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Progress in Geophysical Aspects of the Rotation of the Earth

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Introduction. Rochester (1973) reviewed the subject of the Earth's rotation at the second Geodesy, Solid Earth and Ocean Physics Research Conference, emphasizing the developments made since the publication of Munk and MacDonald's (1960) book, "The Rotation of the Earth". In his introduction, Rochester stressed the many aspects of the geophysical causes and consequences of the Earth's rotation and how the subject has drawn the attention of scientists working in fields ranging from astronomy to palaeontology. This multiplicity renders a completely satisfactory review of the subject unlikely and I limit myself here to some of the more pertinent results obtained since Rochester's review, results which were already foreshadowed in the Conference Report by Kaula et al. (1973). I have attempted a more complete review elsewhere (Lambeck, 1978a).

Any discussion of the Earth's variable rotation is conveniently separated into three parts: (i) the motion of the rotation axis in space, precession and nutation, (ii) the motion of the rotation axis relative to the Earth, polar motion, and (iii) the rate of rotation about this axis, or changes in the length of day. I follow this separation in the following review with emphasis on the last two aspects. I do not discuss the methods nor results of the observation process (see Guinot, 1978; Aardoom, 1978; Lambeck, 1978b).

Polar motion

Low frequency variations

The schematic polar motion power spectrum is illustrated in figure 1. The power near zero frequency (not shown) is a consequence of a secular drift in the pole position; in a westerly direction at a rate of 0$''$.002-0$''$.003 yr^{-1}. This motion appears to be real and probably a consequence of an exchange of mass between the Greenland ice sheet and the oceans. If the observed secular rise in sea level of about 1 mm yr^{-1} is due to the melting of this ice cap, then the rate of melting must be of the order of 10 cm yr^{-1} over the entire sheet and this would result in the observed secular pole shift. The pole shift is quite insensitive to the melting of the Antarctic ice (Lambeck, 1978a). There is a suggestion that, superimposed on this drift, a decade scale wobble occurs, mainly an oscillation at right angles to the direction of drift (Figure 2). Markowitz (1970) finds a period of about 24 years, Vincente and Currie (1976) suggest 30 years. The reality of this motion remains obscure. In particular, a comparison of recent annual mean pole positions obtained by the Bureau International de l'Heure (BIH) and the five station International Latitude Service (ILS), indicate differences that are quite similar to this wobble (figure 3). This strongly suggests that it is a consequence of the observing process and not of a real excitation of the rotation axis direction.

Evidence for polar wander over geologic time remains obscure. Goldreich and Toomre (1969) concluded that if "continental drift" occurred, a large scale wandering of the axis of rotation is inevitable and that, for N plates moving at an average velocity σ, the pole moves at a rate of $\sigma\sqrt{N}$. But the more recent paleomagnetic studies do not require such large scale "absolute" motion of the pole to explain the apparent pole paths for the various parts of the world (McElhinny, 1973). This does not imply that the lower mantle viscosity is, after all, very high but is a consequence of the continents not being isostatically compensated in the Goldreich and Toomre model. The actual change in the inertia tensor due to plate tectonics is a second order effect rather than a first order one (Lambeck, 1978a,c).

Chandler wobble

The interesting part of the polar motion spectrum remains the Chandler wobble, centered at a period of about 14 months. This is the Eulerian precession for the non-rigid Earth. The three questions associated with this motion concern its period, dissipation and excitation. Smith (1977) has re-evaluated the lengthening of the wobble period due to the mantle elasticity and core using the normal mode approach suggested by

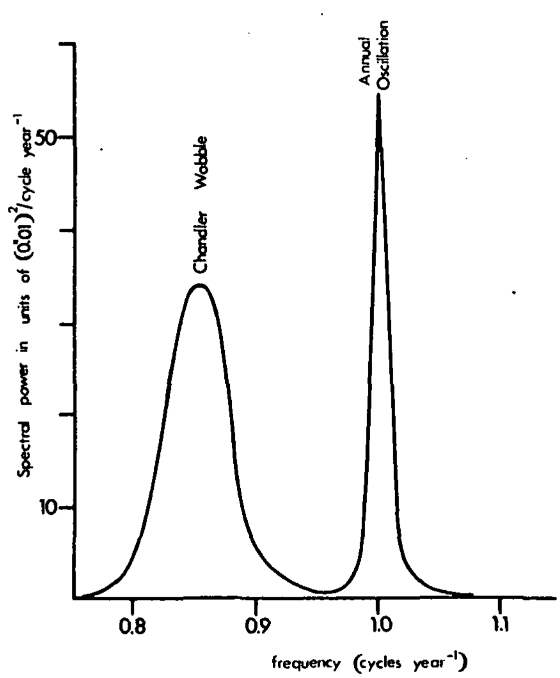

Fig. 1. Schematic Power spectrum of polar motion.

