Long Term Variations in the Length of Day and Climatic Change

Kurt Lambeck
Institut de Physique de Globe, Université Paris VI and Département des Sciences de la Terre, Université Paris VII, 4, place Jussieu, Paris, France

Anny Cazenave
Groupe de recherches de Géodésie Spatiale, Centre National d'Etudes Spatiales, Toulouse, France

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Summary

The long-period (greater than about 10 yr) variations in the length-of-day (LOD) observed since 1820 show a marked similarity with variations observed in various climatic indices; periods of acceleration of the Earth corresponding to years of increasing intensity of the zonal circulation and to global-surface warming; periods of deceleration corresponding to years of decreasing zonal-circulation intensity and to a global decrease in surface temperatures. The long-period atmospheric excitation functions for near-surface geostrophic winds, for changes in the atmospheric mass distribution and for eustatic variations in sea level have been evaluated and correlate well with the observed changes in the LOD; although their total effect represents only about 10 per cent of the required excitation. The computed excitation lags the LOD variations by about 10–15 yr. Upper atmospheric winds or other meteorological related factors appear to be quite inadequate to provide the additional driving force that is required if the LOD changes are of meteorological origin. Instead, it appears that the LOD fluctuations and the climatic variations on a time scale of 20–30 yr may have a common origin as has been suggested by Anderson. That is, the results suggest that indirect solid Earth effects on climate may be important.

Introduction

The long-period variations in the (LOD) of duration of the order of a decade and longer, are usually attributed to core–mantle coupling—Rochester (1970) reviews the various mechanisms. There is no absence of proposed coupling mechanisms but all are marred by a lack of precise information on relevant physical properties of the core or lower mantle. The mechanisms reviewed by Rochester are inertial coupling—including topographic coupling; viscous coupling—including laminar and turbulent layer friction, and electromagnetic coupling. Inertial coupling will have no effect on changes in the LOD unless the axis of symmetry of the core–mantle boundary is badly misaligned with the axis of the mantle. Gans' (1972) recent estimate of the core
viscosity, a calculation based on the assumption that the inner–outer core boundary is a melting transition, gives a value for the dynamic viscosity of less than $2 \times 10^{-1}$ (see Rochester's review for a summary of earlier viscosity estimates) and if this estimate is of the correct magnitude it appears to exclude any form of viscous coupling. Topographic coupling as proposed by Hide (1969) also appears to be ruled out if Lambeck’s (1976) estimate of an upper limit of $200 \text{m}$ or less for the root mean square value of the possible core–mantle topography is valid. Rochester (1970) appears to favour electromagnetic coupling as the possible mechanism and most studies (e.g. Rochester 1960; Roden 1963; Roberts 1972; Yukutake 1972) show that this mechanism is almost capable of transferring the angular momentum between the core and mantle necessary to explain the decade variations in the LOD. In general the longer the duration of the fluctuations the more readily can they be explained by electromagnetic coupling of the core to the mantle. In particular, the non-tidal secular acceleration of the Earth has been attributed to this mechanism (Yukutake 1972) although the acceleration has also been attributed to the observed Earth's delayed response to the unloading of the Pleistocene polar ice and subsequent rise in sea level (Dicke 1969; O’Connell 1971).

Support for electromagnetic coupling has been drawn from the apparently good correlation observed between the westward drift of the dipole field and variations in the LOD as first noted by Vestine in 1952 and as predicted by E. C. Bullard’s theory for the westward drift.

Studies by Vestine & Kahle (1968) and Kahle, Ball & Cain (1969) appear to confirm this correlation but two recent studies—Jin (1974) and Malin & Clark (1974) negate it. According to Malin & Clark more comprehensive collections of models by Barraclough show that before 1940 the scatter in them is too large to permit reliable determination of the variations in the westward drift. Dicke (1966) also discusses the magnetic evidence for the period 1830–1950 and notes that the drift of the low degree (1, 2 and 3) harmonics of the magnetic field have undergone severe disturbances near 1900 at a time when the LOD also changed rapidly. A similar change in the LOD occurring around 1870 is, however, not clearly evident in his magnetic data.

Decade variations in the drift of the geomagnetic field are often attributed to an internal origin whereas variations with periods of less than about 4 or 5 yr are assumed to be of external origin (see, e.g. Currie 1966). This separation appears to be somewhat arbitrary and recent analyses by Alldredge (1975) and by Courtillot & Le Mouël (1975) indicate that fluctuations in the magnetic field with periods up to at least 15 or 20 yr are of external origin. If this is confirmed it removes one further support for the concept of an electromagnetic coupling of the core motion to the mantle at the decade part of the spectrum and stresses the need for a re-investigation of other possible explanations of the long-period variations in the LOD. The observations apparently do not permit a firm conclusion to be drawn about the importance of core–mantle coupling as a means of exciting the LOD variations at lower frequencies such as the changes observed near 1840, 1870 and 1905.

In this paper we investigated the possible meteorological and related oceanic contribution to the Earth’s rotational-excitation function. We do this for three reasons. First, the calculation by Lambeck & Cazenave (1974) of the excitation function from 5 yr of wind data shows that there is apparently significant power in the low frequency part of this spectrum. Secondly, a casual examination of numerous climatic indices for the last two centuries show very similar trends to the variations observed in the LOD. Third, if the LOD changes cannot be a consequence of the meteorological phenomena yet the two correlate, a mechanism must be sought that explains both and this may eventually lead to a better understanding of these two phenomena.

