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Despinning of the earth rotation in the geological past and geomagnetic paleointensities

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Abstract

Length of day (l.o.d.) values deduced from fossils and tidal deposits suggest that the despinning rate was, on the average, about 5 times smaller during the Proterozoic than during the Phanerozoic and, moreover, that between 250 and 100 million years ago, there was a slight non-linear variation super-imposed on the overall linear trend of the Earth's rotation rate. To explain these observational facts, it has recently been argued that formation of the inner structure of the Earth (mass redistribution within the mantle and/or core formation) had not been fully completed before the Proterozoic, and that the decrease of the inertia moment associated with the evolving terrestrial interior compensated to some extent the rotational effects of tidal friction. There is an another plausible explanation to account for the difference of despinning rates during the Proterozoic and Phanerozoic, namely: the distribution of the continents had been significantly different during these epochs and the world ocean had been much shallower in the Proterozoic than in the Phanerozoic. We used published data for the Phanerozoic, Proterozoic and Archean in order to check whether there had been significant long-term changes of geomagnetic intensity. Our results are based on robust statistical analysis; they indicate that during a time interval coinciding roughly with the Mesozoic, the geomagnetic dipole moment underwent a minimum in a quite similar way as the l.o.d. data. For the Proterozoic (2500–570 million years ago) and the late Archean (3000–2500 million years ago), it is very difficult to draw a conclusion concerning the variation in time of the intensity of the geomagnetic field: the data set we used is incomplete and the statistical scatter is larger than the derived mean value. Nevertheless, we tentatively conclude that the values of the average geomagnetic moment were approximately the same in the Phanerozoic and in the Proterozoic + late Archean, and that there is no significant long-term change in the geomagnetic intensity detectable before the Phanerozoic.

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1. Tidal friction and the evolution of the earth-moon system

The history of the Earth–Moon system is of central interest for the understanding of the origin and evolution of the Earth. At the same time, on the basis of tidal observations and theory, it is possible to infer the changes of the Earth's kinetic parameters over much of the geological past (Denis, 1986). Hence it is of foremost importance to gather information on the long-term changes of the Earth–Moon distance. If the lunar orbit would come close to the Earth, it would become significantly inclined with respect to the ecliptic. Such a circumstance is in contradiction with the scenarios of formation of the Moon and the Earth–Moon system which are presently accepted as realistic. The latter place the orbit of the Moon close to the ecliptical plane right from the beginning. Moreover, the knowledge of the tidal and rotational history is valuable for understanding the kinematic and dynamic processes acting within the Earth on a geological time scale. The importance of tidal friction as far as energy exchange is concerned can be illustrated by the following estimates (Maslov, 1991):

(a)	solar energy received by the Earth:	2.1×10^{24} J/a
(b)	loss of energy by heat flow:	1.0×10^{21} J/a
(c)	dissipation of energy by tidal friction:	1.6×10^{19} J/a
(d)	energy released by tectonic activity:	1.3×10^{19} J/a
(e)	energy released by seismic waves:	1.0×10^{18} J/a

Consequently, the dissipation of tidal energy through processes which at present still remain unclear to some extent, contributes significantly to a braking of the Earth's rotation and a decrease of the Earth's flattening, concomitantly with a variation of the associated geokinetic and geodynamic parameters (Denis, 1986, 1993). For instance, if we assume in a first approximation that hydrostatic equilibrium has hold over long time intervals (Denis et al., 1998), we find that the flattening of the Earth's outer surface has decreased by about 70% over the last 2.5×10^9 years. This corresponds to a reduction of the normal gravity at the poles by 690 mGal (0.0069 m/s^2) and an increase at the equator by 2340 mGal (0.0234 m/s^2). Since the late Permian, some 1.9×10^8 years ago, normal gravity has decreased at the poles by about 90 mGal (0.0009 m/s^2) and increased at the equator by 33 mGal (0.00033 m/s^2). Thus the average annual increase of normal gravity at the equator amounts to approximately 2 nGal, i.e. 2×10^{-8} m/s^2 (Varga et al., 1998).

