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TIDAL FRICTION AND ITS CONSEQUENCES IN PALAEOGEODESY, IN THE GRAVITY FIELD VARIATIONS AND IN TECTONICS

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Abstract—Fossils and tidal deposits as well as the possibility to compute values of the lunar tidal torque for different geological epochs allow us to model the variations in time of the Earth's figure, assuming that the latter remains, on a global scale, close to a hydrostatic equilibrium figure. On this basis we were able to infer the variations of the Earth's most important kinetic parameters over much of the geological past. Thus, the geometrical oblateness of the outer surface has decreased from 0.005 to 0.003 over the last two and a half billion years. This slow but continuous change of the Earth's curvature brought about by tidal friction must have led to continuous stress accumulation in the uppermost part of the lithosphere, where the temperature is below 400°C and the rheological behaviour is likely to remain brittle over geological time scales. We investigate the inevitable tectonic consequences of this stress buildup, and try to find some evidence in present-day worldwide seismicity, with a negative result. An interesting result of our study, which may open a new field of gravimetric research, is embodied in the fact that tidal friction causes a secular increase of the Earth's normal gravity component at the equator at a rate of about 2 ngals yr^{-1}, and a concomitant decrease at the poles of about 0.5 ngals yr^{-1}. This tiny secular signal may just lie within observational reach of superconducting gravimeters. © 1997 Elsevier Science Ltd

CONSERVATION OF ANGULAR MOMENTUM IN THE EARTH-MOON SYSTEM

According to Bretterbauer (1987) a phenomenon relevant to tidal friction seems to have been recognized for the first time in 1695, when Edmund Halley observed the apparent acceleration of the Moon. This observation made him conclude that the Moon did not strictly comply with Isaac Newton's law of motion. Later, in 1754, Immanuel Kant suggested that the recently discovered synchronism of the Moon's orbital and spin motions was a result of tidal friction. Julius Robert Mayer gave in 1848 a qualitative description of tidal friction, apparently without knowing Kant's writings (Münzenmayer, 1982). Quantitative investigations were performed by William Ferrel in 1853 and 1865, by Charles Delaunay in 1865, and especially by George

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Howard Darwin (Darwin, 1908) in a series of outstanding papers written after 1878 (Sündermann and Brosche, 1984).

If we put aside meteorological and geoelectromagnetic phenomena, the observed secular despinning of the Earth is generally explained in terms of a tidal gravitational interaction with the Moon. The study of long-term variations of the length of the day (l.o.d.), and consequently the study of the evolution of the Earth-Moon system, is commonly based on the law of conservation of angular momentum written under the following (simplified) form:

$$\frac{d}{dt}(C\omega+C_m\omega_m+C_s\omega_s)= \tag{1}$$

$$\frac{MM_m}{M+M_m}n_m a_m^2\sqrt{1-e_m^2}\left(\frac{\cos i_m}{3n_m}\frac{dn_m}{dt}+\frac{e_m\cos i_m}{1-e_m^2}\frac{de_m}{dt}+\sin i_m\frac{di_m}{dt}\right)+$$

$$+\frac{M_s(M+M_m)}{M_s+M+M_m}n_s a_s^2\sqrt{1-e_s^2}\left(\frac{\cos i_s}{3n_s}\frac{dn_s}{dt}+\frac{e_s\cos i_s}{1-e_s^2}\frac{de_s}{dt}+\sin i_s\frac{di_s}{dt}\right)+N$$

In this equation, the symbols M,C and ω denote respectively the mass, principal inertia moment about the polar axis and spin of the Earth. The same symbols with subscripts m and s denote the corresponding quantities for the Moon and the Sun, respectively. The parameters n_m, a_m, e_m, i_m are, respectively, the angular speed, the semi-major axis, the eccentricity, the inclination of the lunar orbit, and the parameters n_s, a_s, e_s, i_s are the solar counterparts. The correction term $+N$ on the right hand side accounts for a possible imbalance resulting from non-tidal external torques applied to the Earth-Moon system. According to Kostelecký (1990) $de_m/dt=(0.021\pm0.003)\times10^{-9}yr^{-1}$ and $di_m/dt=(0.78\pm0.07)\times10^{-9}yr^{-1}$. We may safely neglect the terms which depend on these small quantities. Moreover, partly because we lack the necessary information for more sophisticated modelling in the geological past, partly because sophistication is not needed for our purpose, we assume that the Moon revolves around the Earth on a circular orbit in a plane which coincides with the terrestrial equator: $e_m=i_m=0$. As a consequence of the minor effect which the Sun plays in the tidal evolution of the Earth-Moon system over a time interval of about three billion years, we neglect the solar terms in Eqn. (1). Finally, we note that $C_m\omega_m<C\omega$, and therefore we omit $dC_m\omega_m/dt$ in Eqn. (1), taking for granted that the Moon had already achieved synchronous rotation 3×10^9 years ago. Referring to the quantity

$$L=\frac{1}{3}\frac{MM_m}{M+M_m}a_m^2\frac{dn_m}{dt} \tag{2}$$

as the tidal torque, we therefore express the conservation of angular momentum in the simple form

$$C\frac{d\omega}{dt}+\omega\frac{dC}{dt}=L+N \tag{3}$$

Denis (1986) has thoroughly discussed the usual assumption made in studies related to the tidal evolution of the Earth-Moon system, *i.e.* $dC/dt\approx0$. This assumption is essentially based on the hypothesis (or "theory" according to many authors) that the Earth's internal structure remained basically unchanged over the last 4.2×10^9 years. In particular, it is believed that the formation of the Earth's core was finished soon after the formation of the Earth itself (Varga and Denis, 1990; Le Mouël *et al.*, 1994).

Having formulated the principal physical law which is being used over and over again in investigations on tidal evolution, we shall now have a short look at the sparse data which can be used as constraints for quantitative modelling.

DATA SOURCES FOR THE STUDY OF LONG-TERM VARIATIONS

Information about long-term variations in the Earth's spin can be obtained from several sources, but it remains sparse and to some extent controversial, especially for epochs earlier than the Phanerozoic. Modern astronomical data yield evidence that the terrestrial rotation has slowed down over the last three centuries at an average rate of $(-5.4 \pm 0.5) \times 10^{-22}\ rad\ s^{-2}$ (Lambeck, 1977), amounting to an increase of $(2.0 \pm 0.2)\ ms\ cy^{-1}$ in l.o.d. This spindown rate agrees perfectly with the value found earlier by Pariisky (1959), and with the value $d\omega/dt = -5.6 \times 10^{-22}\ rad\ s^{-2}$ given more recently by Brosche (1987). The latter is identical with the secular despinning rate determined by Stephenson (1978) from early information on ancient eclipses. It does not agree too well, however, with the value $d\omega/dt = -6.07 \times 10^{-22}\ rad\ s^{-2}$ obtained by Christodoulis *et al.* (1988) by means of satellite laser ranging, nor with the value $d\omega/dt = -5.98 \times 10^{-22}\ rad\ s^{-2}$ found by Newhall *et al.* (1986) by means of lunar laser ranging. In a paper which appeared quite recently, Stephenson and Morrison (1995) analysed all the reliable historical eclipse observations in the period 700 B.C. to A.D. 1990 which are pertinent to the question of the long-term variability of terrestrial rotation, and reached the conclusion that the average increase in l.o.d. over the past 2700 years is $(1.70 \pm 0.05)\ ms\ cy^{-1}$. It corresponds to $d\omega/dt = (-4.5 \pm 0.5) \times 10^{-22}\ rad\ s^{-2}$. This rate is about 20% lower in absolute value than earlier estimates. The present understanding is that this average observed value is the net result of two different physical phenomena. The first is tidal dissipation, which produces an increase in l.o.d. by $(2.3 \pm 0.1)\ ms\ cy^{-1}$, and corresponds to a despinning rate $d\omega/dt = (-6.1 \pm 0.4) \times 10^{-22}\ rad\ s^{-2}$. The second is of non-tidal origin and produces a decrease of l.o.d. at the average rate of $-(0.6 \pm 0.1)\ ms\ cy^{-1}$. The corresponding upspinning rate is $d\omega/dt = (+1.6 \pm 0.4) \times 10^{-22}\ rad\ s^{-2}$. It seems tempting to ascribe the latter wholly or partly to post-glacial rebound. Moreover, Stephenson and Morrison (1995) found a long-term fluctuation in l.o.d., with a semi-amplitude of about 4 ms and a period of roughly 1500 yr. They tentatively assume that this variation is caused by phenomena occurring in the Earth's liquid core which affect the solid mantle through core–mantle coupling.

The astronomical data are useful for geodynamic studies only insofar as one admits that they can be extrapolated to remote epochs in the past. For the modelling we have in mind in this paper, we must rely on palaeontological and palaeosedimental data. The relevant information comes from fossils (bivalves, brachiopods and corals) of the Phanerozoic (Pz), from stromatolites mainly of the Proterozoic (Ptz), and from palaeodeposits of the Proterozoic. The study of annual and daily growth rings in corals was pioneered by Wells (1970), and evidence of a monthly periodicity was pointed out by Scrutton (1970). An early review of this work in connection with Earth rotation was provided by Stoyko (1970). Later investigations on Carboniferous corals was performed by Johnson and Nudds (1975), and extended to bivalves and stromatolites by Berry and Barker (1968, 1975), Mazullo (1971), Mohr (1975), Pannella (1972, 1975, 1976), and Scrutton (1978). A summary of some of these data can be found in Lambeck (1978). Important work on tidal rhythmites, or tidalites, was published by Williams

(1989a,b) and very recently by Sonett *et al.* (1996). Complementary information pertaining to the latter topic is provided by Piper (1990) and Richardson (1990).

Despite this long list of references, reliable data concerning l.o.d. in the geological past are sparse. Moreover, palaeontologists—unlike physicists or geodesists—are hardly used to provide reliable error bounds. Nevertheless, the data plotted in Fig. 1 are probably the best quantitative empirical evidence we have for modelling the evolution of the Earth-Moon system over geological time spans. Moreover, they have a direct bearing on the tectonic evolution of our planet.