Munk & MacDonald (1960) dismiss meteorological factors as important contributors to the excitation function at low frequencies on the ground that the amplitude
of the decade variations in the LOD are considerably larger than the seasonal variations which are almost entirely of meteorological origin. Dicke (1966) likewise dismisses any meteorological contribution to the low-frequency part of the excitation spectrum with the argument that the winds would have to be in excess of 500 km hr$^{-1}$ at the equator. But as the effect of winds on the LOD is the difference between easterly and westerly circulations this conclusion is not immediately evident. Simple model calculations show that latitudinal shifts of about 10° in the zones of maximum and minimum winds suffice to explain the observed changes in the LOD while ground-pressure observations indicate that shifts of up to 5° have occurred in the near-surface winds during the last 100 yr (see, e.g. the pressure indices of Lamb & Johnson 1966). This calculation, plus the observation that a number of climatic indices follow the LOD variations, leads us to reconsider the meteorological role in the low-frequency range of the LOD spectrum.

Meteorological phenomena can affect the Earth's rotation in three ways. In the first instance there is an exchange of angular momentum between the solid Earth and atmosphere—it is the mechanism responsible for most of the seasonal and higher-frequency variations observed in the LOD (Lambeck & Cazenave 1973, 1974). Secondly the constant redistribution of mass results in a time-dependent inertia tensor. Such changes play a dominant role in the annual variations in polar motion (Munk & Hassan 1961; Wilson 1975) but they do not contribute in any important way to the seasonal part of the LOD spectrum (Munk & Hassan 1961; Lambeck & Cazenave 1973). They can perhaps make an important contribution if the long-period variations in mass distribution occur near the surface and the little information available suggests that this may be the case. Further changes in rotation can come about indirectly from meteorological causes, in particular by changes in mean sea level through variable ice storage in the polar caps. Likewise, meteorological changes can also cause changes in the ground-water storage and change the mass distribution of the Earth, or it can modify the ocean currents thereby modifying the angular-momentum balance of the Earth. Of these indirect factors, only the sea effect is amenable to facile treatment although variations in sea level are difficult to interpret because of a multitude of factors that play on the sea surface—not all of which result in a modification of the inertia tensor; only the eustatic changes are of interest. But again a trend in the sea-level curve appears (albeit less clear than the trend in the atmospheric data) that follows the long-period variations in the LOD. Eustatic sea-level variations can also be estimated from the ice and snow budgets of polar-land ice and mountain glaciers, but the method is no more discriminatory now than it was when reviewed by Munk & MacDonald in 1960.

In the following sections we establish that the meteorological- and oceanic-excitation functions are inadequate by an order of magnitude to explain the observed changes in LOD at the decade part of the spectrum despite the fact that the total excitation correlates with the LOD changes. The analysis leads to two alternative hypotheses: either there is a further excitation that has been ignored in the analysis, or climatic change and LOD are related by a third, as yet unknown, mechanism. One possible mechanism involving volcanic activity and seismically-induced changes in the Earth's inertia tensor has been proposed by Anderson (1974, 1975). A third possibility, that the climatic changes are a consequence of the LOD changes must be discarded as being energetically inadequate.

The astronomical data and the excitation function

Observations of the LOD go back to the seventeenth century and the most reliable data comes from observations of occultations of stars by the Moon from about 1820 onwards (Brouwer 1952). The most recent compilation and analysis of the available data is by Morrison (1973) and is based in part on the work of C. F. Martin. The
observed quantity is the integrated amount $\tau$ by which the Earth is slow or fast compared with a uniform reference time. For observations before 1955 this reference time is ephemeris or Newtonian time, a scale based on the observed motions of the Sun, Moon and planets. The integration time is 1 yr. Variations in the LOD are related to the $\tau$ by

$$\frac{\Delta \text{ (LOD)}}{\text{LOD}} = -\frac{d\tau}{dt} = m$$

where, by convention, $\tau$ is positive if the Earth is slow. The $m$ represent variations in the speed of rotation with respect to the uniform time scale. From the annual $\tau$ values the $m$ have been computed using a 3-point Lagrangian interpolation and computing the time derivative at the required epoch. Fig. 1 gives, 5 yr running, mean values of $m$. It is these variations that we wish to interpret in the following sections. Part of the secular deceleration in $m$ is due to tidal dissipation and the astronomical observations lead to a deceleration of $\dot{m} = -(1.4 \pm 0.1) \times 10^{-17} \text{ s}^{-1}$ (Muller & Stephenson 1975) after correction for the solar-tidal contribution (Lambeck 1975). Ocean tidal calculations give $-(1.2 \pm 0.1) \times 10^{-17} \text{ s}^{-1}$ (Lambeck 1975). The remaining secular change in $m$ is of non-tidal origin. Markowitz (1970), following Brouwer (1952) argues that the variable rotation can be explained by sudden changes in the accelerations $\ddot{m}$. The $m$ curve can then be represented by the segments of straight lines while the acceleration $\ddot{m}$ can be represented by a series of step changes (Fig. 1). Changes in $m$ of $5 \times 10^{-8}$ have been observed for the years 1905–1935 and of almost $10^{-7}$ from 1870 to 1900.
The excitation function $\Psi$ describing all phenomena acting on the Earth that cause the variable rotation is related to the rotation by $\dot{m} = \Psi$. This function takes into account changes in the Earth’s relative angular momenta $\Psi(H)$, changes in the Earth’s inertia tensor $\Psi(\Delta I)$ and changes in the torques acting on the Earth $\Psi(L)$. The complete function is given as (Munk & MacDonald 1960, p. 38)

$$\Psi = \Psi(\Delta I) + \Psi(H) + \Psi(L)$$

with

$$\Psi(\Delta I) = -\Delta I_{33}(t)/I_{33}$$

$$= - \frac{R^4}{I_{33}} \int \Delta q(\Phi, \lambda, h, t) \cos^2 \Phi \, ds. \quad (1)$$

$$\Psi(H) = -H_3(t)/\Omega I_{33}$$

$$= \frac{R^4}{I_{33} \Omega} \int \int \rho(h) v_3(\Phi, \lambda, h, t) \cos \Phi \, dh \, ds \quad (2)$$