The Earth's despinning over very long time scales can be explained by the tidal–gravitational interaction between Earth and Moon and, to a lesser extent, between Earth and Sun. Such investigations are usually based on the Euler-Liouville equations, which express the conservation of the total angular momentum of the Earth–Moon–Sun system, and are generally kept at a minimum of complexity by assuming that the Moon revolves about the Earth on a circular orbit situated in the plane of the ecliptic, and that the tidal effect of the Sun may be neglected on the time scales involved. Thus, considering the gravitational effect of the Moon alone, conservation of angular momentum can be expressed as

$$C\mathrm{d}\omega/\mathrm{d}t + \omega\mathrm{d}C/\mathrm{d}t = L + N \tag{1}$$

where the symbols C and ω denote the inertia moment about the polar axis and the angular spin velocity, respectively, and where d/dt denotes the time derivative. The symbol L stands for the tidal retarding torque and N is a correction term accounting for possible non-tidal external torques.

2. Length of day in the geological past

Using information obtained by a number of authors from different kinds of fossil clocks, including corals, bivalves, brachiopods, stromatolites and tidalites, Varga et al. (1998) compiled values of the length of day (l.o.d.) in the remote geological past. They analysed this database by means of statistical techniques both for the Phanerozoic and the Proterozoic, i.e. for an interval of time amounting altogether to 2.5 eons (1 eon = 10^9 a). Their results strongly suggest that the Earth's despinning rate had, on the average, been quite significantly lower during the Proterozoic than during the Phanerozoic.

For the Phanerozoic epoch, which we fix somewhat arbitrarily here from 640 million years ago to the present, the sedimentological lower limit of the Cambrian being rather 570 than 640 million years ago, we find the following linear trend:

$$\text{l.o.d.} = 24.00 - 4.98\,\tau \text{ for } 0 \leqslant \tau \leqslant 0.64 \tag{2}$$

Here τ denotes the time before the present expressed in eons. The abbreviation l.o.d. denotes the length of day at any epoch expressed in hours (h). Hence, in the Phanerozoic, the increase of l.o.d. amounted on the average to 4.98 h/eon, i.e. 1.79 ms/century.

On the other hand, a linear regression performed on the Proterozoic data set yields

$$\text{l.o.d.} = 21.435 - 0.974\,\tau \text{ for } 0.64 \leqslant \tau \leqslant 2.5 \tag{3}$$

leading to an average increase of l.o.d. of only to 0.97 h/eon, or 0.35 ms/century. Admittedly, we need to account for large errors both in the l.o.d. values and in the age determinations, especially as far as Proterozoic data based on tidalites (tidal rhythmites, see Williams, 1989a,b) are concerned. Nevertheless, the fact that the Earth's despinning rate during most of the Proterozoic had been about 5 times lower than during most of the Phanerozoic is hardly questionable (Fig. 1).

If we try with the use of Eq. (3) to estimate the value of l.o.d. at the time of Earth formation (about 4.6×10^9 years ago) or at the end of the Hadean epoch (about 4.0×10^9 years ago), contemporary with the end of the early bombardment and differentiation of the Earth, we find 17.0 and 17.5 h, respectively, for the length of the solar day at those remote times. The Earth–Moon distance was 3.375×10^8 m for $\tau = 4.6\times10^9$ a, and 3.382×10^8 m for $\tau = 4.0\times10^9$ a (i.e. roughly 88% of its present value).

Owing to the fact that in the remote geological past the Moon was closer to the Earth than nowadays, the conclusion that the despinning rate in the Proterozoic should have been 5 times smaller than during most of Phanerozoic seems paradoxical at first view. Indeed, when the Moon was closer, it should, in principle, have raised larger tides and given rise to stronger tidal friction, thus leading to an enhanced despinning rate during the Proterozoic. In fact, there seem to be only two plausible explanations for this paradox: either the tidal retarding torques acting during the