Lambeck (1980) performed a linear regression on Phanerozoic l.o.d. data, and obtained a

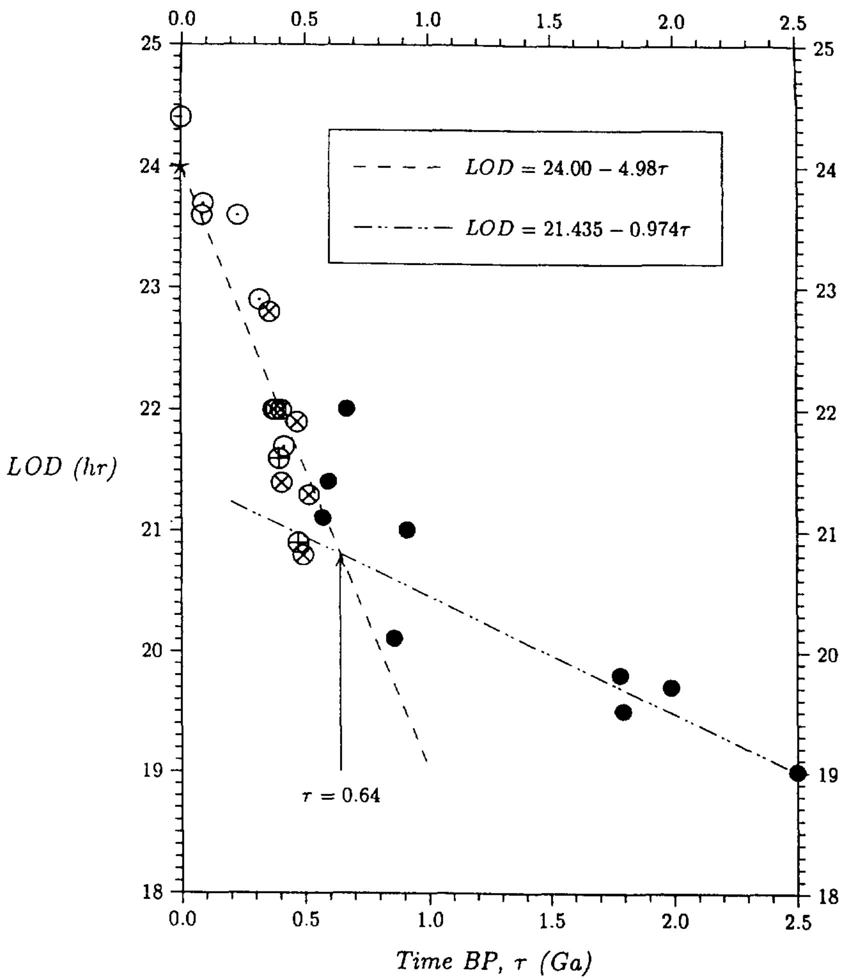

Fig. 1. Variation of l.o.d. during the Phanerozoic and the Proterozoic. A star (*) denotes an astronomical datum, the symbols $\odot$, $\otimes$, $\oplus$, refer respectively to bivalve, coral and brachiopod data, and a bullet (•) denotes information from stromatolites or tidalites. Note that no error bars are given. These are often difficult to obtain from the relevant literature, but in any case they are quite large both in abscissae and in ordinates.

(constant) despinning rate of $-5.4 \times 10^{-22}\ rad\ s^{-2}$. This despinning rate is obviously too large, in absolute value, to be typical of the Proterozoic, for it would yield, three billion years ago, a l.o.d. value shorter than the Poincaré limit (1.15 hour) and an Earth-Moon distance smaller than the Roche limit (18,500 km). The geological record shows no trace of a dramatic event which would necessarily have occurred under such critical circumstances.

Figure 1 suggests that the despinning rate of the Earth has indeed not been constant over the geological past. On the average, it seems to have been smaller than the present rate. Moreover, it appears clearly that the mean despinning rate was smaller in the Proterozoic than in the Phanerozoic. The linear trend in the variation of l.o.d. in the Phanerozoic may be modelled as

$$LOD = 24.00 - 4.98\tau \tag{4}$$

where τ is the time before present expressed in 10^9 years. This law is in essential agreement with Lambeck's value for the tidal despinning rate. On the other hand, for the linear trend in the Proterozoic we suggest tentatively

$$LOD = 21.435 - 0.974\tau \tag{5}$$

An investigation of the l.o.d. data using robust statistical modelling techniques (Lecoutre and Tassi, 1987) performed by Varga *et al.* (1992), also reported in Varga *et al.* (1993), shows indeed that

—the despinning rate during the Proterozoic was on the average about 7.5 times smaller than during the Phanerozoic;
—a statistically significant relative minimum took place in the Earth's despinning rate during the Phanerozoic between 350 and 150 million years ago;
—the average despinning rate during the Phanerozoic was similar to its present value.

Varga *et al.* (1992) tried to explain the non-linear behaviour of the rotation law in terms of a physical mechanism. Using the fact that presently the main contribution to the tidal torque L comes from *oceanic* tides—specifically from the oceanic M_2-tide which accounts for about 84% of the total oceanic tidal torque (Varga and Denis, 1990)—they were able to correlate the despinning rate with the M_2 oceanic tidal torque. The tidal palaeotorques were estimated on the basis of 17 palaeocotidal maps for M_2. The latter had been computed for oceans bordered by shorelines reconstructed from palaeogeographic data for different geological epochs by Krohn and Sündermann (1982) and by Gotlib and Kagan (1985). When continents are gathered, as occurred for instance at the time of Pangaea some 250 million years ago, diurnal tides are enhanced with respect to semi-diurnal tides. This corresponds to a relatively small value of the oceanic torque, and thus of L, which can be connected with the relative minimum of tidal despinning which has occurred between -350×10^6 and -150×10^6 years. On the other hand, when several continents are spread over the Earth's surface (e.g. at the present epoch or in the early Cambrian), semi-diurnal tides are enhanced, the tidal torque is largest and tidal despinning is most effective.

The result concerning the low despinning rate in the Proterozoic solves the problem of the Moon having been ever too near to the Earth, but is rather paradoxical if we consider that the Moon has nonetheless been closer to the Earth in the past. There must in principle have been more tidal dissipation, not less. At least two compensating mechanisms may be invoked, but both of them are liable to be criticized. The first involves the idea that the world ocean was much less deep two or three billion years ago than it is now, the second idea (which we favour for the moment) is that the formation of the core was not completed entirely soon after the Earth itself was formed. Thus, iron has continually been transferred from the mantle to the core through a core–mantle boundary expanding in time. This idea of a core growing throughout geological time had first been suggested by Urey (1952 pp. 87–92) and was later advocated by Runcorn (1962, 1964 and 1965). It is not readily accepted anymore because geochemical and cosmochemical arguments (Oversby and Ringwood, 1971; Anderson and Hanks, 1972) seem to indicate that core formation is a run-away process leading to completion soon after planetary formation. Nevertheless, to some extent, ongoing growth of the core may be considered as a thermodynamic necessity (Majewski, 1995).

PALAEOGEODETIC CONSEQUENCES OF TIDAL FRICTION

On the basis of palaeontological and palaeosedimental evidence it makes sense to adjust some curve by mean-squares to the palaeo-*LOD* values. Although the scatter in the data is admittedly quite large, we believe that the rotation law embodied in Eqns. (4) and (5), namely $LOD=24.00-4.98\ \tau$ if $\tau<0.64$, $LOD=21.435-0.974\ \tau$ if $\tau\geq 0.64$, represents a plausible average time trend of the length-of-day values for the last 2.5×10^9 years. Here, *LOD* is expressed in hours, the time τ in 10^9 years. In what follows, we shall use this law for illustrative purposes, noting that the exact despinning rates are not essential for most of the conclusions we shall draw in this paper. However, an important consequence of the data plotted in Fig. 1 is that the Moon has probably never been very close to the Earth, its orbital radius remaining within about 75% of the present value (Varga *et al.*, 1991). This inference agrees with the conclusion reached by Deubner (1990).

We are thus in a position to calculate the variation of the Earth's kinetic parameters over a large part of the geological history assuming that for the time scales considered, the Earth's evolving shape can be conceived as a succession of hydrostatic equilibrium figures (Denis, 1986). We computed these palaeokinetic parameters by means of a proprietary numeric code based on the algorithm explained in Denis (1989) (see also Denis *et al.*, 1996a). This code actually solves the Clairaut-Laplace-Lyapunov integro-differential system in the third-order approximation.

Assuming that the mass distribution within the Earth did not significantly change over the last two and a half billion years—assuming in particular that core formation was essentially completed 2.5×10^9 years ago and that the Earth's total mean radius R=6,371,000 metres did not change appreciably—we computed for the Earth model CGGM (Denis *et al.*, 1996b) the values of the geodynamic constant $m=\Omega^2R^3/GM$, of the surface flattening $f=(a-c)/a$, of the precessional constant $H=(C-A)/C$ and of the dynamic shape factor $J_2=(C-A)/Ma^2$, for spin periods $T=2\pi/\Omega$ ranging from 18 to 24 hours. Here, $GM=3.986\times 10^{14}\ m^3\ s^{-2}$ is the geocentric gravitational constant, M is the Earth's total mass, a and c are respectively the equatorial and the polar radii, C and A are principal moments of inertia, respectively, about the polar axis and about an equatorial axis. The spin period is related to the actual l.o.d. value by the relation $T=0.99727$ *LOD*. Table 1 shows the results obtained by these model computations. Because the surface flattening depends on the internal density structure only by terms which are of the second order of smallness (Denis, 1989), our assumption concerning the formation of the core is not crucial for the values stated in Table 1, provided R has remained nearly constant. However, the corresponding values of f and H for the inner core boundary (ICB) and the core-mantle boundary (CMB), provided in Table 2, depend rather critically on the assumption that the core structure remained essentially the same throughout the considered time span. Figure 2 shows the variation of the reciprocal flattening at the surface, at the CMB, and at the ICB, as a function of geological

Table 1. Hydrostatic parameters as a function of the rotation period T, computed for Earth model CGGM of Denis *et al.* (1996b)

T(hours)	$10^3 \times m$	$10^3 \times f$	$10^3 \times H$	$10^3 \times J_2$
18.0	6.100	5.933	5.763	1.915
18.5	5.774	5.616	5.456	1.813
19.0	5.474	5.324	5.173	1.719
19.5	5.197	5.055	4.912	1.632
20.0	4.941	4.805	4.670	1.551
20.5	4.703	4.573	4.445	1.477
21.0	4.481	4.358	4.236	1.407
21.5	4.275	4.157	4.042	1.343
22.0	4.083	3.970	3.860	1.282
22.5	3.904	3.796	3.691	1.226
23.0	3.736	3.632	3.532	1.173
23.5	3.579	3.479	3.383	1.124
24.0	3.431	3.336	3.244	1.078

time τ, supposing that the rotation law assumed here is indeed representative of the average values of l.o.d. in the remote past.