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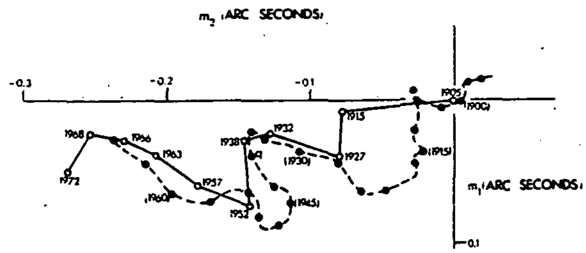

Fig. 2. Motion of the mean pole according to A. Stoyko. The pole positions present running means based on six year intervals. Positions every 3 years are indicated. The motion of the mean pole according to W. Markowitz is indicated by the open circles.

Gilbert (1971). In this method, the rotational deformations are expressed by the normal-mode eigen functions which themselves can be verified from studies of the Earth's free oscillations. Smith's treatment also places fewer constraints on the nature of the permissible core motion than do the earlier Jeffreys-Vicente and Molodensky theories. Sasao et al. (1977, 1978) discusses these theories in more detail, in particular, they introduce viscous coupling between the core and mantle. Smith has considered core models with different degrees of stratification, expressed by the Brunt-Vaisala frequency N_{B-V}, but only in the case of a strong stable stratification is the wobble period affected by a significant amount (Table 1). Seismic data remains inadequate to

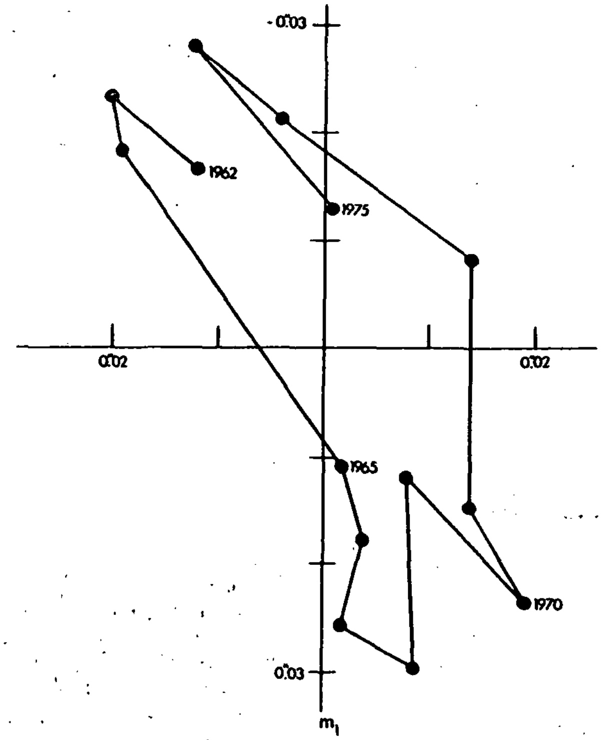

Fig. 3. Annual mean differences between the B.I.H. and ILS pole positions from 1962 to 1975.

TABLE 1. Summary of theoretical contributions to the Chandler wobble period.

	Model 1 ($N^2_{BV}=0$)			Model 2 ($N^2_{BV}=3.38\times10^{-7}$)	
Mantle and Core (Smith, 1977)	403.6			405.2	
Mantle anelasticity					
$Q_m = 300$ } Constant Q model			3.9		
$Q_m = 600$ } Constant Q model			1.8		
frequency dependent Q model			7.6		
Ocean equilibrium theory			27.4		
Non equilibrium ocean tide					
$Q_w = 20$; $\varepsilon = 22°$			25.2		
$Q_w = 40$; $\varepsilon = 11°$			26.8		
$Q_w = 80$; $\varepsilon = 6°$			27.2		
Total period	Q_m=300	Q_m=600		Q_m=300	Q_m=600
$Q_w = 20$	432.7	430.6		434.3	432.2
$Q_w = 40$	434.3	432.2		435.9	433.8
$Q_w = 80$	434.7	432.6		436.3	434.2
$Q_w = \infty$	434.9	432.8		436.5	434.6
Observed period (Jeffreys, 1968)			434.3 ± 2.2		

TABLE 2.

Period	12 hrs	24 hrs	6 m	12 m	430 days
Q = 10	42.3	44.5	60.6	61.6	62.8
50	9.7	10.3	14.8	15.2	15.4
100	5.0	5.2	7.6	7.8	7.9
200	2.6	2.7	3.8	3.9	4.0
400	1.3	1.4	2.0	2.0	2.0
600	0.93	1.0	1.4	1.4	1.4
1000	0.60	0.63	0.80	0.80	0.80
600	0.5	0.7	4.1	5.2	5.5

Percentage changes in k_2 due to dispersion as a function of the tidal frequence. The first 7 lines are based on the assumption that Q is independent of frequency from about 1 sec periods to the period in question. The last line is based on the assumption that Q is proportional to $(\text{frequency})^{1/3}$.

determine the extent to which the core may be stratified.