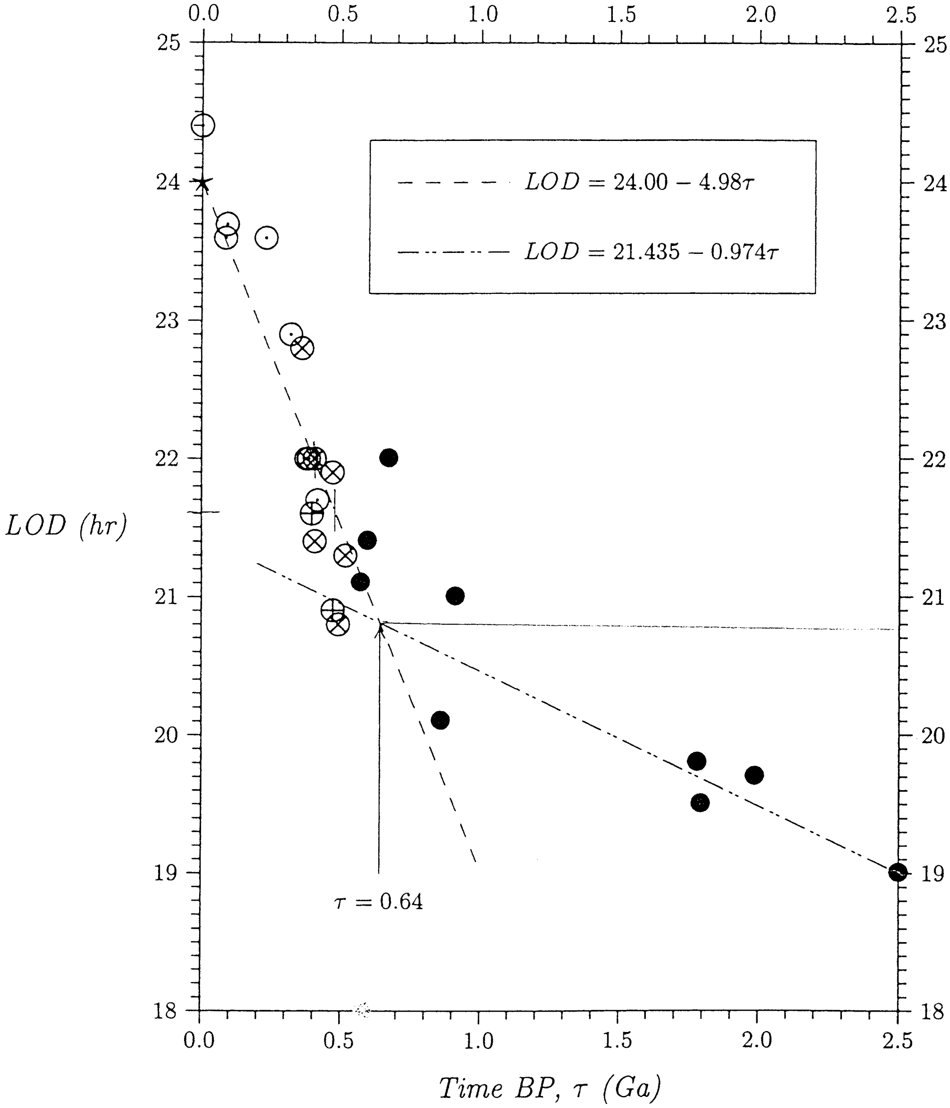

Fig. 1. Variation of l.o.d. during the Phanerozoic and Proterozoic (i.e. during the interval of time extending from the present to 2.5 eons in the past). The star denotes the astronomical datum. The symbols ⊙, ⊗ and ⊕ refer to bivalve, coral and brachiopod data, respectively. The symbol ● denotes information stemming from stromatolites or tidalites (Varga et al., 1998).

Proterozoic were on the average much smaller than those acting during the Phanerozoic because for some reason the oceanic tidal friction was much less efficient, or the tidal torques were approximately the same all the time, but there existed during the Proterozoic a compensating effect to tidal friction brought about by core growth and/or by continuous mantle differentiation and crust formation, possibly associated with plate tectonics. Denis (1986) had already drawn attention to the fact that for most mechanisms, except a growing core, the second term ($\omega \mathrm{d}C/\mathrm{d}t$)

on the l.h.s. of a Eq. (1), is much smaller than the first term ($C\ \mathrm{d}\omega/\mathrm{d}t$). A similar conclusion was reached by Varga (1975) when the possible time variation of C was estimated with the use of the variation of the density models.

Tomecka-Suchon and Denis (1999) investigated the effect on l.o.d. of a growing core, according to a theory of core formation proposed a long time ago by Runcorn (1964). They found that Runcorn's idea very nicely matches the observed or inferred average l.o.d.-curve since Earth formation. However, despite the obvious fact that "... hypotheses, like cats, have nine lives ..." (Tarling and Runcorn, 1973), not many geophysicists would agree at present that the Earth's outer core had been growing significantly a very long time after the Earth itself had been formed. The essential arguments invoked for an early core formation come from isotope geochemistry (Oversby and Ringwood, 1971; Ringwood, 1979). On the other hand, the early differentiation of the Earth's mantle produced an upper mantle which is chemically distinct from the lower mantle (different Mg/Si ratios). Plate tectonics gives rise to continuous planetary differentiation (Carlson, 1994). Moreover, Archean rocks indicate that the early atmosphere-ocean system was very different (Ebinger, 2000) with respect to its present state. To summarize, it is believed that mantle formation had started very early, has been active over all of the Earth's history, most probably at variable rates, and has not yet stopped.