Denoting latitude by φ, we may write in a first-order approximation for the normal gravity field on the Clairaut spheroid (Heiskanen and Vening Meinesz, 1958 p. 49; Stacey, 1992 p. 93)

$$\gamma_0 \approx \gamma_e \left[1 + \left(\frac{5\Omega^2 a^3}{2GM} - f \right) \sin^2\varphi \right] \tag{6}$$

where in the same approximation the equatorial gravity γ_e is given by

Table 2. Hydrostatic parameters as a function of the rotation period T, computed at ICB and CMB for Earth model CGGM of Denis *et al.* (1996b)

T(hours)	$10^3 \times f_{ICB}$	$10^3 \times f_{CMB}$	$10^3 \times H_{ICB}$	$10^3 \times H_{CMB}$
18.0	4.439	4.584	4.429	4.571
18.5	4.202	4.340	4.193	4.328
19.0	3.983	4.114	3.975	4.103
19.5	3.781	3.905	3.774	3.896
20.0	3.594	3.713	3.588	3.704
20.5	3.421	3.533	3.415	3.525
21.0	3.260	3.367	3.255	3.360
21.5	3.110	3.212	3.105	3.205
22.0	2.970	3.068	2.966	3.061
22.5	2.839	2.933	2.835	2.927
23.0	2.717	2.806	2.713	2.801
23.5	2.603	2.688	2.599	2.683
24.0	2.495	2.577	2.492	2.573

$$\gamma_e \approx \frac{GM}{a^2}\left(1+f\frac{3\Omega^2 a^3}{2GM}\right) \tag{7}$$

Moreover, $\Omega^2 a^3/GM \approx \Omega^2 R^3/GM = m$, $a = R(1+f/3)$. Thus, in terms of the quantities R and m used before, γ_e may be written

$$\gamma_e \approx \frac{GM}{R^2}\left(1+\frac{f}{3}-\frac{3m}{2}\right) \tag{8}$$

and Eqn. (6) becomes

$$\gamma_0 \approx \frac{GM}{R^2}\left[1+\frac{f}{3}-\frac{3m}{2}+\left(\frac{5m}{2}-f\right)\sin^2\varphi\right] \tag{9}$$

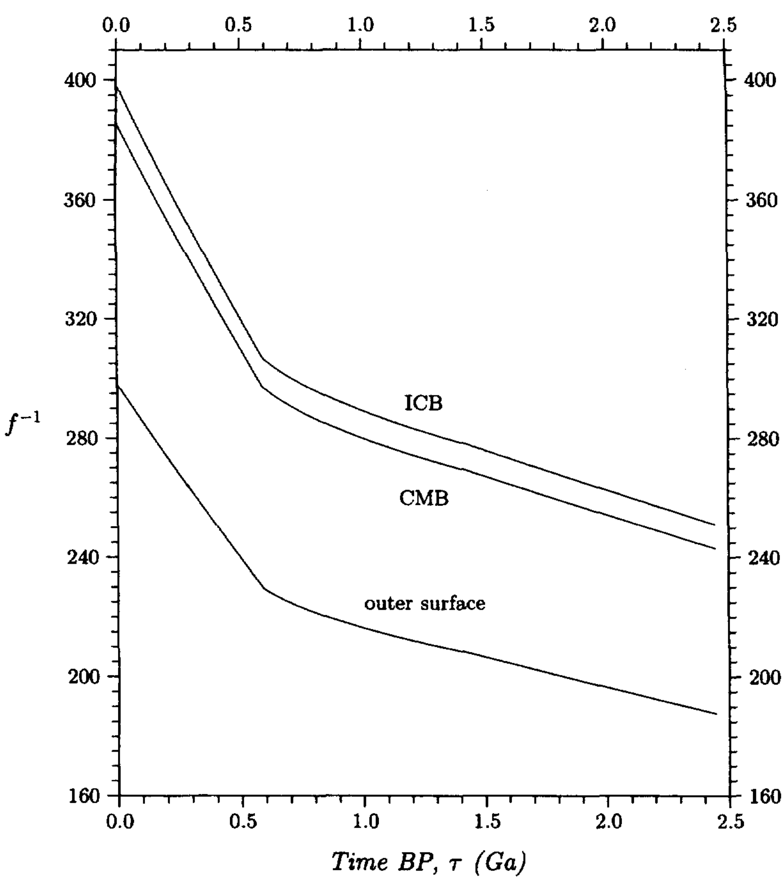

Fig. 2. Inferred reciprocal flattening at the outer surface (f^{-1}), at the inner core boundary ((f^{-1}_{ICB})) and at core-mantle boundary ((f^{-1}_{CMB})) throughout geological time.

We have taken care to express γ_0 in terms of the total equivolumetric mean radius R instead of the equatorial radius a (the latter being more commonly used in the literature) because we assume that R has remained constant throughout time, whereas a obviously has not. On this basis, we have drawn in Fig. 3 the normal gravity field as a function of latitude for a number of selected epochs.

Thus, according to our model results, the normal gravity has increased at the equator from 9.7570 to 9.7804 m/s^2 over the last 2.45×10^9 years, i.e. an increase of 2340 mgals. Over the same time span, normal gravity at the poles changed from 9.8391 to 9.8322 m/s^2, amounting to a decrease of only 690 mgals. This demonstrates the fact that the changing centrifugal acceleration affects more the gravity changes than the changing distances from the Earth's centre. This fact is immediately apparent from relation (9) if we notice that the coefficients of m embody the direct contribution of the centrifugal acceleration to gravity, whereas the coefficients of f represent an indirect contribution via a change of shape.

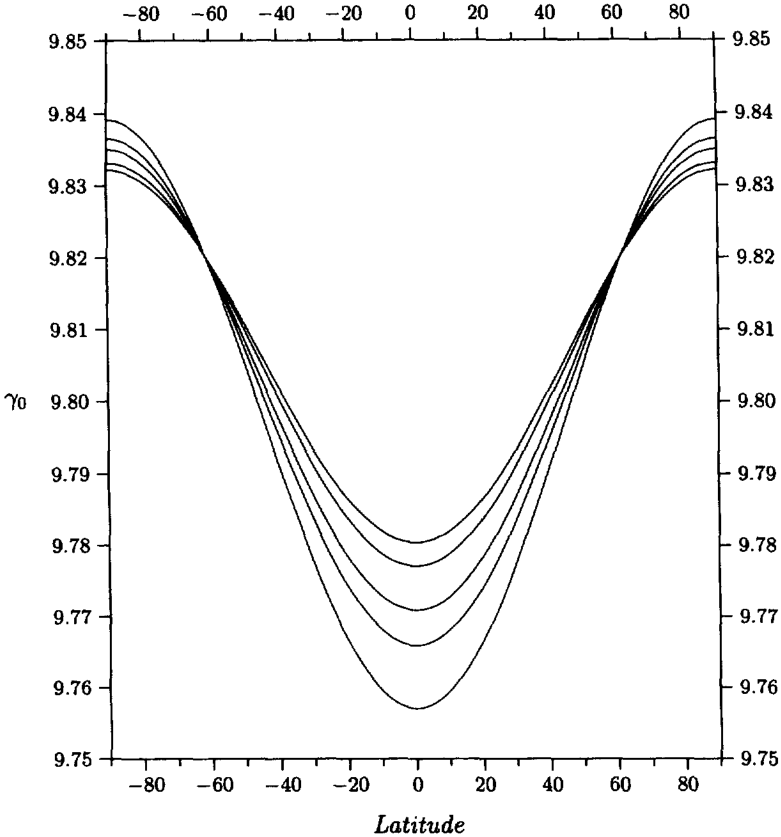

Fig. 3. Inferred normal gravity γ_0 as a function of latitude φ on the Clairaut spheroid throughout geological time. Latitude is in degrees (°), normal gravity in m/s^2. The five curves, ordered with decreasing values of normal gravity at the equator (or increasing values at the poles), correspond to the present epoch (τ=0.0), to the Upper Permian (τ=0.19 *Ga*), to the Lower Ordovician (τ=0.49 *Ga*), to the late Algonkian (τ=0.90 *Ga*) and to the early Algonkian (τ=2.45 *Ga*), respectively.

Moreover, we notice that since the late Permian, 190 million years ago, the normal gravity has decreased at the poles by about 90 mgals. At latitudes 50° North and South, it has increased concomitantly by about 80 mgals. The increase at the equator was about 330 mgals, which leads to an average increase rate of about 2 ngals yr^{-1}. This very small value is close to the sensitivity limit of modern superconducting gravimeters and may just be within our observational possibilities. Although some time will pass before this tiny signal may actually get identified, because of instrumental drift problems and screening by non-secular time changes of the Earth's gravity field which are greater by several orders of magnitude, we believe that our result opens an interesting field of investigation and represents an ambitious challenge for geodesists.

STRESSES DUE TO THE CHANGES IN TIME OF THE GEOMETRICAL FLATTENING

Using the values of the rotation speed at different geological epochs, the broad changes of the Earth's shape and related parameters can be inferred with some confidence from the theory of hydrostatic figures, supposing that the gross mass distribution inside the Earth has more or less remained the same. This goal has been achieved in the previous section. Denis and Varga (1990) have argued that the secular changes of the geometrical flattening build up an incremental stress field in the brittle outer layers of the Earth, namely in the effective elastic lithosphere. It is assumed that the latter is too thin to change the actual flattening appreciably with respect to its hydrostatic value. Turcotte (1979) states that there is extensive observational evidence from flexure studies of the oceanic lithosphere that the tectonic plates, or at least the upper part of these plates where the temperature is below 400–600°C, behave elastically over long geological time scales and has introduced the concept of membrane stress tectonics (Turcotte, 1974; Freeth, 1979).

Laboratory studies by Goetze and Evans (1979) indicate that common rocks cannot sustain shear stresses exceeding 0.1 to 1 *GPa*, the exact value of the yield stress depending strongly on temperature. Once the applied shear stress becomes larger than this yield stress, the materials cease to behave elastically. Therefore, Goetze and Evans suggest a rheological model for the lithosphere which agrees with the model on which the conclusions reached here are based. This model comprises a very thin cool upper layer which is elastic-brittle, and a thick hot lower layer which is ductile and over long time spans behaves like a fluid. Bodine *et al.* (1981); McNutt and Menard (1982); Watts and Ribe (1984); Diament (1987), among many others, have used *in situ* observations—including geoid heights and bathymetric profiles around seamounts and volcanic island chains—to study flexure of the oceanic lithosphere under load. Their results support the idea that the long-term rheological behaviour of the lithosphere can be conceived as a thin, continuous and homogeneous, elastic-brittle layer on top and a visco-elastic material at the bottom. The effective elastic thickness decreases with the age of the load, but tends asymptotically to a finite limit comprised between 10 and 20 km.