In most recent calculations of Love-numbers and wobble period, the elastic parameters K,μ have been assumed to be frequency independent and values deduced from body waves or free oscillations have been used. This is at variance with the current interpretation given to the observation that the mantle Q is nearly constant over a wide range of seismic frequencies (Anderson et al. 1977; Kanamori and Anderson, 1977; Liu et al. 1976), and dispersion effects may be of consequence. In particular, the real parts of Love numbers become frequency dependent. Table 2 summarizes some results for a homogeneous incompressible Earth model and Table 1 gives corrections to the wobble period (see also Dahlen, 1978; Lambeck, 1978a). Very recently, Anderson and Minster (1978) suggested that Q may be proportional to $(\text{frequency})^{1/3}$, based on an observation that "high temperature background" may reflect as important dissipation mechanism. Table 2 also summarizes the resulting frequency dependent Love numbers in this case.

Dahlen (1976) has developed the equilibrium pole tide theory to allow for self attraction and the yielding of the Earth under the variable tidal load. But whether or not this tide follows an equilibrium theory remains obscure. Work by Miller and Wunsch (1973) and Currie (1975) confirm that, while the pole tide is an ocean-wide phenomena, available evidence is quite inadequate to map its global characteristics. Especially the evidence for any lag of the tide behind the forcing function remains unsatisfactory (Hosoyama, 1976). Lambeck (1978a) has modelled the consequences of dissipation of the pole tide on the rotation in a manner similar to that used for dissipation in the lunar tides (Lambeck, 1977) and the modification of the wobble period may be of the order of 0.5 days (Table 1). This remains below the noise level of the astronomical estimates of the wobble period (Jeffreys, 1968).

As the wobble period depends on the rate of rotation and, to a lesser degree on the ocean-land distribution, the period will have changed during geological time. In particular it could have approached that of the annual frequency (Cannon, 1974) but the geological consequences of such a resonance do not appear to be very severe (Lambeck, 1975a).

Excitation mechanisms of the wobble have continued to receive attention since Rochester's 1973 review. The then current state of the seismic excitation hypothesis was reviewed in Kaula et al. (1973). At that time, differences of opinion existed on (i) the correct treatment of the core-mantle interface in the theoretical evaluation of the earthquake induced changes in the inertia tensor, (ii) on the appropriate moment-magnitude relationship, (iii) on the most satisfactory manner of evaluating the cumulative effects of seismic activity and, (iv) on the observational evidence for discontinuities in the curvature of the pole path. The theoretical aspects now appear to be resolved (Mansinha et al. 1978) and all recent estimates of the pole shift due to an earthquake of given source parameters are in agreement (Table 3). O'Connell and Dziewonski (1976), using the free oscillation

TABLE 3.

Event		Wobble excitation Magnitude (0".01)	Direction	Author
Alaska				
1964		0.72	201°	M.L. Smith (1977)
"		0.73	202°	Dahlen (1973)
"		1.11	203°	O'Connell and Dziewonski (1976)
Chile				
1960	(1)	2.12	114°	M.L. Smith (1977)
	(2)	2.80	118°	M.L. Smith (1977)
"	(1)	2.56	109°	O'Connell and Dziewonski (1976)
"	(1)	2.2	101°	Mansinha et al. (1977)

Comparison of estimates of the shift of the inertia axis due to the Alaskan (1964) and Chile (1960). For the Chile event (1) refers to the main shock while (2) refers to the precursor.

formulation, evaluated the cumulative seismic excitation function from 1900 to 1970 and conclude that it is adequate to maintain the wobble against damping, although Kanamori (1976, 1977a) argues that their seismic moments are overestimated. It is now clear that no single moment-magnitude relationship describes all earthquakes, and that the magnitudes, determined from 20-100s period seismic waves, may be an inadequate measure of the overall seismic moment. In particular, there may be a considerable low frequency or aseismic contribution to the moment. Evidence for this comes from observations of (i) precursors (Kanamori and Cipar, 1974; Kanamori and Anderson, 1975; Dziewonski and Gilbert, 1975; Thatcher, 1974), (ii) tsunami earthquakes (Kanamori, 1972), (iii) studies of aftershock areas (Kanamori, 1977a; Stuart and Johnston, 1974) and, (iv) discrepancies between seismic slip and plate motions (Chen and Molnar, 1977; Kanamori, 1977b). If the aseismic slip occurs over time intervals that are short compared to the Chandler wobble period, they may contribute significantly to the excitation of this wobble. This suggests that, since it is not well understood which particular earthquakes are associated with significant aseismic slip, a more useful measure for comparing the wobble with seismic activity than seismic moments of the largest events, is the frequency (N) of earthquakes above a certain magnitude. This is shown in figure 4 together with the elastic energy E_e released by large earthquakes. Both N and E_e show trends that are comparable to the fluctuations in the wobble amplitude.

The other excitation mechanism that has recently been revised is the variability in the atmospheric mass distribution (Wilson and Haubrich, 1976). But the meteorological data is neither sufficiently reliable nor complete to permit an unambiguous interpretation. About the Chandler frequency, Wilson and Haubrich estimates an average power in the meteorological excitation spectrum that is about one sixth of that required to maintain the wobble. Part of the missing power may be a consequence of an absence of surface pressure data over central Asia, of an overestimation of the astronomic spectrum due to noise in the data and to hydrological fractors, but the evidence remains tenuous. In particular, differences found between the results by Wilson and Haubrich, Siderenkov (1973) and Jochmann (1976) for the annual atmospheric excitation function points to an unsatisfactory situation (Lambeck, 1978a).