Therefore, it seems promising to investigate whether geomagnetic paleointensities can provide a clue, or possibly an explanation, for the low despinning rate during the Proterozoic, and teach us something about core formation. On the other hand, it seems also interesting to investigate how the paleorotation data can help us interpret the reported paleointensity values of the geomagnetic field.

3. Geomagnetic paleointensities for the last 3.5 eons

The strength of the Earth's magnetic field during geological time is known only quite unsatisfactorily. Notorious difficulties are related with the physical nature of the primary remanent magnetization and with experimental and observational aspects (Prévot et al., 1990; Prévot and Perrin, 1992; Perrin and Sherbakov, 1997). Considering the large time span (3.5 eons) elapsed since the Early Archean, the available geomagnetic paleointensity data are rather sparse, even though we do not request high quality data. This is the case, in particular, for the Proterozoic and Archean epochs, i.e. for by far the longest part of the Earth's history. Nevertheless, we believe that some general conclusions about the geomagnetic field and its relation to the Earth's despinning rate can be drawn, provided our conclusions for the Proterozoic and Archean epochs are formulated with caution.

The geomagnetic paleointensity data set we used in this study is given in the Appendix. Fig. 2 sketches these data graphically. These geomagnetic data were compiled from information published by McElhinny and Evans (1968), Kobayashi (1968), Schwartz and Symons (1969), Hale (1987), Prévot et al. (1990), Prévot and Perrin (1992), Solodovnikov (1995), Morimoto et al. (1997), Perrin and Sherbakov (1997; also reported in Greff-Lefftz and Legros, 1999) and Truhin et al. (1999). The whole data set consists of 218 items, among which 135 pertain to the Phanerozoic 73 to the Proterozoic and 10 to the late Archean. The paleointensity data were extracted from diverse publications, which used different units. In order to achieve an acceptable degree of homogeneity for the database, we converted all the data to relative units by expressing the published quantities, most often average virtual dipole moments, as ratios of the ancient (M) and present (M_o) magnetic moments. Notice that