Amalvict and Legros (1986, 1991 and 1993) derived formulae for incremental lithospheric stress in simple gravito-elastic models. Their results extend earlier results (Vening Meinesz, 1947; Melosh, 1977). In terms of an effective shear modulus μ_{eff} of the elastic outer shell, the incremental meridional stress $\sigma_{\varphi\varphi}$ is given as a function of latitude φ by

$$\sigma_{\varphi\varphi}(\varphi) = -\frac{\mu_{eff}\Delta f}{11}[5 - 3\cos 2\varphi - 4\varepsilon(3 + 7\cos 2\varphi)] \tag{10}$$

Here, Δf denotes the change of flattening with respect to a reference value f_0 at a given epoch

τ_0, and ϵ is the thickness of the effective lithosphere divided by the Earth's total radius. The incremental azimuthal stress $\sigma_{\lambda\lambda}$ (λ denoting longitude) as a function of latitude is

$$\sigma_{\lambda\lambda}(\varphi)=+\frac{\mu_{eff}\Delta f}{11}[1+9\cos 2\varphi-4\varepsilon(5+\cos 2\varphi)] \tag{11}$$

The resulting incremental stress difference $\Delta\sigma$ as a function of latitude is

$$\Delta\sigma(\varphi)=\sigma_{\varphi\varphi}(\varphi)-\sigma_{\lambda\lambda}(\varphi)=-\frac{\mu_{eff}\Delta f}{11}(6-32\varepsilon)(1+\cos 2\varphi) \tag{12}$$

We note that the stress pattern is symmetrical with respect to the equator, in agreement with Pierre Curie's Symmetry Law (Curie, 1894; Denis, 1989). Figure 4 shows the stress patterns for ϵ=0.0, 0.01 and 0.1, assuming a flattening decrement $\Delta f=-0.001$ and an effective shear modulus $\mu_{eff}=10^{11}$ *Pa*. The latter is a typical value for surface rocks. From the information gathered in sections 2 and 3, we find that the flattening exceeded the present flattening by 0.001 at some moment in the late Infra-Cambrian or early Cambrian, roughly 580 million years ago (epoch names according to Foucault and Raoult, 1984). The length of the day was then about 21.1 hours. Because $\epsilon<0.003$ according to the observations reported above, it is in general a good approximation to neglect the finite thickness of the Earth's effective lithosphere.

Some 200 million years ago, in the late Triassic when the supercontinent Pangaea broke into pieces, the length of the day was about 23 hours, according to Eqn. (4). From Table 1 we see that the surface flattening at this remote epoch was close to 0.0036 whereas the present-day flattening is close to 0.0033. Thus, the flattening decrease since the late Triassic is about 0.0003, i.e. $\Delta f\approx-3\times10^{-4}$. Because Δf intervenes linearly in the Eqns. (10)–(12), the stress increments since the breakup of Pangaea may be obtained by dividing the values indicated in Fig. 4 roughly by 3. We notice that if Δf is negative, the associated incremental stress difference $\Delta\sigma$ is positive, and *vice-versa*. For $\Delta f<0$, $\Delta\sigma$ reaches a maximum value at the equator ($\varphi=0°$) and vanishes at the poles ($\varphi=\pm 90°$). Now, if we remember that a positive stress is tensional and a negative stress is compressional, we see that the meridional (or latitudinal) stress $\sigma_{\varphi\varphi}$ is tensional at all latitudes *provided the effective lithosphere is not too thick*, whereas the azimuthal (or longitudinal) stress $\sigma_{\lambda\lambda}$ is compressional up to some critical latitude φ_{crit}, poleward of which it becomes tensional as well. For an infinitely thin lithosphere, for which $\epsilon=0$, $\varphi_{crit}=\pm48.19°$. If we consider increasingly thicker lithospheric models, the parallel on which $\sigma_{\lambda\lambda}=0$ migrates from mid-latitudes to lower latitudes.

Let us now assume that σ_{rr}, $\sigma_{\varphi\varphi}$, $\sigma_{\lambda\lambda}$ are principal stresses. Since the Earth's effective lithosphere is very thin, we may assume that the radial stress increment σ_{rr} vanishes. Under these circumstances we may explain the stress pattern of Fig. 4 in terms of Anderson's theory of near-surface faulting (Anderson, 1951). In particular, the stress pattern shown in Fig. 4 establishes the existence of two major tectonic provinces, namely a strike-slip fault province ($\sigma_{\varphi\varphi}>\sigma_{rr}>\sigma_{\lambda\lambda}$) which extends from the equator to mid-latitudes, and a normal fault province ($\sigma_{\varphi\varphi}>\sigma_{\lambda\lambda}>\sigma_{rr}$) which extends from mid-latitudes to the poles.

To reach this conclusion, we have neglected the global shrinkage in time associated with the tidal despinning. This shrinkage is a consequence of the changing radial term of the centrifugal potential. In fact, in agreement with the Clairaut-Laplace-Lyapunov (CLL) theory of hydrostatic equilibrium figures, we assumed until now that the Earth behaves like a *permanent rotator*. This amounts in particular to saying that rotational deformation is incompressive (Tassoul, 1978). Thus, only the quadrupole term of the centrifugal potential needs to be taken into account and no change in volume, local or global, occurs. It does actually not imply that the Earth material be incompressible.

Stoneley (1924) considered shrinkage of the Earth's volume caused by tidal despinning in an attempt to explain orogenesis. He computed, with the then admitted elastic parameters describing the mechanical structure of the Earth's interior, that a change of l.o.d. from 12 to 24 hours would diminish the Earth's mean radius by about 1.6 km. For the same change in l.o.d., the CLL theory leads to a decrease of the equatorial radius of about 16 km, and a concomitant increase of the polar radius of approximately 32 km (with no change of the mean radius). Hinderer and Legros (1990) find a contraction of 470 m over the last 400 million years. Their

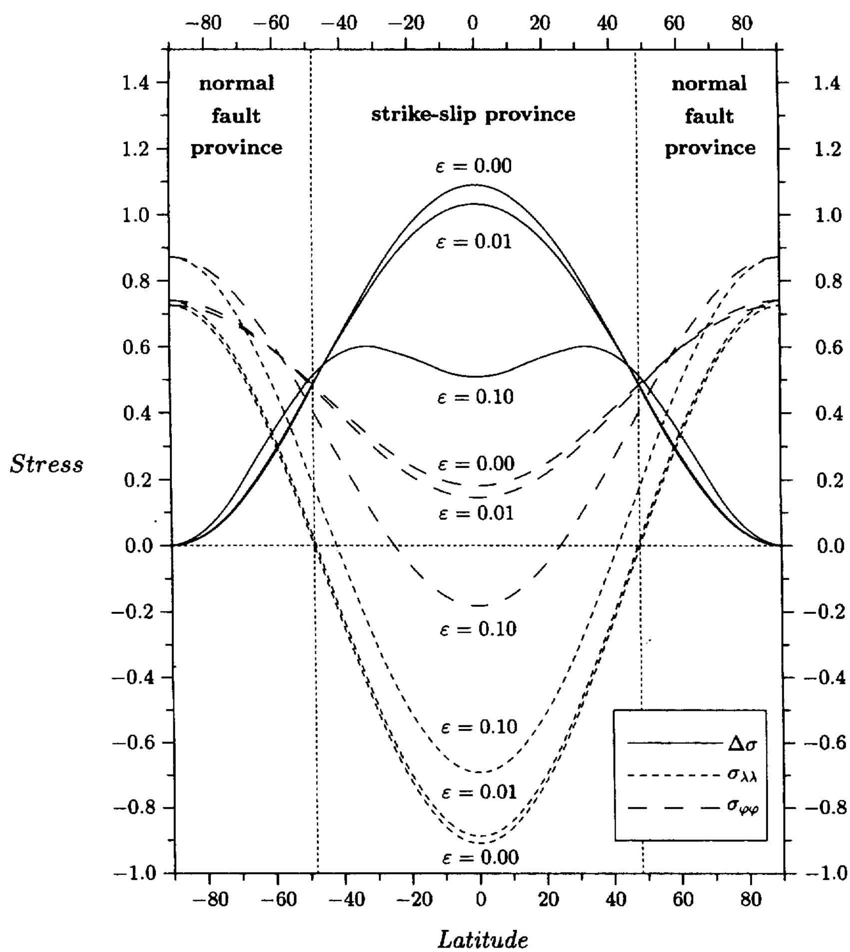

Fig. 4. Incremental shear stresses (expressed in kilobar) in the effective lithosphere caused by a decrease in the flattening of the Earth's outer surface by 0.001. The effective shear modulus of the lithosphere is assumed to be 1 megabar. Broken lines with long dashes correspond to meridional stress increment $\sigma_{\varphi\varphi}$, broken lines with short dashes to azimuthal stress increment $\sigma_{\lambda\lambda}$, and solid lines to the stress difference $\Delta\sigma = \sigma_{\varphi\varphi} - \sigma_{\lambda\lambda}$. The curves go by sets of three corresponding respectively to a normalized thickness ϵ of 0.00, 0.01, and 0.10. Notice that the behaviour of $\Delta\sigma$ changes appreciably when the lithosphere becomes thicker.

result is compatible with Stoneley's value, which was obtained for a much longer time span. On the other hand, computations based on CLL theory yield a decrease of the equatorial radius of 1.5 km and an increase of the polar radius of 3.0 km for the last 400 million years. We may therefore argue with some confidence that stress increments caused by tidal shrinkage are an order of magnitude smaller than the stress increments caused by the change of shape.

It seems worthwhile to pursue the topic of the Earth's shrinkage slightly further. Indeed, if we use experimental values of the compressibility of common rocks, and if we suppose that the gravity increased over the Earth's history as a consequence of tidal shrinkage (mainly because of the changing centrifugal force) by 3%, we may infer a relative change of the Earth's mean radius, say $|\Delta R|/R$, comprised between 0.1% and 1%. Our estimate is based on a paper by Birch (1968) which shows that a doubling of the present value of gravity would lead to a decrease of the Earth's radius by 400 km at most. Hence, an overall increase of gravity by 0.3 m/s^2, resulting from an increase of l.o.d. from 8.4 to 24 hours, would lead to $\Delta R \approx -12$ km, compared to $\Delta a \approx -50$ km and $\Delta c \approx +100$ km. $|\Delta R| = 12$ km is obviously an extreme upper limit for the change of the Earth's total radius caused by tidal friction.