Possibly a combination of seismic and atmospheric processes contribute to the excitation although the combined excitation (Figure 5) is not much better than either one alone (see also Wilson and Haubrich, 1977). Inversion of the wobble data, after removal of the atmospheric contribution, for the seismic moments has not led to conclusive results, and it is not yet possible to infer parameters defining large seismic or aseismic events from past Chandler wobble data.

The third Chandler wobble question concerns its energy sink. Oceans are a probable sink but observational evidence for the departure of the pole tide from equilibrium remains totally inadequate. Wunsch (1974) has modelled dissipation of the pole tide in the North Sea and concludes that, by extrapolating to the world's shallow seas, the oceans may just provide an adequate sink. In the pole tide problem one requires the lag of the 2,1 harmonic in the tide expansion to describe the total rate of dissipation and a value of 5° is adequate to explain the observed wobble Q. But observed values for the lag are very variable and unreliable.

Significant dissipation in the mantle is generally ruled out because estimates for the wobble Q are less than the corresponding seismic Q's. Stacey (1977) has stressed, however, that a direct comparison of seismic and wobble Q's is not valid since the two are defined differently. The former is defined as being proportional to $\Delta E/E_e$ where ΔE is the amount of energy dissipated in one cycle of the motion and E_e is the peak elastic energy stored in the cycle (see, for example, O'Connell and Budiansky, 1978). The

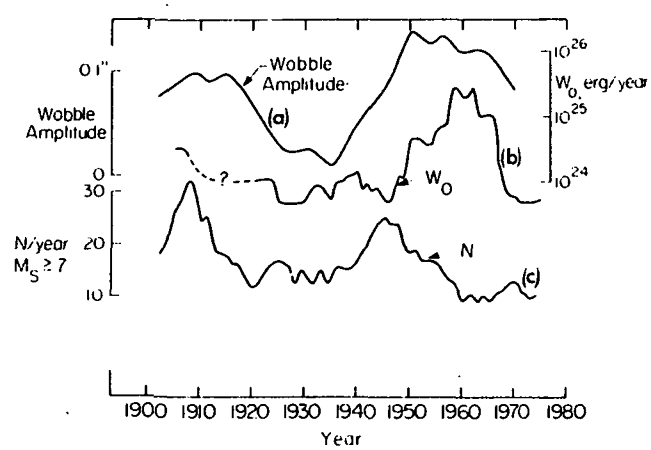

Fig. 4. Comparison of the amplitude of the Chandler wobble (curve a), the elastic energy realeased by earthquakes (5-year running means, curve b) and the annual number of earthquakes of $M_s \geq 7.0$ (5-year running means, curve c). From Kanamori (1977a).

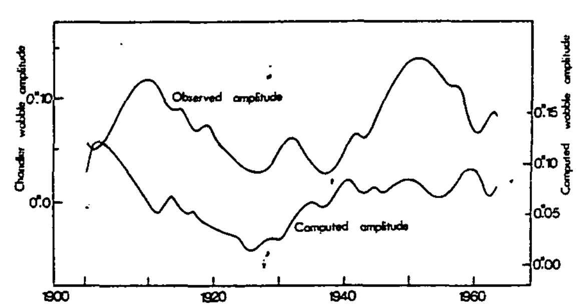

Fig. 5. Fluctuations in the Chandler wobble amplitude as observed (upper curve) and as computed from the seismic and atmospheric data. Note that the vertical scales of the two curves have been displaced.

wobble Q, however, is usually defined as being proportional to $\Delta E/E_k$, where E_k is the kinetic energy of the rotational motion. But E_e is about 15% of E_k (Merriam and Lambeck. 1978) and a wobble Q of, say 100, implies a seismic Q of about 15. The lower Q values at the Chandler wobble frequency than at seismic frequencies may not be wholly unexpected if Anderson and Minster's Q model is valid.

Annual frequency

The comparison of meteorological and astronomical estimates of the seasonal polar motion remains unsatisfactory for several reasons, including (i) incomplete meteorological, oceanic and hydrological data, (ii) global year to year variability in these data, (iii) poor wobble data at the annual frequency prior to the introduction of the B.I.H. polar motion results. Siderenkov (1973), Jochmann (1976), Wilson and Haubrich (1976) have discussed the atmospheric contributions but significant differences between the results exist. The contribution from ground water storage is important but only very approximately known. Seasonal variations in the ocean volume also are not insignificant. The unsatisfactory situation is discussed in more detail by Lambeck (1978a).

Length of day

Astronomers observe the integrated amount by which the Earth is slow or fast relative to a uniform time scale. This gives the difference between Universal Time and Atomic Time, or UT-AT. Of greater geophysical interest is the proportional change in the length of day (l.o.d), or

$$m_3 = \frac{\omega_3 - \Omega}{\Omega} = - \frac{\Delta(\text{l.o.d})}{\text{l.o.d}} = - \frac{d}{dt}(\text{UT-AT}),$$

where ω_3 is the instantaneous rotation velocity and Ω is a nominal or mean velocity. Figure 6 illustrates the schematic spectrum of m_3. It consists mainly of a continuum upon which a number of periodic phenomena are superimposed.