$$\Psi(L) = \frac{1}{\Omega I_{33}} \int^t_0 L(t) \, dt. \quad (3)$$

$\Delta I_{33}(t)$ is the time-dependent variation in the mean moment of inertia $I_{33}$ referred to a body fixed axis $x_3$ that lies close to the mean position of the instantaneous-rotation axis. $H_3(t)$ is the angular momentum about the $x_3$ axis due to motion relative to the body fixed axes $x_i$. $L_3(t)$ is the torque component along the $x_3$ axis. $\Omega$ is the mean rotation rate of the Earth. We are only concerned with mass distributions near or on the Earth’s surface and if $\Delta \rho$ is the density change, $\Delta q = \Delta \rho h$ is the surface-load distribution on the surface $S$. $R$ is the Earth’s radius, $v_3$ the zonal component of velocity; $\Phi$ denotes latitude. If the change in $m$ of $5 \times 10^{-8}$ observed during the interval 1905–1935 is attributed to changes in angular momentum then $|\Delta H_3(t)| \sim 2.9 \times 10^{33}$ g cm$^2$ s$^{-1}$. Attributing this change in $m$ to a modification of the inertia tensor requires that $|\Delta I_{33}| \sim 4 \times 10^{37}$ g cm$^2$.

**Circulation trends since 1800**

Observations of climatic fluctuations during the last two centuries show two principal types of atmospheric circulation alternating typically every 20–40 yr. The first type (type I) is characterized by an increasing intensity of the zonal circulation at all latitudes and with a poleward migration of the belts of maximum wind intensities. The circulation is accompanied by a decrease in the overall range of surface-air temperatures between the equator and the poles, and by an overall increase in the mean global surface-air temperatures. Ocean-surface temperatures also tend to increase at high latitudes. The type II circulation is characterized by a weakening of the zonal circulation, by a migration of the main streams to lower latitudes and by an overall decrease in temperature. For both types of circulation the migration in latitude and the changing intensities are global phenomena, occurring at all longitudes and in the northern and southern hemispheres although the trends in different regions are not always in phase. Both easterly and westerly winds increase with the type I circulation and both decrease during the type II circulation. Lamb (1972) discusses in detail the available observational evidence for these two main types of circulation during the last 150 yr.

Various meteorological indices exist that reflect these two types of circulation patterns (Fig. 2). Difference in pressure between two latitudes give a measure of the
FIG. 2. Various climatic indices observed since 1820 and the change in the length of day (curve J). (A) Pressure differences between 30°-10° north latitudes in the Atlantic ocean (after Lamb & Johnson 1966). (B) January position indices of the intertropical trough in Australia (after Lamb & Johnson 1966). (C) Zonal circulation indices in the northern hemisphere after Girs (curve C1) and after Dzerdzievski (curve C2). (D) Frequency of south westerly surface winds in England (after Lamb 1972). (E) Trends in mean annual surface temperatures for three latitude ranges (after Mitchell). (F) Air temperatures observed in the Greenland ice sheet (after Johnsen et al. 1970). (G) Mean January temperatures over central England (after Manley 1954). (H) Snow accumulation in Antarctica (after Fletcher). (I) Rainfall index of Santiago, Chile (after Taulis). (J) Variations in the Earth's rotation as expressed by the m.
near-surface geostrophic winds and Lamb & Johnson (1966) give such indices for various locations in both hemispheres from about 1800 onwards although more global data are available only from about 1860 onwards. Lamb & Johnson (1966) also give the variation in latitude of some of the principal pressure extrema. B. L. Dzerdzievski (Lamb 1972, p. 272) gives an index of the frequencies of certain zonal-circulation patterns in the northern hemisphere. A similar zonal index is given by A. A. Girs (Lamb 1972, p. 451)—both these indices are available only from 1900 onwards. Local indices of the frequency of westerly winds are given by Lamb (1972) for England and these parallel the Girs and Dzerdzievski indices for the period 1900-1960. Mean global temperatures going back to about 1890 are given by Mitchell (Strahler & Strahler 1973, p. 149) and further measures of air temperature going back to the nineteenth century and earlier can be deduced from the variable concentration of oxygen isotopes $^{18}O/^{16}O$ observed in the Greenland icesheet (Johnsen, Dansgaard & Clausen 1970). Variations of the mean temperature over central England are given by Manley (1953) from 1698 onwards. As for the wind indices, the regional- and global-temperature curves show important similarities for the overlapping period suggesting that Manley's and Johnsen's curves do reflect global phenomena. Other northern hemisphere indices showing similar trends include the north–south movement of Arctic sea ice—L. Koch (Lamb 1972, p. 302) and the growth rate of trees in northern Scandinavia and in the polar Urals (Adamenko 1963).

Apart from the pressure differences and temperature observations, information on southern-hemispheric climatic changes are limited but they show trends that are similar to those in the northern hemisphere further testifying to the global extent of the climatic changes. This is seen, e.g. in rainfall indices of Santiago, Chile, given by E. Taulis (Lamb 1972, p. 302); in snow accumulation at the South Pole, and ice movements in the Weddell Sea—J. Fletcher (Lamb 1972, p. 337). Fig. 2 illustrates a number of these climatic indicators. We are not concerned here with the relations that exist between the various observed quantities and we are interested only in their long-term trends as possible indicators of global climatic change. Not all indices give clear trends and they are not always in phase but from the ensemble the following emerge.