Thus, the results obtained for the incremental stress pattern in the effective lithosphere, shown in Fig. 4, should not change dramatically if we account for the Earth's finite compressibility. Shrinkage caused by tidal spindown may possibly give rise to a narrow thrust province in the equatorial region. However, the Earth's volume might not have decreased at all in any significant way, or it might even have slightly increased over geological time. Indeed, thermal dilatation caused by radioactive or tidal heating or expansion caused by a possible decrease of the gravitational constant are effects that oppose contraction due to despinning. If any global expansion has occurred, the strike-slip province at low latitudes may extend to somewhat higher latitudes.

TECTONIC IMPLICATIONS OF TIDAL DESPINNING

The results obtained in the previous section invite us to speculate on the implications of tidal friction for the tectonic evolution of the Earth, and suggest a plausible scenario (Denis and Varga, 1990; Denis, 1993). More specifically, they allow us to discuss in general terms the dynamics of the uppermost layers in the very remote past. Unfortunately, geological data concerning the Earth's early history are rather sparse and unreliable, and therefore give rise to much debate. Nevertheless, there seems to be a large consensus that the Earth's crust was mainly granitic more than 3.5 billion years ago. After this epoch, roughly between 3.5 and 2.7 billion years in the past, larger quantities of sedimentary and volcanic rocks were added to the granitic crust. Then, probably between 2.7 and 2.0 billion years ago, stable continental blocks were formed, the present continental cratons. During this time span of 700 million years, a gradual transition from great tectonic mobility and crustal thickening to the present style of global plate tectonics seems to have occurred (Lambeck, 1980).

The continental drift which takes place today is essentially governed by convective motions deep inside the Earth's mantle. However, because the asthenosphere uncouples to some extent motions in the lower part of the mantle from deformation of the lithosphere, it is not obvious that mantle convection can account for the original breakup of the lithosphere into a number of plates, as suggested by the work of Ichiye (1971). On the other hand, if lithospheric breakup is caused by tidal despinning, it is easy to understand why no compelling evidence of earthlike plate tectonics has been found on Mars and Venus, two telluric planets with no massive companion. Actually, for Venus the issue is not definitively settled yet: recent evidence seems to point to some kind of active tectonics, which may or may not be similar to plate tectonics on Earth (Phillips and Hansen, 1994). In this respect, it is noteworthy that Liu (1974) suggested the breakup of the lithosphere to be due to polar wandering, by which the lithosphere experiences a change of curvature similar to the one caused by tidal despinning.

The shear strength S found in laboratory measurements under confining pressures and temperatures relevant to the effective lithosphere is in the range 0.1–0.3 kbar for granite, and 0.3–0.5 kbar for basalt. The angle of internal friction δ is comprised between 56° and 58° in granite, between 48° and 50° in basalt (Jaeger, 1978). Shear failure in brittle materials such as granite and basalt occurs, according to the Coulomb-Navier-Mohr theory, if

$$|\Delta\sigma-(\tan^2\gamma-1)\sigma_{\lambda\lambda}|\geq|2S\tan\gamma| \tag{13}$$

with $\gamma=45°+1/2\delta$. The failure takes place in a vertical plane at an angle γ to the longitudinal stress.

For granite, the tensile strength is close to 0.4 kbar. The compressive strength is much larger, about 1.4 kbar. Thus, shear fracture occurs in granite if

$$|\Delta\sigma-10.4\sigma_{\lambda\lambda}|\geq 1.35\text{ kbar} \tag{14}$$

assuming that $S=0.2$ kbar and $\delta=57°$. It happens in basalt if

$$|\Delta\sigma-6.16\sigma_{\lambda\lambda}|\geq 2.14\text{ kbar} \tag{15}$$

assuming that $S=0.4$ kbar and $\delta=49°$. Typical shear moduli for granite and basalt are 234 and 307 kbar, respectively (Clark, 1966).

These criteria, together with the values of Δf, $\sigma_{\lambda\lambda}$ and $\Delta\sigma$ found in sections 3 and 4, give strong evidence that an original crust mainly made up of granite must have undergone severe strike-slip faulting at low latitudes roughly 3.5 billion years ago. If the original crust had been composed mainly of basalt, strike-slip faulting would have occurred 900 million years later, about 2.6 billion years ago. Normal faulting at higher and high latitudes must have happened quite earlier, maybe only 200 to 500 million years after the original crust existed.

The fact that normal faulting was active so early in the Earth's history may have caused a lasting asymmetry between the northern and southern hemispheres. Indeed, at that early epoch the core may still have been small enough to permit some kind of dipolar whole–mantle convection to take place (Runcorn, 1965; Fairbridge, 1972; Denis, 1985). Thus, hot ascending magma should have extruded at high latitudes in one hemisphere, but not at high latitudes in the other hemisphere, where cool material was descending towards the centre, nor at low latitudes on any side of the equator (Denis, 1993).

Reverse faulting—particularly thrust faulting—is not typical of a tidal spindown regime. It is presumably associated mainly with collisions of moving plates. This idea suggests that folded orogenic belts showing large thrust faults can only have occurred much later, say less than 2.5 billion years ago. Volcanic mountain chains, however, should have occurred on Earth about 200–500 million years after the original granitic crust had been formed. It is not possible yet to decide whether faulting due to stress buildup in the effective lithosphere by tidal despinning was of an episodic nature or was rather continuous in time, after the original lithosphere had been fractured into several pieces. Because tidal despinning is a very slow but continuous process, and because rupture occurs suddenly for a certain stress level, there is a definite possibility that major strike-slip fault systems have occurred episodically before an actual plate tectonic regime was established. On the other hand, the creation of major rift systems should be a more or less stationary process.

Drifting of plates is, at least in principle, not related directly with tidal friction. Nevertheless, this phenomenon should give rise to stress buildup in plates in much the same way as tidal despinning does. Indeed, a plate possessing a southward or northward drift component, such as the African plate presently has, experiences a change in curvature which builds up intraplate membrane stress and may possibly lead to intraplate earthquakes. Membrane stress tectonics of the African plate has been considered by Freeth (1979). Much of the theoretical formulation of this field is due to Turcotte (1974). Since the flattening of the Earth was greater in the past, intraplate tectonics was presumably more efficient in the Cryptozoic than in more recent times.

The global evolutionary scenario explained here seems plausible enough, but needs some observational support. Obviously, the existence of the fundamental ingredients—tidal spindown and a concomitant decrease of the Earth's flattening—cannot be questioned. The existence of an effective lithosphere capable of sustaining an elastic-brittle behaviour over hundreds or thousands of million years is not readily accepted by many geophysicists who believe that over geological times *all* shear stresses in solid rock are getting relaxed by creep phenomena. However, this belief is not based on experiment or theory, but rather on general ideas about creep processes in the mantle extrapolated to the upper lithosphere. The mere observation that the very ancient surface features have been preserved for hundreds of millions of years on Earth, and even more distinctly on the Moon, Mercury, Mars and other planetary bodies where erosion is less intensive, points to the fact that the cool outer lithospheres of earthlike planets are indeed capable of sustaining shear stress over geological intervals of time. We simply must admit that we still ignore the exact rheologic law which rules the behaviour of the terrestrial lithosphere on time scales ranging from less than 100 to more than 1000 million years.

Nevertheless, as already stated in section 4, flexure studies of the oceanic lithosphere show strong evidence that over time spans and stress levels relevant to our investigation, the transition from an essentially elastic to an essentially ductile regime occurs in the Earth at an isotherm of ca. 400°C (Turcotte and Schubert, 1982). If such is the case, the thickness of the effective lithosphere must lie between 10 and 20 km. These bounds are smaller by a factor of 5 to 10 than the thickness of the lithosphere determined by seismological studies. Such a thin shell cannot exert an appreciable bending stress, and thus cannot support a significant non-hydrostatic rotational bulge at the equator. Consequently, the hydrostatic figure theory ought to be an excellent approximation for inferring the overall shape of the effective lithosphere at any moment. Knowing this shape, we may use thin elastic shell theory (Kraus, 1967) to compute incremental strains and stresses produced by a change of curvature. These considerations justify the procedure followed and the results obtained in sections 3 and 4.

It is obvious that once a plate-tectonic process had started, the most striking tectonic features became dominated by the dynamics of this process. Tectonic effects implied by tidal spindown are likely to be masked and smeared out almost completely by the dominating effects produced by the motions of the lithospheric plates. Nevertheless, there seems to remain a slight chance to observe a tiny signature of tidal despinning in present-day seismicity statistics. Indeed, there is *a priori* no reason to suspect some particular latitude dependence in the occurrence of shallow-focus earthquakes caused by plate interactions. On the other hand, stress buildup caused by tidal despinning has a striking latitude pattern (Fig. 4), which might be reflected to some extent in earthquake statistics.

To investigate this point, we plot in Fig. 5 the function occurring on the l.h.s. of the Coulomb-Navier-Mohr failure criterion (13), namely $|\Delta\sigma-(\tan^2\gamma-1)\sigma_{\lambda\lambda}|$, as a function of latitude. For γ we use 71° corresponding to the plausible value $\delta=52°$. The plot is drawn for the values of $\Delta\sigma$ and $\sigma_{\lambda\lambda}$ given in Fig. 4, and thus correspond to $\mu_{eff}=1$ megabar, $\Delta f=-0.001$ (and $\epsilon=0.0$).

Although the maximum value reached at the equator is 4 to 6 times above the failure stress, the actual values of this curve are not important for the purpose we have in mind here; only the general pattern is important. This pattern is again quite striking, with an absolute maximum reached at the equator, a relative maximum reached at the poles, and an absolute minimum reached for latitudes close to 50° North and South. No shear failure occurs at this minimum, at any time. We tried to find some similar tectonic signature in catalogues of shallow-focus earthquakes.

As a first step we studied the seismic energy released during the strongest (magnitude $M \geq 7.9$) earthquakes, using as observational basis Richter's (1958) catalogue.