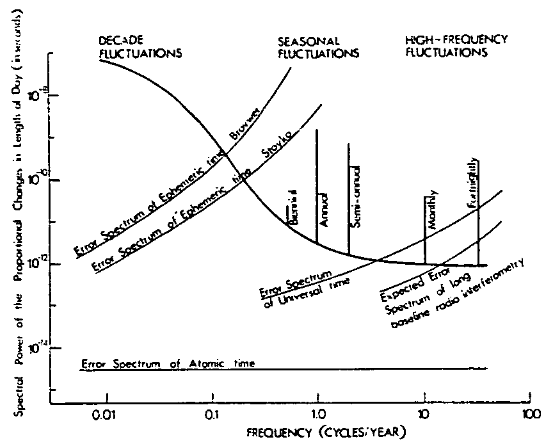

Fig. 6. Schematic power spectrum of the proportional changes in length of day.

Secular acceleration

The secular acceleration of the Earth's rotation, $\dot{\Omega}$, is mainly a consequence of the dissipation of energy of the Moon and Sun raised ocean tides (see Lambeck, 1977, for a review of the problem). A further contribution may come from slow changes in the mass distribution within the Earth or from non-tidal torques acting on the mantle. Denote these two contributions to $\dot{\Omega}$ by $\dot{\Omega}_T$ for the tidal part and $\dot{\Omega}_{NT}$ for the non-tidal part. Associated with $\dot{\Omega}_T$ is an acceleration $\dot{n}$ of the Moon in longitude and, as astronomical observations give $\dot{n}$ and $\dot{\Omega}$, the tidal contribution to the spin can be separated from the non-tidal part. This separation has been evaluated in detail by Lambeck (1975b, 1977). Table 4 summarizes recent results. The astronomical estimates of $\dot{\Omega}$ and $\dot{n}$ come mainly from the analysis of discrepancies between computed and recorded eclipse paths. Most of these records, made during the last three millenia, have been revised by Newton (1970, 1972), Stephenson (1972), Muller and

TABLE 4.

	Astronomical Estimates	Tidal Estimates	Satellite Estimates
$\dot{n}_{☾}$ $(10^{-23} s^{-2})$	-1.35 ± 0.10	-1.49 ± 0.5	-1.33 ± 0.25
$\dot{\Omega}_T\vert_{\dot{n}_{☾}}$ $(10^{-22} s^{-2})$	-5.48	-6.05	-5.40
$\delta\,\dot{\Omega}_T$ $(10^{-22} s^{-2})$	-1.47	-1.47	-1.47
$\dot{\Omega}_T$ $(10^{-22} s^{-2})$	-6.95	-7.52	-6.87
$\frac{dE}{dt}$ $(10^{19}\text{ergs } s^{-1})$	-3.94 ± 0.30	-4.26 ± 0.45	-3.90 ± 0.70

Recent astronomical, tidal and satellite estimates of the tidal and non-tidal accelerations of the Earth and Moon. $\delta\dot{\Omega}_T$ is the contribution to $\dot{\Omega}_T$ from tidal terms other than those contributing to $\dot{n}$.

Stephenson (1975), and Muller (1975, 1976), with the consequence that we have not only a more extensive data set but also more confidence in the reliability of the results. Muller (1975, 1976) estimates that a principal uncertainty now is due to the constants used in the lunar theory. Telescope observations give $\dot{n}$ but not $\dot{\Omega}$ since they cover too short a time interval (about 170 years) to permit a separation of $\dot{\Omega}$ from long-period fluctuations (Morrison, 1978). Perturbations in the motions of artificial Earth satellites, due to the ocean and body tide potentials, also permit $\dot{n}$ to be estimated (Lambeck, 1975; Cazenave et al. 1977). Recent estimates of tidal parameters are by Cazenave and Daillet (1977), Goad and Douglas (1978) and Felsentreger et al. (1978). The satellite deduced results for $\dot{n}$ are comparable with those obtained from the astronomical observations. Lunar laser ranging also gives a comparable results for $\dot{n}$ (Williams et al., 1978).

At present, $\dot{\Omega}_T \approx 53.1\dot{n}$ (Lambeck, 1977) and this value has been used to determine the tidal acceleration $\dot{\Omega}_T$. The difference between this value and the observed acceleration is small and uncertain and Muller (1976) attributes most of it to a change in the gravitational constant. He finds, for different cosmologies, $\dot{G}$ of the order 2-5 parts in 10^{11} year^{-1} from an analysis of historical records of astronomical events, telescope observations and tidal theory.

There has never been a shortage of geophysical explanations for the non-tidal acceleration but this quantity now appears to be sufficiently small to conclude that there have not been significant changes in the Earth's moment of inertia during the past 3000 years. Improvements in the value of $\dot{\Omega}_{NT}$ will come only if $\dot{\Omega}$ can be determined with greater accuracy and this will only come about if the record of eclipse observations can be extended. The method used by Sawyer and Stephenson (1970), to locate the valuable Ugarit record of 1375 BC, leaves hope that new observations may still be found in, for example, the Sumarian records. The Chinese records of the court astronomers have been fully explored for astronomical references by Stephenson (1972) but, according to C.P. Fitzgerald, provincial records may provide valuable additional accounts of eclipses back to about 200 AD. Furthermore, ancient Chinese astronomical records of occulatations and conjunctions are many (e.g. Ho, 1966) and these have not yet been fully investigated for purposes of determining $\dot{n}$ and $\dot{\Omega}$.