The years 1900–1930 are typical of the type I circulation, being characterized by an increasing strength of the zonal circulation at mid latitudes that reach a maximum for the years 1930–1935. There is a poleward shift by a few degrees of all the pressure–extrema belts and a global-surface warming—a secondary maximum occurred as late as 1950. Subsequent years are of the type II circulation with an increasing weakening of the circulation becoming persistent and global by the 1960's. Before 1900 the indices are less reliable but the following global pattern emerges (Lamb 1972): the years 1870–95 are of type II, with the zonal circulation being weak: the years 1840–1870 are an interval of increasingly vigorous zonal circulation (type I) while the period 1820–1840 is characteristic of the type II circulation. Indices of the Arctic Sea ice and of the high-latitude tree growth rates suggest that this last period may have commenced as early as about 1790 and that it was preceded by type I circulation from about 1700 onwards. These dates are of course quite approximate and it does not appear possible to establish relative amplitudes of the intensity of the global circulation patterns for the different intervals in which either type of circulation dominated. A general trend of circulation changes is however indicated and is sketched schematically in Fig. 3. If we assume that the wind circulation is related in some way to the variable rotation of the Earth then this figure should be proportional to the time derivative of the excitation function $\Psi$. If we assume that $\Psi$ is positive for type I circulation, the agreement with $\dot{m}$ is strongly suggestive of a causal relationship, but a quantitative analysis is not possible with the limited nature of the meteorological data set, and without a clear understanding of the relationship between the meteorological indices and the excitation function.
In meteorological discussions, the Earth’s rotation is usually held to be constant so that angular momentum in the atmosphere must be conserved. Thus with the poleward shift of the main zonal streams the tropical easterlies are expected to decrease in amplitude as they will cover an increasingly larger surface area. Independently, the increasing strength of the westerlies requires a corresponding increase in the strength of the easterlies. If we assume that these circulation changes cause the perturbations in the LOD the above correlation between $\dot{m}$ and the two circulation types however suggests an incomplete balance between easterly and westerly winds: the angular momentum of the easterlies apparently increasing more in amplitude than that of the westerlies and there is a net angular-momentum transfer between the solid Earth and the atmosphere.

**Evaluation of the meteorological function**

In the absence of actual wind data the excitation function can be evaluated from pressure data using the geostrophic assumption. Ground-level pressure alone is available and only near-surface geostrophic winds can be computed. Limited global-pressure data extends back to 1846 but only from about 1890 onwards is there a sufficient number of stations reporting monthly mean-pressure values to enable the excitation function to be evaluated with some reliability. The data set used is that compiled by the National Center for Atmospheric Research (NCAR), Boulder, Colorado.

For the geostrophic condition, the pressure gradient is related to the zonal-wind velocity $v_z$ by

$$ \rho v_z = \frac{-1}{2R\Omega \sin \Phi} \frac{\partial p}{\partial \Phi} $$

and the excitation function (equation 2) becomes

$$ \Psi^A(H) = \frac{R^2}{\Omega^2 I_{33}} \int \frac{\cos^2 \Phi}{\sin \Phi} \frac{\partial p}{\partial \Phi} d\Phi dh. $$
Fig. 4 illustrate $\Psi^A(H)$ for the two hemispheres and for the total geostrophic near-surface winds with $dh = 3$ km. The $\Psi^A(H)$ for the two hemispheres show quite similar trends but the northern hemispheric contribution is dominant and lags the southern hemispheric contribution by a few years. The change in $\Psi^A(H)$ is of the order $0.6 \times 10^{-8}$ for the time interval 1910–1950 and represents some 10 per cent of the change in $m$ for the corresponding period. There is an indication that $\Psi^A(H)$ lags $m$ by as much as 10–15 yr. The total $\Psi^A(H)$ can be increased to about $1.0 \times 10^{-8}$ if the ground-pressure data is assumed to be representative of geostrophic winds for the first 5 km of atmosphere rather than 3 km as supposed above. Upper tropospheric winds may contribute further to $\Psi^A(H)$ but their magnitudes are unknown. The 5-yr wind data set used by Lambeck & Cazenave (1974) indicates that most of the power in the low frequency part of the excitation spectrum comes from winds in the lower troposphere below about 5 km.

A second atmospheric contribution to be considered is the excitation $\Psi^A(\Delta I)$ due to the mass transfer in the atmosphere. Munk & MacDonald (1960) show that the ratio $\Psi^A(H)/\Psi^A(\Delta I)$ depends on the ratio of the wind velocity averaged over all altitudes (weighted by density) to the wind velocity at ground level. For low-altitude winds, as may be the case for the long-period trends, this ratio is small and the

![Fig. 4. Various excitation functions $\Psi$ and $m$. (A) $\Psi^A(H)$ for geostrophic winds in the southern hemisphere. (B) $\Psi^A(H)$ for geostrophic winds in the northern hemisphere. (C) $\Psi^A(H)$ for geostrophic winds in the two hemispheres. (D) $\Psi^A(\Delta I)$ for the variable atmospheric inertia tensor. (E) Mean annual global surface temperatures for the atmosphere. (F) $\Psi^o(\Delta I)$ for the variable Eustatic sea level (curve F1) and the residual excitation function (curve F2) after elimination of a secular trend. (G) The total excitation function equal to $a + b + d + f$. (H) The observed $m$ (solid line) and the total excitation function (broken line) translated by 15 yr. Note the different scales for these two functions.]
Correlation coefficients and phase lags between various excitation functions and the astronomically observed $m$ for the time interval 1870–1970

<table>
<thead>
<tr>
<th>Functions</th>
<th>Correlation coefficient</th>
<th>lag (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$m$, $\Psi$ (winds)</td>
<td>0.66</td>
<td>-15</td>
</tr>
<tr>
<td>$m$, $\Psi$ (inertia)</td>
<td>0.67</td>
<td>-3</td>
</tr>
<tr>
<td>$m$, $\Psi$ (oceans)</td>
<td>0.76</td>
<td>-15</td>
</tr>
<tr>
<td>$m$, $\Psi$ (total)</td>
<td>0.75</td>
<td>-15</td>
</tr>
<tr>
<td>$m$, global temperature (from 1900)</td>
<td>0.85</td>
<td>-5</td>
</tr>
<tr>
<td>North Hemisphere Temperature</td>
<td>0.91</td>
<td></td>
</tr>
</tbody>
</table>

Notes:
The negative phase lag indicates a lag of the meteorological function with respect to $m$. The last line gives the correlation between the computed and observed surface temperatures based on the hypothesis (b) (see text).