The annual frequency of such big seismic events is low, about 1 to 2 events per year, but they contribute to almost 60% of the annually released seismic energy (Kasahara, 1981) and have little or no observational bias. For binning purposes we use latitude zones of equal width (15°). Therefore, for making meaningful energy considerations, we multiplied the energy released in any individual latitude zone by the ratio of the area of the equatorial zone [0°···15°] to the area

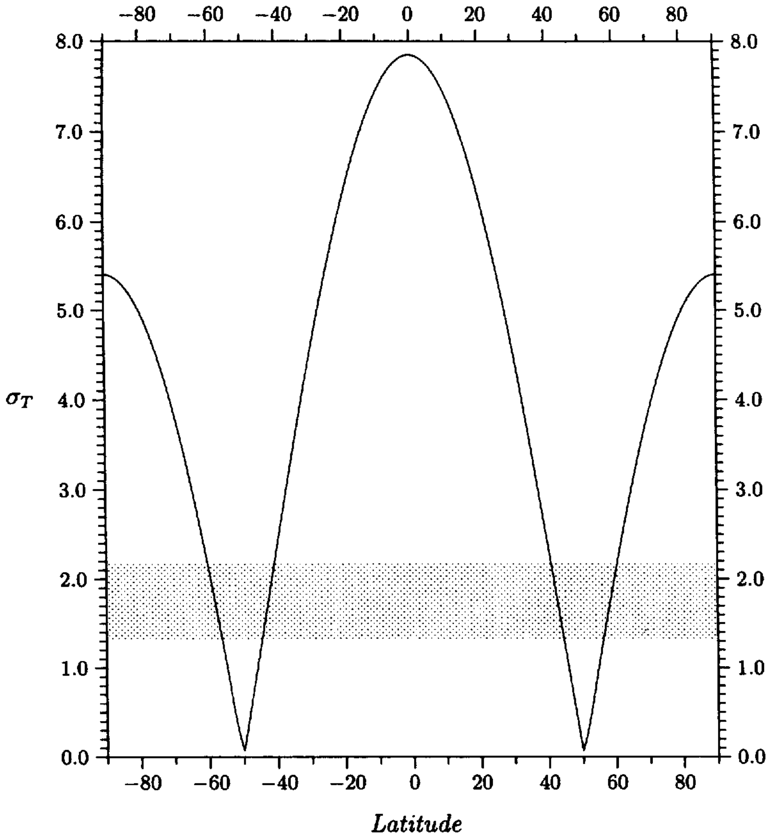

Fig. 5. Characteristic stress function for the Coulomb-Navier-Mohr shear failure criterion $\sigma_T = |\Delta\sigma + (1 - \tan^2\gamma)\sigma_{\lambda\lambda}|$ as a function of latitude. The plot is drawn for $\mu_{eff} = 1 \times 10^{11}$ *Pa* $\Delta f = -0.001$, $\epsilon = 0.0$, and $\gamma = 71°$. Shear failure in the lithosphere occurs for stress levels situated somewhere in the shaded area.

Table 3. Latitude distribution of shallow-focus seismic activity for earthquakes with magnitudes M≥7.9, according to the Richter (1958) earthquake catalogue. Owing to the relatively small number ($N=194$) of events, latitude zones symmetrical with respect to the equator were merged, and binning was made for latitude zones with a 15° width

Latitude zone (°)	Number of events	Normalized energy release ΣE^*_j (10^{18} J)	Normalized length of tectonic lineaments
[0···15[	54	1.99	1.00
[15···30[	47	1.74	0.80
[30···45[	32	1.61	0.61
[45···60[	42	2.47	0.65
[60···75[	19	1.55	0.40
[75···90]	0	0.00	0.00

of the considered latitude zone. The normalized values ΣE^*_j thus obtained are provided in the third column of Table 3. The seismic energy E_j (expressed in Joule) released by an individual earthquake of magnitude M was computed by means of the Gutenberg-Richter formula in the slightly amended form given by Kasahara (1981), *i.e.*

$$E_j=4.8+1.5M \tag{16}$$

Because of the restricted number of data, we actually assumed symmetry with respect to the equator and grouped any given northern latitude zone with the corresponding southern one. Moreover, it should be clear that a study based on the elastic energy released during particular earthquakes suffers from an uncertainty in magnitude which generally amounts to 0.2–0.3, but may be greater (Kanamori, 1977). The latter may influence in an essential way the calculated results. Therefore some caution is necessary when an interpretation is based on the distribution of seismic energy. The fourth column of Table 3 contains the relative lengths of active linear structures (tectonic lineaments) related to global tectonics in the same latitude zones. They are normalized for the equatorial region and were obtained from the inspection of the world map of the planetary tectonic structures (Beatty and Chaikin, 1991).

From the information gathered in Table 3 and based on the Richter (1958) catalogue we draw the following conclusions:

—no shallow events with $M \geq 7.9$ seem to occur at latitudes higher than 65°;
—the distribution of the number of large seismic shallow-focus events is quite similar to the distribution of the tectonic lineaments;
—there is no correlation between the distribution along the latitude of seismic energy release and the occurrence of the strongest earthquakes.

However, an important hint about a possible signature of seismic activity caused by tidal despinning would be provided if the seismicity statistics would indicate a pattern symmetrical with respect to the equator showing some correlation with the pattern illustrated in Fig. 5. For this purpose, we need to consider more events than those of Richter's catalogue. Therefore, as a second step, we used 698 events listed in the seismic moment catalogue of large shallow earthquakes from 1900 to 1982 (Pacheco and Sykes, 1992). We binned these data again into latitude zones of 15°, this time considering the northern and southern hemispheres separately.

Unlike seismicity caused by plate tectonics, for which an activity index is best defined along arcs, "background seismicity" caused by a change in the Earth's curvature should be proportional to the area considered. For this reason we applied a correction factor to the observed number of events and to the seismic energy released in each latitude zone. The gross data and the corrected values are given in Table 4. We should like to mention that the latitude zones with a width of 15° were not selected in a totally subjective manner, but were imposed on us by a consideration of empirical criteria for selecting an optimum of categories for binning N data. These criteria (*cfr.* Schönwiese, 1985 pp. 15–16) indicate that for N=698 events a number of classes comprised between 10 (according to the Sturges-Strauch criterion) and 14 (according to the Panofsky-Brier criterion) should be chosen. Twelve classes thus appear to us as a fair choice.

Histograms for the corrected numbers of events and for the corrected energies are shown in Fig. 6 and Fig. 7, respectively. Although these histograms are different if we look at them in detail, both exhibit a multimodal pattern which we wish to correlate with the typical multimodal aspect of the function plotted in Fig. 5. In particular, we notice a rough symmetry with respect to the equator for the occurrence of earthquakes.

Consider the N pairs of variables (X_1,Y_1), (X_2,Y_2), $\cdots$ (X_N,Y_N), where X stands for instance for $\bar{\sigma}_T$ and Y for N^*. Denote the mean of the X-values by $\bar{X}$, the mean of the Y-values by $\bar{Y}$ and the corresponding standard deviations respectively by s_X and s_Y. Here, $N=12$, $\bar{X}=4.268$, $\bar{Y}=69.667$, $s_X=2.391$ and $s_Y=63.258$. Defining the sample correlation coefficient between the two data sets $\{X_i\}$ and $\{Y_i\}$ as (Davis, 1986)

$$r=\frac{\sum_{i=1}^{N}(X_i-\bar{X})(Y_i-\bar{Y})}{(N-1)s_X s_Y} \tag{17}$$

Table 4. Latitude distribution of shallow-focus seismic activity for earthquakes compiled from the Pacheco and Sykes (1992) seismic moment catalogue. The first column ($\Delta\varphi$) shows the 12 latitude zones. The other columns give characteristic values for each latitude zone: column 2 (CF) provides the correction factor, column 3 (N) the number of events, column 4 (N^*) the corrected number of events, columns 5 and 6 (ΣE respectively ΣE^*) the total energy and the corrected energy in 10^{18} J, column 7 ($(\bar{\sigma}_T)$) values of the stress function for the shear failure criterion (defined as in the caption of Fig. 5) evaluated at the mid-point of each latitude zone

$\Delta\varphi$	CF	N	N^*	ΣE	ΣE^*	$\bar{\sigma}_T$
[−90, −75[	7.60	0	0	0.000	0.000	5.18
[−75, −60[	2.59	2	5	0.004	0.010	3.47
[−60, −45[	1.63	20	33	0.285	0.464	0.49
[−45, −30[	1.25	33	41	1.088	1.360	2.94
[−30, −15[	1.07	64	69	0.842	0.904	5.91
[−15, 0[	1.00	146	146	1.386	1.386	7.62
[0, 15[	1.00	85	85	0.854	0.854	7.62
[15, 30[	1.07	106	114	1.494	1.603	5.91
[30, 45[	1.25	154	192	2.307	2.883	2.94
[45, 60[	1.63	76	125	1.362	2.218	0.49
[60, 75[	2.59	10	26	0.278	0.721	3.47
[75, 90]	7.60	0	0	0.000	0.000	5.18

it is easy to show that the latter is comprised between the values -1 and $+1$. $|r|=1$ denotes an exact *linear* relationship between the X and Y data, $r=0$ indicates non-existence of such a *linear* relationship (but a non-linear correlation may well exist between the data sets). In the specific case considered here, we find $r=0.001$ and a value for the t-statistic, namely $t=r\sqrt{(N-2)/(1-r^2)}$, of 0.003. The latter is three orders of magnitude smaller than the critical value for rejection of the null hypothesis ($t_{crit}=\pm 1.372$ at a significance level $\alpha=10\%$ for 10 degrees of freedom). In other words, on the basis of the adopted criterion we do not see any statistically significant correlation between earthquake occurrence in a given latitude zone and stresses induced by tidal despinning, despite the fact that Fig. 6 shows a multimodal pattern which is roughly symmetrical with respect to the equator. A similar conclusion may be drawn if for Y we consider ΣE^* instead of N^*. Indeed, we then obtain $\bar{Y}=1.034$, $s_Y=0.821$, $r=-0.092$, and $t=-0.292$. Notice that in this case the computed correlation coefficient would even indicate a negative correlation, which again is statistically not significant. Considering the vastly different time scales on which tidal friction on one hand, plate motions on the other hand, build up tectonic stress, it would indeed be very astonishing to find a significant correlation between seismicity and tidal despinning. This does not mean that tectonics is not influenced on long time scales by tidal friction. From our investigation we are forced to conclude that it definitely is. A