The paleontological evidence for the Earth's past rotation has been reviewed by Lambeck (1978a, d) and Scrutton (1978). Since 1973 there have been few new observations of the frequencies of growth rhythms and the interval has been characterized by caution not always evident in the earlier studies (Clarke, 1974; and papers in the volume edited by Rosenberg and Runcorn, 1975). Lambeck (1978d) has estimated $\dot{\Omega}$ and $\dot{n}$ from the better documented coral and bivalve growth rhythms since the Ordovician and finds that $\dot{\Omega}$ has, on the average, been about 75% of the present value and that there is no evidence in the data for a significant non-tidal acceleration during the last 4×10^8 years (Table 5). The results

TABLE 5.

	$\dot{\Omega}$ $(10^{-22}s^{-2})$	$\dot{n}_{☾}$ $(10^{-23}s^{-2})$
Coral Data	-6.3±0.7	-1.5±0.4
Bivalve Data	-5.9±0.6	-1.3±0.3
Combined Data	-5.2±0.2	-1.2±0.2
With constraint that $\dot{\Omega}_{NT}=0$	-5.2±0.2	-1.0±0.1

Summary of estimates of the Paleorotation of Moon and Earth.

and some geophysical consequences are discussed further in Lambeck (1978c).

Long period fluctuations

Evidence for the long period (from about 10 to 300 years) comes from the telescope observations made since the eighteenth century and, according to F.R. Stephenson, there is some hope that timed eclipse observations may aid in extending the record further back into time. Morrison (1972) has transformed the data since 1820 into a uniform system and is now carrying out a complete revision of the data. Figure 7 illustrates the results for m_3 and its derivative $\dot{m}_3$ (see Lambeck, 1978a). The causes of these changes remain obscure. Electromagnetic coupling of core motions to the mantle, possibly reinforced by topographic coupling, is the main contender. In particular, they can explain the changes occurring over periods of several decades and longer (Yukutake, 1973; Watanabe and Yukutake, 1975). The main unknowns in the theory are the strength of the torroidal field at the core-mantle boundary and the value

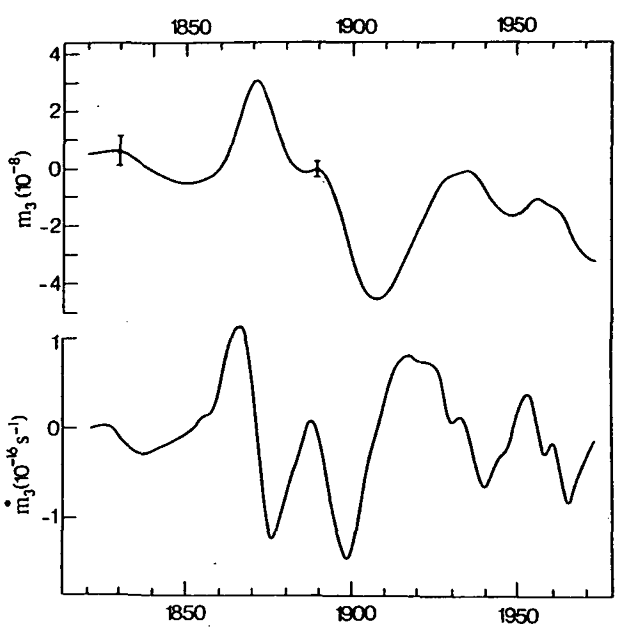

Fig. 7. m_3 and dm_3/dt based on telescope observations from 1820 to 1970.

for the lower-mantle electrical conductivity σ_m. Values of $2\text{-}3\times10^2$ ohm m^{-1} are required to explain the decade fluctuations but several recent investigators has suggested that it may be much higher than this (Kolomiyseva, 1972; Braginski and Nikolaichik, 1973; Alldredge, 1977; Stacey et al. 1978). The observational evidence, however, remains inadequate (see, for example, Courtillot and LeMouel, 1977). An argument by Hide (private communication, July 1978) also opposes these high conductivity values, on the grounds that his proposed method of estimating the core radius from magnetic data (Hide, 1978) would otherwise not work, and his good agreement with the seismic estimate for the core radius would have to be considered as fortuitous.

Topographic coupling was proposed by Hide (1969) when electromagnetic coupling appeared inadequate. Hide (1977) has reviewed the state of the mechanism. To estimate the total torque on the mantle exerted by the flow past the irregular core-mantle boundary, requires the form of the topography and the flow-field. Even their power spectra are unknown although, from considerations of the Earth's gravity field, it does not appear that the height of the irregularities can exceed a few hundred meters (Lambeck, 1976). The mechanism, however, does not necessarily require large values for the topography since the drag coefficient is dependent on the topography itself and high topography does not necessarily imply strong topographic coupling. Moffatt (1978) has investigated the dependence of the drag coefficient on the strength of the magnetic field within the core and concludes that this coefficient may be significantly greater than unity, thereby enhancing the effectiveness of the topographic coupling mechanism without requiring large amplitude topography.