$\Psi^A(\Delta I)$ may become relatively important. We compute $\Psi^A(\Delta I)$ using equation (1) with $\Delta q = \Delta p/g$ where $\Delta p$ is the variation in the ground level pressure and $g$ is gravity. Only the zero and second-degree zonal harmonics in $\Delta p$ will contribute to $\Psi^A(\Delta I)$. Any yielding of the solid Earth under this surface load will reduce the excitation function but this modification has been ignored since $\Psi^A(\Delta I)$ is already quite small when compared with the $\Psi^A(H)$. The ocean response to the pressure variations should also be considered, a change in sea level atmospheric pressure modifying the level of the ocean surface (see Munk & MacDonald 1960). This correction has not been applied either since the additional excitation from observed sea-level variations, containing in part this response, has also been computed (see below). The trends in $\Psi^A(\Delta I)$ (Fig. 4) follow closely those of $\Psi^A(H)$ but they amount to only about 30 per cent of $\Psi^A(H)$. Thus the total atmospheric-excitation function is dominated by the northern-hemisphere surface winds. The functions $\Psi^A(H)$, and their sum all lag $m$ by about 15 yr. The correlation coefficient between these functions and $m$ is of the order 0.7–0.8 (Table 1).

The global mean atmospheric temperature $T$, computed from monthly mean surface temperatures also compiled by NCAR, gives the curve shown in Fig. 4 and follows closely that computed by Mitchell. This curve again shows similar trends to the excitation functions $\Psi^A(H)$ and $\Psi^A(\Delta I)$ as well as to the astronomical $m$ although the correlation coefficient between $T$ and $m$ is somewhat higher than for the other correlations and the lag of $T$ with respect to $m$ is less than that of the excitation functions (Table 1).

Oceanic effects

Changes in sea level affect the Earth’s rotation only if these changes result in a modification of the second-order inertia tensor. Long term variations, greater than a few years duration, can result from various factors, including:

(i) variations in the amount of ice stored in the polar ice caps and mountain glaciers;

(ii) variations in the mean atmospheric pressure and the wind stresses on the ocean surface;

(iii) variations in the mean water temperature and other oceanographic factors such as salinity and ocean currents.
(iv) long-period tides;
(v) changes in the holding capacity of the ocean basins, due to subsidence or uplift of the sea floor;
(vi) subsidence or uplift of the coastlines resulting in apparent changes in mean sea level.

Only those factors that result in changes in the inertia tensor through a horizontal redistribution of mass are of interest. The 18.7-yr tide arising from the principal nutation term in the lunar motion has an equilibrium amplitude of

$$\xi = 18.4 (1 + k_2 - h_2) (\sin^2 \Phi - \frac{1}{2}) \text{ mm}$$

and the real tide may follow the equilibrium form (Proudman 1960) although observational evidence for this is sparse (Rossiter 1967). The effect of this tide on the LOD is an order of magnitude smaller than the corresponding body tide which itself perturbs the $m$ by only $1.6 \times 10^{-9}$ (Woolard 1959). If the station distribution is truly global the effect of the tides on the mean sea-level curve will vanish. Other than tectonic changes on a geological time scale, (v) and (vi) are important now due to the post-glacial adjustment of the Earth. Melting of the late Quaternary icecaps has resulted in a very significant redistribution of mass, to which the Earth's rotation would long ago have adjusted were it not for the backsurge flow in the asthenosphere which is still very active today (see, e.g. the collection of papers edited by Andrews 1974) and to which the Earth's secular, non-tidal acceleration has been attributed (Dicke 1969; O'Connell 1971).

Changes in sea level due to temperature variations can be important but will not modify significantly the inertia tensor since the mass displacements are essentially vertical and small. Changes in Atlantic mean surface temperatures of about $0.05 \text{ yr}^{-1}$ have been recorded (Wahl & Bryson 1975) but these appear to be restricted to the upper 100–200 m of water. The accompanied sea-level changes are of the order of 1–2 mm yr$^{-1}$. These changes also relate to the main circulation patterns and their effect on the sea-level curve will be less than these values but during the type I circulation, an increase in global mean sea level can be expected, while during type II circulation sea level will tend to decrease. Changing global temperatures can also lead to sea-level changes through a modification of the amount of ice stored in the polar caps, but the nature of the response is not immediately evident: increasing temperatures can increase precipitation and increase the amount of water stored in the polar caps or they can increase melting and decrease the amount of ice stored.

Changes in atmospheric pressure of as much as 10 mbars over periods of about 30 yr have been recorded (see for example the pressure indices of Lamb & Johnson 1966), corresponding to a change in sea level of 10 cm if the ocean response is that of an inverted barometer. These pressure variations are related to the north–south migration of the main circulation streams and their contribution to the global sea level will be reduced if the sea-level observations are globally distributed. A complete discussion requires a unified treatment of the air pressure, sea level and solid Earth interaction—outlined by Munk & MacDonald (1960). However this has not been done here as the oceanic excitation function is already quite small when compared with $m$, and as the $\Psi^A(\Delta I)$ has not been corrected for the sea-level response either, the two effects will largely cancel. Wind-stress effects may make important contributions to the sea-level observations in shallow seas but their effect on the global sea level should vanish.

Annual mean values of salinity or density are poor and variations in their global mean values are presumed to be small.