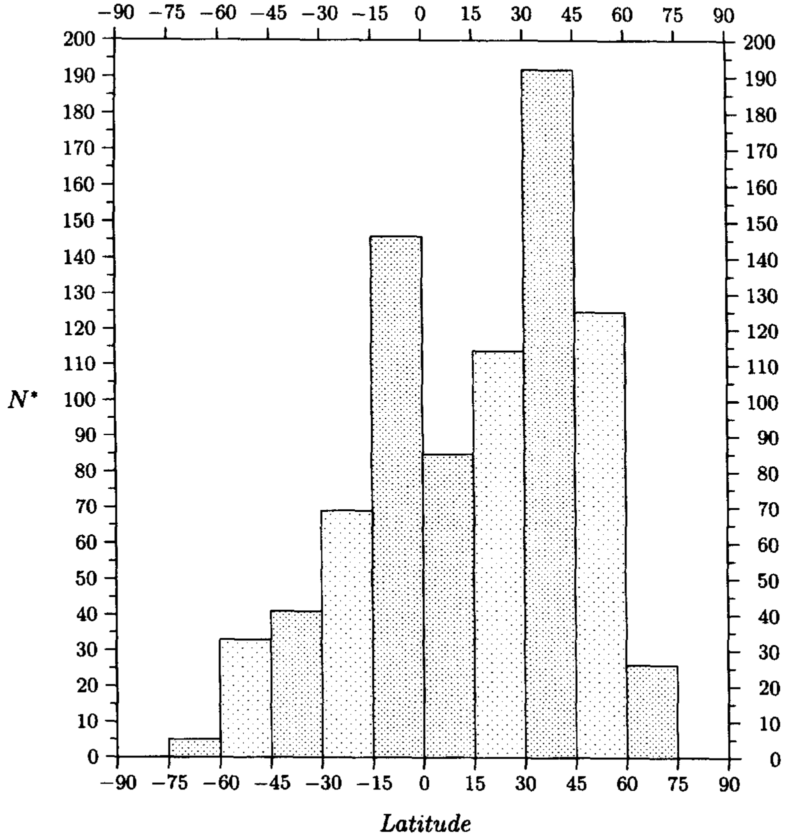

Fig. 6. Area-corrected numbers of earthquakes for 12 latitude zones, each being 15° wide.

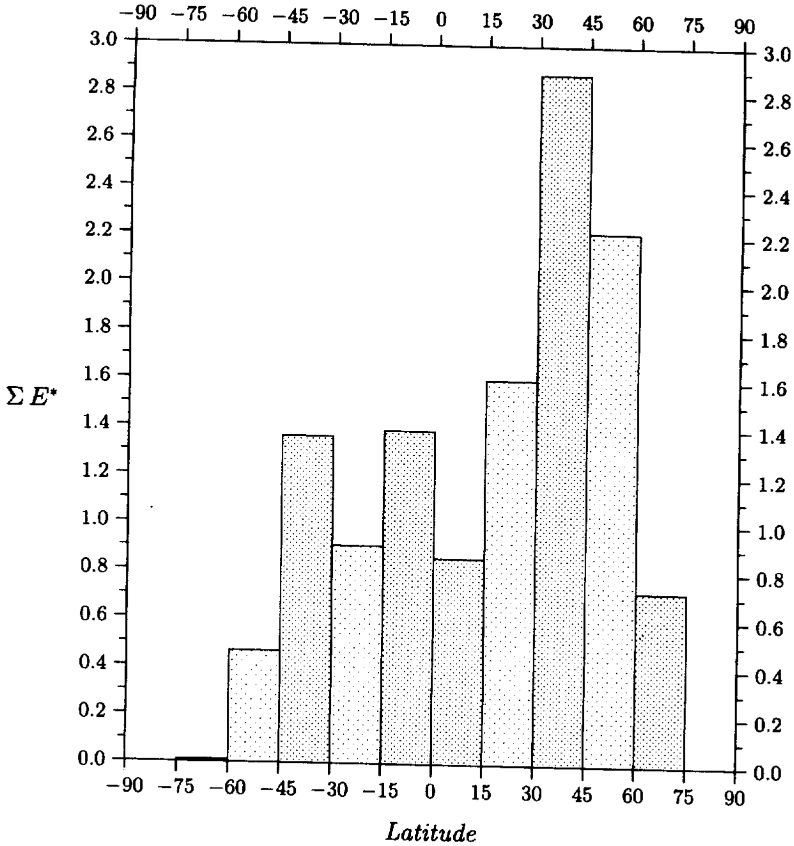

Fig. 7. Area-corrected seismic energy released in 12 latitude zones, each being 15° wide.

slight correlation might actually be found if we would consider a large number of *intraplate* earthquakes, but such a catalogue is not readily available in the literature and ought first to be prepared.

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REFERENCES

Amalvict, M. and Legros, H. (1986) Lithospheric stresses: gravito-elastic model and geometric deformation. *Manuscripta Geodaetica* **11,** 197–206.

Amalvict, M. and Legros, H. (1991) Stresses in the lithosphere induced by geophysical processes of degree two. *Manuscripta Geodaetica* **16,** 333–352.

Amalvict, M. and Legros, H. (1993) Stresses in the lithosphere induced by the Earth's rotation.

In *Physics and Evolution of the Earth's Interior, volume 6: Dynamics of the Earth's Evolution*, ed. R. Teisseyre, pp. 348–389. PWN, Warszawa and Elsevier, Amsterdam.

Anderson, D. L. and Hanks, T. C. (1972) Formation of the Earth's core. *Nature* **237,** 387–388.

Anderson, E. M. (1951) *The Dynamics of Faulting*. Oliver and Boyd, Edinburgh.

Beatty, J. K. and Chaikin, A. (1991) (Eds). *The New Solar System*. Cambridge University Press and Sky Publishing Corporation.

Berry, W. B. N. and Barker, R. M. (1968) Fossil bivalve shells indicate longer month and year in Cretaceous than present. *Nature* **217,** 938–939.

Berry, W. B. N. and Barker, R. M. (1975) Growth increments in fossil and modern bivalves. In *Growth Rhythms and the History of the Earth's Rotation*, eds G. D. Rosenberg and S. K. Runcorn, pp. 9–24. John Wiley, London.

Birch, F. (1968) On the possibility of large changes in the Earth's volume. *Phys. Earth Planet. Int.* **1,** 141–147.

Bodine, J. H., Steckler M. S. and Watts A. B. (1981) Observations of flexure and the rheology of the oceanic lithosphere. *Journal of Geophysical Research* **B86,** 9905–9918.

Bretterbauer, K. (1987) The palaeorotation of the Earth. In *Rotating Earth, Sopron Winter Seminar in Geodynamics 1987*, pp. 1–26. Geodetic and Geophysical Research Institute, Sopron.

Brosche, P. (1987) Tidal friction and the Earth's rotation. In *Proc. Int. Symp. Figure and Dynamics of the Earth, Moon, and Planets*, ed. P. Holota, pp. 491–496. Czechoslovak Academy of Sciences, Prague.

Christodoulis, D. C., Smith, D. E., Williamson, R. G. and Klosko, S. M. (1988) Observed tidal braking in the Earth-Moon-Sun system. *Journal of Geophysical Research* **B93,** 6216–6236.

Clark, S. P. jr. (1966) (Editor), Handbook of Physical Constants (Revised Edition), *Geol. Soc. Am.*, Memoir 97.

Curie, P. (1894) Sur la symétrie dans les phénomènes physiques. Symétrie d'un champ électrique et d'un champ magnétique et appliquée (3[e] série). *Journal de Physique théorique et appliquée* **3,** 393–415.

Darwin, G. H., Bewegung der Hydrosphäre, *Encykl. mathem. Wiss.*, **VI**, 1908, pp. 3–83. (English translation in: G. H. Darwin, Scientific Papers, **IV**, 1911, pp. 185–263. Cambridge University Press).

Davis, J. C. (1986) *Statistics and Data Analysis in Geology* (2nd Edn). Wiley, New York.

Denis, C. (1985) Core formation and rotation of the Earth in the geological past. *C. R. Journées Luxbg. Géodyn.* **60,** 12–17.

Denis, C. (1986) On the change of kinetical parameters of the Earth during geological times. *Geophys. J. R. astr. Soc.* **87,** 559–568.

Denis, C. (1989) The hydrostatic figure of the Earth. In *Physics and Evolution of the Earth's Interior, volume 4: Gravity and Low-Frequency Geodynamics*, ed. R. Teisseyre, pp. 111–186. PWN, Warszawa and Elsevier, Amsterdam.

Denis, C. (1993) Global deformations and evolution of the Earth. *Acta Geod. Geoph. Mont. Hung.* **28,** 15–131.

Denis, C., Amalvict, M., Rogister, Y. and Tomecka-Suchoń, S. (1996a) Methods for computing the internal flattening, with applications to the Earth's structure and geodynamics, *Geophys. J. Intl.*, (submitted).

Denis, C., Rogister, Y., Amalvict, M., Delire, C., Denis A. İbrahim and Munhoven, G. (1996b) Hydrostatic flattening, core structure, and translational mode of the inner core, *Phys. Earth Planet. Int.* (in press).

Denis, C. and Varga, P. (1990) Tectonic consequences of the Earth's variable rotation on geological time scales. In *Earth's Rotation from Eons to Days*, eds. P. Brosche and J. Sündermann, pp. 146–162. Springer-Verlag, Berlin.

Deubner, F. L. (1990) Discussion on Late Precambrian tidal rhythmites in South Australia and the history of the Earth's rotation. *Journal of the Geological Society London* **147,** 1083–1084.

Diament, M. (1987) Isostasie, réponses mécanique et thermique de la lithosphère, Thèse de doctorat d'Etat, Université de Paris-Sud, Centre d'Orsay, Paris.

Fairbridge, R. W. (1972) Planetary spin-rate and evolving cores. *Ann. N. Y. Acad. Sci.* **187,** 88–107.

Foucault, A. and Raoult, J.-F. (1984) *Dictionnaire de Géologie* (2[e] édition). Masson, Paris.

Freeth, S. J. (1979) Deformation of the African plate as a consequence of membrane stress domains generated by post-Jurassic drift. *Earth Planet. Sci. Lett.* **45,** 93–104.

Goetze, C. and Evans, B. (1979) Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophys. J. R. astr. Soc.* **59,** 463–478.

Gotlib, V. Y. and Kagan, B. A. (1985) A reconstruction of the tides in the Paleocean: results of a numerical simulation. *Deutsche Hydrograph. Ztschr.* **38,** 190–209.