Variations on the decade time scale in the atmospheric and oceanic mass redistributions have been evaluated by Lambeck and Cazenave (1976) but, while they find variations that exhibit similar trends to those observed in length of day, the magnitudes are inadequate. In particular, periods of an accelerating Earth, as occurred from 1840 to 1870 and again from 1900 to 1940 are associated with increasing strength in the zonal wind circulation and with increasing surface temperatures, while periods of deceleration occur when there is a general decrease in the strength of the zonal wind circulation and surface temperatures decrease (Figure 8). These correlations require further study but they do raise the possibility that length of day changes and the global climatic changes, over intervals of a few decades, may have a common origin.

Seasonal and higher frequency variations

The Earth's rotation appears to exhibit considerable fluctuation in speed due to its apparent sensitivity to changes in the zonal wind circulation. Thus the astronomical observations may provide useful measures of the strengths and frequencies of the global wind circulation. The atmospheric induced changes in length of day appear to be important for periods of several years down to a few days and, as such, they need to be known if other excitation mechanisms are investigated. Examples of such excitations include the study of changes in length of day caused by large earthquakes and aseismic motion or by core-mantle coupling. Lambeck and Cazenave (1977) have reviewed the subject. The seasonal changes in the length of day are clearly of atmospheric origin, due to a periodic exchange of angular momentum between the zonal wind circulation and the Earth's mantle (Lambeck and Cazenave, 1973). Changes in the inertia tensor associated with atmospheric, oceanic and hydrological mass redistributions, play only a minor role at the semi-annual frequency are tidal contributions significant. The astronomical data since 1955 have indicated that there have been some significant changes in the year-to-year amplitudes of the seasonal terms and this is indicative of a substantial change in the year-to-year circulation of the atmosphere. The significance of the year to year fluctuation in the semi-annual oscillation amplitude in length of day is less clear, since Okazaki (1975) has suggested that part of this may be a consequence of the astronomical reduction process, at least for the data prior to 1962. The length of day observations also suggest an intermittent quasi-biennial oscillation, variable in both period and in amplitude. This is a further consequence of an exchange of angular momentum between the mantle and the atmosphere. Meteorological evidence for longer period global wind cycles is minimal, particularly for winds at higher altitudes. Siderenkov (1969) has used the surface torque approach to estimate the angular momentum changes in the atmosphere and finds some evidence that length of day changes over roughly 5 year periods are a consequence, at least in part, of zonal winds. Okazaki (1977) came to a similar conclusion as did Lambeck and Cazenave (1974). Recently Rosen et al. (1976) computed the yearly

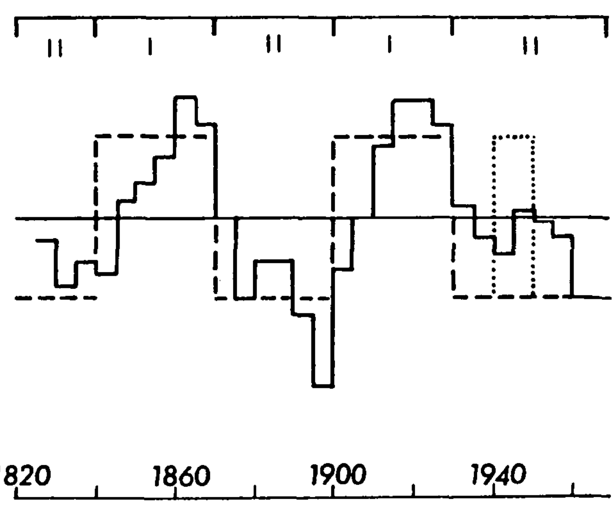

Fig. 8. Schematic representation of the circulation pattern (broken line). Periods of increasing strength of the zonal circulation are denoted by I, those of decreasing circulation are denoted by II. The astronomically observed $\dot{m}_3$, running mean values over 15 years, are defined by the solid line.

value of the zonal angular momentum in the northern hemispheric circulation for a ten year period. The fluctuations are comparable to those deduced from the l.o.d. but smaller (Figure 9). This is further evidence for an important atmospheric excitation functions for l.o.d changes over periods up to about 10 years and stresses the need once more for detailed atmospheric angular momentum calculations if one wants to study core-mantle coupling mechanisms.

At frequencies above 2 cycles per year, the length of day data shows considerable perturbation and much of this can be attributed to the zonal winds (Lambeck and Cazenave, 1974; Okazaki, 1977). Studies of the global atmospheric circulation spectrum show a broad continuum from about 8 months to a few days (Mitchell, 1976) with peaks rising above the average near the semi-annual and fortnightly frequencies. The circulation patterns become more regional at the higher frequencies and it is not clear at what frequency they cease to have inconsequential effects on the rotation. Abrupt aperiodic changes in circulation patterns are often observed and may cause some of the irregular week-to-week changes in rotation occasionally reported (e.g. Guinot, 1968). Here, however, both the astronomical and meteorological data become less reliable and a more detailed and precise compilation of the zonal winds is required for all latitudes and altitudes.