Several recent analyses of sea-level variations have been made. For European waters the most complete study is that by Rossiter (1967); for North American waters that by Hicks (1971). Both studies confirm that sea level has been increasing by about
I mm yr\(^{-1}\) but both results are heavily influenced by the post-glacial rebound that occurs in both these regions. This rebound can be either positive or negative depending on the position of the forebulge produced by the elastic upward bending of the lithosphere subject to the ice load (see, e.g. Walcott 1970). Fairbridge & Krebs (1962) have updated Gutenberg's (1941) analysis and give the global sea-level curve for the years 1863–1963. Their analysis is based on tide-gauge records for the Pacific, Indian and Atlantic Oceans, but the data for the first two oceans and for the South Atlantic is sparse and their world curve is dominated by the North Atlantic data, reflecting, at least partially, meteorological conditions with a time scale characteristically of about 20 or 30 yr.

A second approach to estimating changes in sea level is through the analysis of the ice and snow budgets of the polar regions. Since Munk & MacDonald's discussion, several new studies of the ice storage have been published but there has not been any significant improvement in the estimates of the resulting change in sea level—as the problem is one of estimating the small difference between the uncertain rates of accumulation and ablation. The various estimates of the ice and snow budgets for Greenland and Antarctica do not even agree in whether there is a net loss or a net gain, although both ice sheets appear to have been quite stable and surrounded by receding outlets glaciers (see, e.g. Lliboutry 1965). For Greenland which contains 9 per cent of the total land snow and ice, Bauer (1966) estimates a net loss of 0.12 \times 10^{18} \text{ cm}^3 \text{ yr}^{-1}. H. Bader (Lliboutry 1965, p. 487) places an upper limit of a net gain of 0.12 \times 10^{18} \text{ cm}^3 \text{ yr}^{-1}. Bauer’s value will lead to a rise in sea level of 0.3 mm yr\(^{-1}\) while Bader's estimate results in a fall in sea level by the same amount. Antarctica, containing about 90 per cent of the total land ice likewise leads to conflicting values. F. Loewe (Strahler & Strahler 1973, p. 437) estimates a net gain of about 0.24 \times 10^{18} \text{ cm}^3 \text{ yr}^{-1}. Other estimates, quoted by Lliboutry (1965, p. 513) are 0.5 \times 10^{18}—Dolgouchine; –0.4 \times 10^{18}—Buitnisky; and 0.1 \times 10^{18}—d’Averyanov. These values result in changes in sea level that range from +1.4 to –1.1 mm yr\(^{-1}\). Mountain glaciers other than in polar regions contain only about 1 per cent of the total land ice but this amount is much more variable than the polar ice and may make a significant contribution to any change in sea level. Observations of European and other glaciers suggest that there has been a global recession for at least the last 100 yr (see Lliboutry 1965, for a review). Finsterwalder (1954) estimated that the secular loss of eight typical glaciers in the Eastern Alps for the years 1856–1950 is of the order of 500 mm yr\(^{-1}\) (unit of surface area). Superimposed upon this are fluctuations of the order of 200 mm yr\(^{-1}\) (unit of surface area) which follow the observed temperature fluctuations on a time scale of 20–40 yr. According to Manley (1954) these glacier loss–gain variations follow closely those observed in Iceland and Norway (see also, Le Roy Ladurie 1973). Assuming that these rates are typical of all European glaciers of total surface area 4 \times 10^5 \text{ km}^2 the expected rise in sea level is of the order of 0.5 mm yr\(^{-1}\). The regional trends for Europe are also reflected by glaciers in North America and Asia. If similar loss rates occurred everywhere the rise in sea level would be of the order of 1 mm yr\(^{-1}\), similar to the observed rate. This would represent an upper limit since not all glaciers will exhibit the same loss rates, nor will the loss–gain rates always be in phase but it does point to an important role of loss–gain rates of non-polar glaciers in eustatic sea-level variations.

Sea-level response to the fluctuations in the glaciers—if they are indeed of global extent—would be one of receding level up to about 1880 and one of increasing level from this epoch until the present. Fluctuations of 20–30 yr duration could be superimposed on this trend. The global sea-level curve of Fairbridge & Krebs (1962), shows such a general increase from 1880 to 1960, suggesting that this variation in level is related to the variable land ice storage rather than being of a direct meteorological origin whose fluctuations are more rapid and which dominate longer term variations. If the secular trend for this period is eliminated the shorter-period variations become
evident and follow the global atmospheric temperature curve, suggesting that these sea level fluctuations are real (Fig. 4—curve F). As the direct effects, primarily the rise in sea level due to temperature variations, will presumably be in approximate phase with eustatic sea level, the use of the observed sea-level curve will give an upper limit to the oceanic excitation function.

Munk & Revelle (1952) and Munk & MacDonald (1960) derive the excitation function \( \Psi^0 \) resulting from a redistribution of mass associated with the variable storage in the ice caps. If \( \xi' \) is the net rate of increase of the ice or snow level in Greenland, \( \xi'' \) that in Antarctica and \( \xi \) the rate of increase of sea level then

\[
\Psi^0(\Delta I) = \frac{8\pi R^4}{5} \rho_w (1 + k'_2) \xi \left( a'_{20} \xi + a''_{20} \xi'' + a''_{00} \xi'' \right) \cdot a_{00}
\]