Heiskanen, W. A. and Vening Meinesz, F. A. (1958) *The Earth and Its Gravity Field.* McGraw-Hill, New York.

Hinderer, J. and Legros, H. (1990) Global consequences of the tidal secular deceleration for the solid Earth and its fluid core. In *Earth's Rotation from Eons to Days*, eds P. Brosche and J. Sündermann, pp. 177–187. Springer-Verlag, Berlin.

Ichiye, T. (1971) Continental breakup by nonstationary mantle convection generated with differential heating of the crust. *Journal of Geophysical Research* **76,** 1139–1153.

Jaeger, J. C. (1978) *Elasticity, Fracture and Flow (With Engineering and Geological Applications).* Chapman and Hall, London.

Johnson, G. A. L. and Nudds, J. R. (1975) Carboniferous coral geochronometers. In *Growth Rhythms and the History of the Earth's Rotation*, eds G. D. Rosenberg and S. K. Runcorn, pp. 9–24. John Wiley, London.

Kanamori, H. (1977) The energy release in great earthquakes. *Journal of Geophysical Research* **82,** 2981–2987.

Kasahara, K. (1981) *Earthquake mechanics.* Cambridge University Press.

Kostelecký, J. (1990) Celestial mechanics of present tidal friction. In *Earth's Rotation from Eons to Days*, eds P. Brosche and J. Sündermann, pp. 58–68. Springer-Verlag, Berlin.

Kraus, H. (1967) *Thin Elastic Shells.* Wiley, New York.

Krohn, J. and Sündermann, J. (1982) Paleotides before the Permian. In *Tidal Friction and the Earth's Rotation II*, eds P. Brosche and J. Sündermann, pp. 190–209. Springer-Verlag, Berlin.

Lambeck, K. (1977) Tidal dissipation in the oceans: astronomical, geophysical and oceanographic consequences. *Phil. Trans. R. Soc. Lond.* **A287,** 545–594.

Lambeck, K. (1978) The Earth's palaeorotation. In *Tidal Friction and the Earth's Rotation*, eds P. Brosche and J. Sündermann, pp. 145–153. Springer-Verlag, Berlin.

Lambeck, K. (1980) *The Earth's Variable Rotation (Geophysical Causes and Consequences).* Cambridge University Press.

Lecoutre, J. P. and Tassi, Ph. (1987) *Statistique non paramétrique et robustesse.* Economica, Paris.

Le Mouël, J.-L., Poirier, J.-P. and Assoumani, M. (1994) Sur le refroidissement du noyau terrestre. *C. R. Acad. Sci. Paris (Série II)* **319,** 1497–1501.

Liu, H. S. (1974) On the breakup of tectonic plates by polar wandering. *Journal of Geophysical Research* **79,** 2568–2572.

Majewski, E. (1995) Thermodynamic model of the core–mantle boundary . *Publ. Inst. Geoph. Pol. Acad. Sci* **A25**(275), 1–61.

Mazullo, S. J. (1971) Length of the year during the Silurian and Devonian periods. *Bull. Geol. Soc. Am.* **82,** 1085–1086.

McNutt, M. and Menard, H. W. (1982) Constraints on yield strength in the oceanic lithosphere derived from observations of flexure. *Geophys. J. R. astr. Soc.* **71,** 363–394.

Melosh, H. J. (1977) Global tectonics of a despun planet. *Icarus* **31,** 221–243.

Mohr, R. E. (1975) Measured periodicities of the Biwabik stromatolites and their geophysical significance. In *Growth Rhythms and the History of the Earth's Rotation*, eds G. D. Rosenberg and S. K. Runcorn, pp. 43–55. John Wiley, London.

Münzenmayer, H. P. (1982) Julius Robert Mayer's ideas on a theory of tidal friction. In *Tidal Friction and the Earth's Rotation* **II**, eds P. Brosche and J. Sündermann, pp. 1–3. Springer-Verlag, Berlin.

Newhall, X. X., Williams, J. G. and Dickey, J. O. (1985, 1986) Earth rotation from lunar laser ranging, *Bur. Int. Heure, Ann. Rept.*, pp. D53-D59.

Oversby, V. M. and Ringwood, A. E. (1971) Time of formation of the Earth's core. *Nature* **234,** 463–465.

Pacheco, J. F. and Sykes, L. R. (1992) Seismic moment catalog of large shallow earthquakes, 1900 to 1989. *Bull. Seism. Soc. Am.* **82,** 1306–1349.

Pannella, G. (1972) Paleontological evidence on the Earth's rotational history since Early Precambrian. *Astrophys. Space Sci.* **16,** 212–237.

Pannella, G. (1975) Paleontological clocks and the history of the Earth's rotation. In *Growth Rhythms and the History of the Earth's Rotation*, eds G. D. Rosenberg and S. K. Runcorn, pp. 9–24. John Wiley, London.

Pannella, G. (1976) Geophysical inferences from stromatolite lamination. In *Stromatolites*, ed. M.R. Walter, pp. 673–685. Elsevier, Amsterdam.

Pariisky, N. N. (1959) On the effect of earth tides on the secular retardation of the Earth's rotation. *Third Int. Symp. Earth Tides*. Trieste, pp. 19-26.

Phillips, R. J. and Hansen V. L. (1994) Tectonic and magmatic evolution of Venus. *Ann. Rev. Earth Planet. Sci.* **22,** 597–654.

Piper, J. D. A. (1990) Implications of some recent sedimentological studies to the history of the Earth-Moon system. In *Earth's Rotation from Eons to Days*, eds P. Brosche and J. Sündermann, pp. 227–233. Springer-Verlag, Berlin.

Richardson, C. A. (1990) Tidal rhythms in the shell secretion of living bivalves. In *Earth's Rotation from Eons to Days*, eds P. Brosche and J. Sündermann, pp. 215–226. Springer-Verlag, Berlin.

Richter, C. F. (1958) *Elementary Seismology*. Freeman, San Francisco.

Runcorn, S. K. (1962) Convection currents in the Earth's mantle. *Nature* **195,** 1248–1249.

Runcorn, S. K. (1964) Changes in the Earth's moment of inertia. *Nature* **204,** 823–825.

Runcorn, S. K. (1965) Changes in the convection pattern in the Earth's mantle and continental drift. *Phil. Trans. R. Soc. Lond.* **A258,** 228–251.

Schönwiese, C. D. (1985) *Praktische Statistik für Meteorologen und Geowissenschaftler*. Borntraeger, Berlin.

Sonett, C. P., Kvale, E. P., Zakharian, A., Chan, M. A. and Demko, T. M. (1996) Late proterozoic and paleozoic tides, retreat of the Moon, and rotation of the Earth. *Science* **273,** 100–104.

Scrutton, C. T. (1970) Evidence for a monthly periodicity in the growth of some corals. In *Palaeogeophysics*, ed. S. K. Runcorn, pp. 11–16. Academic Press, London.

Scrutton, C. T. (1978) Periodic growth features in fossil organisms and the length of the day and month. In *Tidal Friction and the Earth's Rotation*, eds P. Brosche and J. Sündermann, pp. 154–196. Springer-Verlag, Berlin.

Stacey, F. D. (1992) *Physics of the Earth* (3rd Edn). Brookfield Press, Brisbane.

Stephenson, F. R. (1978) Pre-telescopic astronomical observations. In *Tidal Friction and the Earth's Rotation*, eds P. Brosche and J. Sündermann, pp. 5–21. Springer-Verlag, Berlin.

Stephenson, F. R. and Morrison L. V. (1995) Long-term fluctuations in the Earth's rotation, 700 B.C. to A.D. 1990. *Phil. Trans. R. Soc. Lond.* **A351,** 165–202.

Stoneley, R. (1924) The shrinkage of the Earth's crust through diminishing rotation. *Mon. Not.*

R. astr. Soc. Geophys. Suppl. **1,** 149–155.
Stoyko, A. (1970) La variation séculaire de la rotation de la Terre et les problèmes connexes. *Ann. Guébhard (Neuchâtel)* **46,** 293–316.
Sündermann, J. and Brosche, P. (1984) Tidal friction and dynamics of the Earth-Moon system. In *Landolt-Börnstein 2, Geophysik der festen Erde, des Mondes und der Planeten*, eds K. Fuchs and H. Soffel, pp. 125–144. Springer-Verlag, Berlin.
Tassoul, J. L. (1978) *Theory of Rotating Stars*. Princeton University Press.
Turcotte, D. L. (1974) Membrane tectonics. *Geophys. J. R. astr. Soc.* **36,** 33–42.
Turcotte, D. L. (1979) Flexure. *Advanced Geophysics* **21,** 51–86.
Turcotte, D. L. and Schubert, G. (1982) *Geodynamics*. Wiley, New York.
Urey, H. (1952) *The Planets*. Yale University Press.
Varga, P. and Denis, C. (1990) Secular variations of the Earth's moment of inertia and related quantities. In *Earth's Rotation from Eons to Days*, eds P. Brosche and J. Sündermann, pp. 163–176. Springer-Verlag, Berlin.
Varga, P., Denis, C., Varga, T. and Závoti, J. (1992) Statistical modelling of the tidal history of the Earth-Moon system. *C. R. Journées Luxbg. Géodyn.* **73,** 1–7.
Varga, P., Denis, C. and Varga, T. (1993) On temporal variations in the Earth's tidal despinning rate. *Wiss. Z. Techn. Univers., Dresden* **42,** 48–51.
Varga, T., Varga, P. and Denis, C. (1991) Tidal friction and palaeogeodesy. *C. R. Journées Luxbg. Géodyn.* **71,** 7–11.
Vening Meinesz, F. A. (1947) Shear patterns of the Earth's crust. *Trans. Am. Geophys. Un.* **28,** 1–61.
Watts, A. B. and Ribe, N. M. (1984) On geoid heights and flexure of the lithosphere at seamounts. *Journal of Geophysical Research* **B89,** 11152–11170.
Wells, J. W. (1970) Problems of annual and daily growth rings in corals. In *Palaeogeophysics*, ed. S. K. Runcorn, pp. 3–9. Academic Press, London.
Williams, G. E. (1989) Tidal rhythmites: geochronometers for the ancient Earth-Moon system. *Episodes* **12,** 162–171.
Williams, G. E. (1989) Late Precambrian tidal rhythmites in South Australia and the history of the Earth's rotation. *J. Geol. Soc. Lond.* **146,** 97–111.