Precession and Nutation

Revisions of E.W. Woolards rigid body nutation theory have been carried out by Kinoshito (1976) and Murray (1978) but the interest of the subject lies mainly in the departures from this theory due to the Earth's non-rigid response to the solar and lunar torques. The symposium proceedings edited by Federov et al. (1978) discuss various aspects of these motions.

The most complete solution for the forced nutations of realistic Earth models is by Sasao et al. (1977, 1978) although the theory by Smith (1977) could be readily extended to include these forced terms. Sasao et al's theoretical

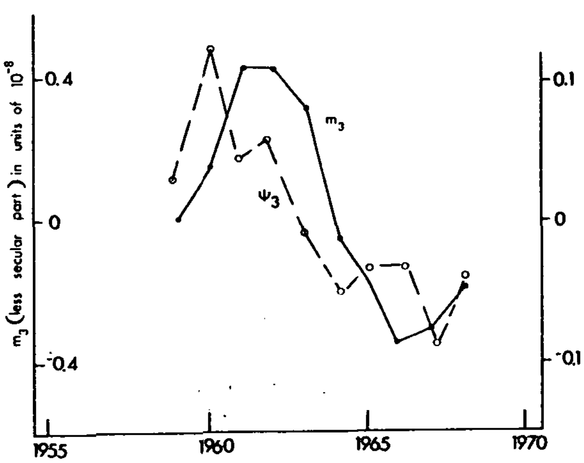

Fig. 9. Estimates of the zonal angular momentum in the northern hemispheric circulation (from Rosen et al. 1976) and changes in length of day.

TABLE 6.

	Period in mean solar days				
	-365.3	-6798	6798	182.6	13.7
Observed* amplitude	1.336	0.9962	1.0030	1.034	1.031
Theoretical* amplitude	1.283	0.9965	1.0030	1.034	1.031

* based on the mean of values given by Sasao et al.

Summary of the theoretical nutation amplitudes (Sasao et al. 1977) and a comparison with the observed periods. The amplitudes are normalized by the rigid body value.

nutuation periods are generally in satisfactory agreement with the observed values (Table 6) although the latter are subject to some uncertainty.

A second nutuation has received considerable attention in the recent literature. This is the nearly diurnal wobble, or what Toomre (1974) calls the principal core nutation. This motion, already discussed in the last century, received only intermittent attention until N.A. Popov claimed to have detected it in the astronomical observations. The motion, a free wobble due to the Earth's liquid core, is a retrograde motion of the rotation axis about the axis of figure and is accompanied by a very much larger motion of the rotation axis in space with a period of roughly 400 days (Toomre, 1974). Yatskiv (1972) and Rochester et al. (1974) summarized the subsequent papers dealing with the search for this wobble in the latitude observatios while Rochester et al. and Capitaine (1975) have searched for it in the declination observations without success. The absence of evidence of this motion in space then indicates that the corresponding wobble amplitude becomes totally insignificant. Yatskiv et al. (1975) object to this conclusion but their argument is obscure.

Calculations of the period of the nearly diurnal wobble have most recently been made for realistic Earth models by Smith (1977) and Sasao et al. (1977, 1978) and the results are not very model dependent. This appears to make the whole problem rather academic.

Some further problems

Secular decrease in obliquity

Astronomical observations have revealed a "secular" change in the obliquity, most of which is really a long period perturbation in the Earth's motion due to the gravitational attractions of the planets. A small discrepancy between the observed and the gravitational results has been explained by frictional coupling between the core and mantle during the precession period (Aoki, 1969; Kakuta and Aoki, 1972). A revision of the astronomical data by Duncombe and van Flandern (1976) now points to an insignificant discrepancy and the theoretical study of Rochester (1976) also indicates an insignificant role of core mantle coupling during the orbital evolution of the Earth-Moon system. This conclusion can also be deduced from the studies by Sasao et al. (1977, 1978).

Evolution of the Earth-Moon system

Kaula and Harris (1975) have given a recent review of the dynamical evolution of the Earth-Moon system and of the possible constraints that this may place on the origin of the Moon. These constraints do not appear to be important. The main problem remains the inclination problem: How to get the Moon into an equatorial orbit when it was close to the Earth. The rate of change of the orbital inclination is the sum of perturbations due mainly to three tidal terms (M_2, K_1, O_1), with two tending to decrease (M_2 and K_1) and one (O_1) tending to increase the inclination with time (Lambeck, 1975) and situations can be contrived in which the overall sign if dI/dt changes simply by introducing an appropriate frequency dependent Q law. Rubincam (1975) discusses one possibility, namely a Maxwell Earth in which Q is proportional to frequency. Anderson and Minster's (1978) suggestion of Q proportional to $(\text{frequency})^{1/3}$ may do the same thing, if the tidal frequencies at a given moment are such as to accentuate the O_1 relative to M_2 and K_1 tides.

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