\( \rho_w \) is the density of the ocean. The \( a_{nm}, b_{nm} \) are the spherical harmonic coefficients of the ocean land distribution, \( a'_{nm}, b'_{nm} \) are the coefficients of the Arctic ice spherical harmonic expansion, \( a''_{nm}, b''_{nm} \) are the coefficients of the Antarctica ice. The harmonic coefficients of the land—sea distribution of Balmino, Lambeck & Kaula (1972) are used. The ice functions have also been re-evaluated and the zero- and second-degree coefficients are given in Table 2. Mountain glaciers being of small areal extent will contribute to \( \Psi^0 \) only through \( \xi \). The factor \( 1 + k'_2 \) (with \( k'_2 = -0.30 \)) allows for the Earth's elastic response to the time-variable surface load and as the variations studied are very short compared with a characteristic time of the viscous response of the Earth of about 2000 yr (Cathles 1975) an elastic response appears adequate. Numerically \( \Psi^0(\Delta I) \approx -0.76 \times 10^{-9} \xi \) where \( \xi \) is expressed in cm/unit of time. Changes in \( m_3 \) of \( 5 \times 10^{-8} \) as observed from 1905–1935 therefore require a drop in sea level of about 60 cm if these changes in the LOD are caused by a redistribution of mass between the polar land ice and the world's oceans. Observations of sea level on the other hand indicate an increase in sea level of about 3 cm for this period. Clearly the observed changes in sea level do not make an important contribution to the sought excitation. Fig. 4 illustrates the sea level excitation function \( \Psi^0(\Delta I) \) using the Fairbridge & Krebs (1962) global sea-level curve. From 1890 to 1960 there appears to be a secular decrease of about \( -3.2 \times 10^{-18} \) s\(^{-1}\) or a total change of \( -0.7 \times 10^{-8} \) over this 70-yr period. Superimposed on this are variations with a magnitude of less than a few parts in \( 10^9 \) and which appear to be out of phase with the \( \Psi^0(\Delta I) \) (Fig. 4) suggesting that the sea-level fluctuations are in part of meteorological origin; an increase in temperature increasing the sea level curve, or an increase in the atmospheric inertia tensor decreasing the oceanic inertia tensor.

**Discussion**

Figure 4 gives the total excitation function \( \Psi^T \) including the atmospheric and oceanic contributions discussed above. Only from about 1890 and onwards can this function be considered to be reliable. As already indicated \( \Psi^T \) is dominated by the geostrophic surface winds in the northern hemisphere. The correlation coefficient

### Table 2

**Spherical harmonic coefficients (unnormalized) of the ocean distribution, Greenland ice and Antarctic ice**

<table>
<thead>
<tr>
<th></th>
<th>Ocean</th>
<th>Greenland ice</th>
<th>Antarctic ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>( a_{00} )</td>
<td>0.697</td>
<td>0.007</td>
<td>0.039</td>
</tr>
<tr>
<td>( a_{20} )</td>
<td>-0.134</td>
<td>0.030</td>
<td>0.170</td>
</tr>
<tr>
<td>( a_{21} )</td>
<td>-0.051</td>
<td>0.006</td>
<td>-0.001</td>
</tr>
<tr>
<td>( b_{21} )</td>
<td>-0.065</td>
<td>-0.005</td>
<td>-0.004</td>
</tr>
</tbody>
</table>
between $\Psi^T$ and $m$ is 0.75 with $\Psi^T$ lagging $m$ by about 15 yr. Fig. 4 also illustrates $m$ and $\Psi^T$ with the latter shifted by this lag. The changes in $\Psi^T$ corresponding to the years 1905–1935 and 1870–1900 represents about 10 per cent of $m$. An increase in the computed meteorological excitation function by the required order of magnitude is possible only if there have been very important changes in the upper atmospheric circulation over periods of about 10 yr and longer—changes that are proportionally more important than those observed at low altitudes. There does not appear to be any observational data available pertaining to such long-period changes. If the geostrophic winds are representative of the circulation in more than the first 3 km of atmosphere the total excitation function can be increased by a factor of about 2 giving a total excitation of the order of 20 per cent of $m$. Two observations of limited reliability indicate that this extrapolation may not be valid. The first is that the limited 5-yr wind data set utilized by Lambeck & Cazenave (1974) suggests that most of the low-frequency contribution to $\Psi$ comes from the lower troposphere. Second, the phase lag of about 15 yr between $\Psi$ and $m$, if real, is troubling as it is difficult to imagine such a lag between upper and lower atmospheric circulation at low frequencies. The ocean contribution to $\Psi^T$ could be over estimated since the variation in sea level will include effects that do not change the inertia tensor and as $\Psi$ (ocean) is out of phase with $\Psi$ (atmosphere) the total excitation function could be increased slightly. Other indirect effects appear to be small. Evidence of long-term global fluctuations in ground-water storage is lacking but as the annual variation contributes only about $3 \times 10^{-10}$ to the total seasonal excitation (Lambeck & Cazenave 1973) any long term effects are not likely to exceed a few parts in $10^{10}$. Long-period tides do not exceed 1 in $10^9$, and uncertainties in their estimated effects due to the ocean tide are unlikely to exceed 10 or 20 per cent. Ocean currents also appear unlikely to be important as their seasonal effects on $m$ do not exceed a few parts in $10^{10}$ (Munk & MacDonald 1960; Wilson 1975).

The two principal observations that can be drawn from the preceding sections are:

(i) there is a close similarity between variations in the LOD and trends in various climatic indices for the last 150 yr for which LOD variations have been observed. In particular the LOD variations correlate well with the global temperature $T$ and with ground pressure, both of which are indicators of global wind circulation. Periods of increasing zonal circulation correlate with an acceleration of the Earth while periods of decreasing zonal circulation correlate with a deceleration of the Earth;

(ii) the excitation function $\Psi^T$ computed from surface pressure- and sea-level data represents only about 10 per cent of the amount necessary to explain the LOD variations;

Two further observations, less well established, are:

(iii) Both $\Psi^T$ and $T$ lag $m$ with $T$ lagging less than $\Psi^T$,

(iv) The global mean temperature $T$ correlates somewhat better with $m$ than do either $\Psi^H(A)$ or $\Psi^A(D)$ or the sum of these two partial excitation functions (see Fig. 5).

Cause and effect cannot be distinguished from the first observation alone and three alternative hypotheses are possible: (a) the atmospheric circulation causes the long period changes in LOD as it does for periods less than 2 or 3 yr. (b) the fluctuations in LOD and climatic change are both the consequence of a third phenomena, or (c) the fluctuations in LOD cause the observed variations in the circulation. For hypotheses (a) and (b) the remaining 90 per cent of the excitation remains to be explained; for the first this must be of meteorological origin while for the second a wider range of further excitation is permitted. For the third hypothesis the total mechanism for exciting the LOD changes remains to be explained. The final choice between these hypotheses
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Fig. 5. Variations in \( m \) (solid line) and in global surface temperature \( T \) (broken line) from 1900–1970.

will rest on the plausibility of any explanations for the missing excitation and on any supporting observational evidence of phenomena that may be related indirectly to this mechanism.

As already discussed it seems improbable that other, unevaluated, meteorological factors can furnish the additional excitation. The third hypothesis has been suggested by Maksimov & Sleptsov-Shelevich (1973) who have drawn attention to the similarity between the variation in LOD and changes in atmospheric pressure at 70° north latitude for the years 1900–1960. If such a relation exists the oceans presumably play an important role if the 10–15-yr phase lag is to be explained. The changes in \( m \) can result in a modification of the circulation and mass distribution in two ways. In the first instance a change in \( m \) will cause the shape of the equipotential surfaces to change, secondly there will be a small change in the geostrophic condition. Easier to evaluate is the energetics of the situation. The change in the potential of the centrifugal force is (Lambeck 1973)

\[
\delta U = \Omega^2 R^2 (\cos \phi - \xi) m
\]

and the equilibrium energy of the tide set up in the ocean is (Munk & MacDonald 1960, p. 218)

\[
E = g \rho \int \left( \frac{\delta U}{g} \right)^2 ds \\
= 2\pi \frac{\rho R^6}{g} \Omega^4 m^2 \int (\cos^2 \phi - \xi)^2 \cos \phi d\phi \\
= 0.25 \times 10^{34} m^2 \text{erg}
\]

where the oceans are assumed to cover the Earth. The rate of energy dissipation is given by

\[
\oint \frac{dE}{dt} dt = \frac{2\pi E}{Q}
\]

or

\[
\frac{dE}{dt} \sim 2\pi \sigma E/Q
\]

where \( E \) is assumed to vary periodically with frequency \( \sigma \). Typically \( \sigma \sim 30 \text{yr}^{-1} \) and \( m \sim 5 \times 10^{-8} \). The ocean \( Q \) for the M\(_2\) tide is of the order 35 (Hendershott 1972) while that for the pole tide is about 25 or more (Wunsch 1974). Thus

\[
\frac{dE}{dt} \sim 0.8 \times 10^{16} \text{erg s}^{-1}
\]
and is exceedingly small when compared with the already small amount of about $10^{12}$–$10^{13}$ erg s$^{-1}$ estimated for the pole tide (Wunsch 1974) or with the $5 \times 10^{19}$ erg s$^{-1}$ dissipated by the semi-diurnal and diurnal ocean tides (Lambeck 1975a).

The second hypothesis (b) is perhaps more convincing even if little quantitative information appears to be available. One phenomenon that has been evoked to explain climatic changes is volcanic activity. Volcanic dust injected into the upper atmosphere will persist for several years and, by modifying the Earth's albedo, may cause important meteorological changes (Lamb 1970). In particular, periods of intense activity such as occurred in the 1880's have been used to explain the changing climate. Anderson (1974) has suggested that this activity may also be responsible for the important change in LOD at the turn of the century although quantitative arguments are lacking. If the effect of volcanic dust cannot change the atmospheric excitation function by the required amount the following scenario is still possible. Volcanic activity affects climate leading to the observed but insufficient excitation function. Associated with the volcanic activity are earthquakes which result in a changing inertia tensor of the solid Earth and which would explain the missing 90 per cent of the sought excitation. Stoyko & Stoyko (1969) have already suggested such a correlation between the LOD variations and seismic activity as do more recent and more detailed studies by Anderson (1974, 1975), Press & Briggs (1975) and M. A. Chinnery (1975, private communication). But there are two potential weaknesses in these arguments. First the seismic energy released appears to be inadequate to change the LOD by the required amounts (Smylie & Mansinha 1971; Dahlen 1973) although aseismic, lithospheric motions could be more efficient. Second, the observed phase lag between $m$ and $\Psi^T$ is not explained unless the volcanic activity lags seismic activity by as much as 15 yr.

A second phenomenon sometimes evoked to explain climatic change is the variation in solar activity but the evidence for this is inconclusive (see Lamb 1972, for a review). Also the possible indirect effect of a variable solar-wind torque acting on the Earth appears to be completely inadequate to explain the missing excitation (Coleman 1971; Hirshberg, 1972).

Whatever mechanism is finally proposed it will have to explain the apparently significant lag that is found between the LOD and the various climatic indices, temperature and excitations. The interest of this lag suggests that the LOD observations can be used as an indicator of future climatic trends, in particular of the surface warmings. Without a better understanding of the interactions between the two phenomena the use of the LOD observations in predicting climate is of very limited value but if the hypothesis is accepted then the continuing deceleration of $m$ for the last 10 yr suggests that the present period of decreasing average global temperature will continue for at least another 5–10 yr. Perhaps a slight comfort in this gloomy trend is that in 1972 the LOD showed a sharp positive acceleration that has persisted until the present, although it is impossible to say if this trend will continue as it did at the turn of the century or whether it is only a small perturbation in the more general decelerating trend.

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Contribution I.P.G. N.S. 197

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