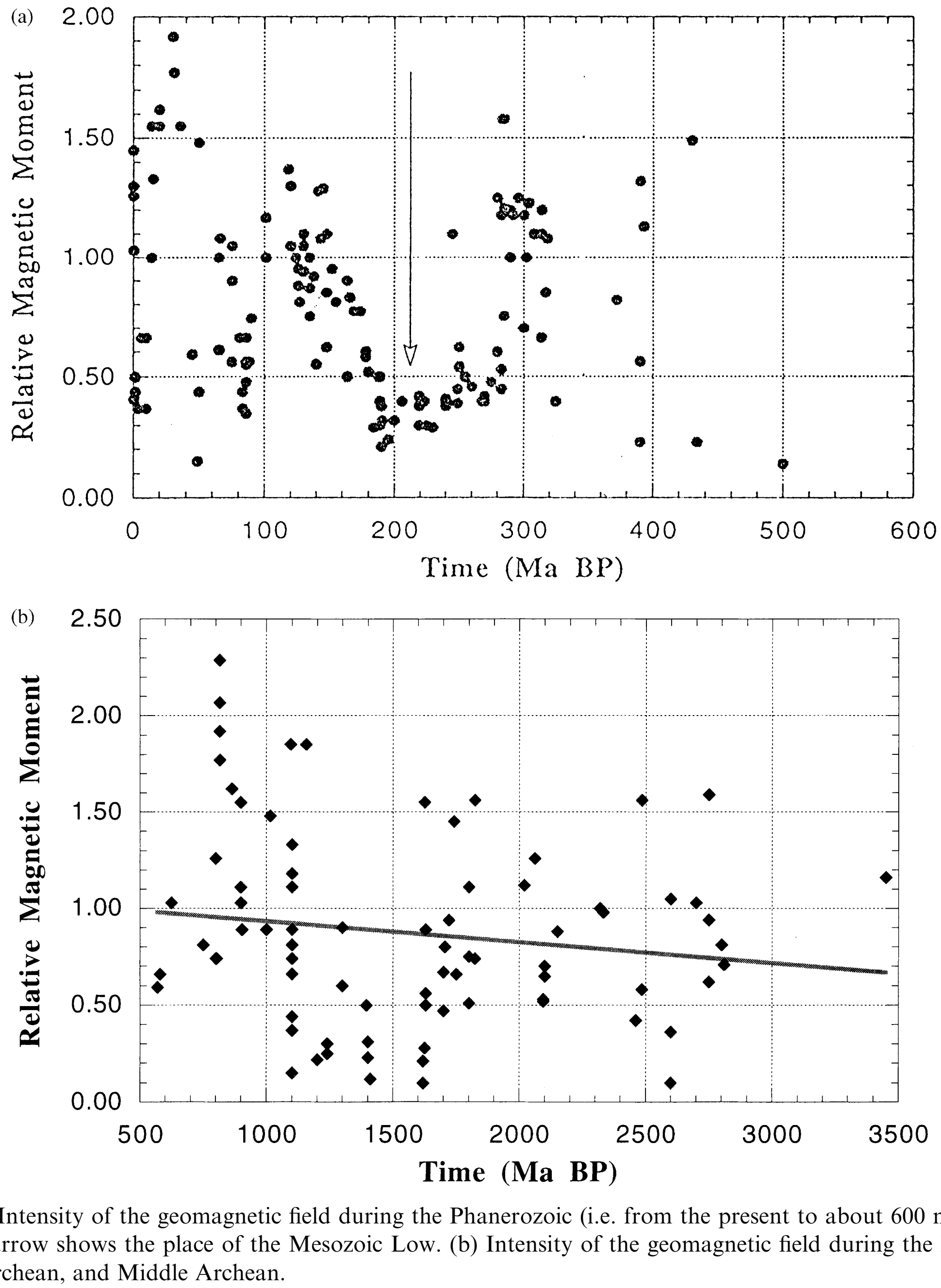

Fig. 2. (a) Intensity of the geomagnetic field during the Phanerozoic (i.e. from the present to about 600 million years ago). The arrow shows the place of the Mesozoic Low. (b) Intensity of the geomagnetic field during the Proterozoic, the Late Archean, and Middle Archean.

$$M = 4\pi Ha^3/\mu_o \tag{4}$$

where $\mu_o = 4\pi\times10^{-7}$ is the magnetic permeability of vacuum, a is the Earth's radius, and H is the horizontal component of the geomagnetic field at the equator. For M_o we took 6.77×10^{22} A m^2 .

Many data were obtained by reading them from enlarged plots. Of course, this circumstance leads unavoidably to small random discrepancies with respect to the original determinations. However, the possible discrepancies remain most certainly within the overall large experimental error bounds, and should have no or only a negligible effect on the approximate reconstruction of the geomagnetic field intensity throughout the geological past. The relative accuracy of the measured magnetic field values is generally comprised between 10 and 30%, but in certain instances the error is larger than the intensity value itself. Quite often, the error bounds on the age determinations are not indicated by the quoted authors. When they are provided, age has most often been obtained radiometrically by means of the K/Ar method. Then, the typical errors are smaller than 10%, and only rarely approach 20%. Altogether, the scatter in the data is quite large, and is most probably not caused only by experimental errors. An essential part of the scatter obviously stems from the fact that the Earth's magnetic field undergoes large-amplitude short-periodic changes. In our study, we are interested in paleointensities averaged over time scales of several million years. Such meaningful averages are hardly available from the limited information at hand.

4. An attempt of complex investigation of geomagnetic intensity and paleorotation data

There is only a very limited amount of reliable paleointensity data available for most of the Earth's history. Therefore, only some very general conclusions on the long-term behaviour of the geomagnetic field with time may be drawn. We summarize these conclusions as follows:

(1) According to Prévot and Perrin (1992), the magnetic moment of the Earth varies mainly between 2×10^{22} and 12×10^{22} Am^2, and this scatter is similar both for the Phanerozoic and the Precambrian. There is no serious observational basis for drawing conclusions on possible long-term variations on a geological time scale between the present epoch and the Middle Archean era, 3.5×10^9 years ago (see also Table 1). An estimate of a possible variation of the magnetic moment can be attempted by means of the scaling law derived by Busse (1976) for the planetary magnetic moment M, namely

$$M \propto \rho^{1/2}\omega a_c^4 \tag{5}$$

which holds for a planet for which the necessary conditions for the existence of a self-sustaining dynamo are fulfilled. In Eq. (5), ρ is the density, a_c is the core radius, and ω is the planetary rotation rate. If we suppose that core formation was finished at the starting point of the time interval considered in this paper, i.e. 3.5×10^9 years ago, we may conclude from Eq. (5) that

$$M_i = M_o\omega_i/\omega_o \tag{6}$$

Table 1
Paleointensity data[a] for the Phanerozoic (from 570 million years ago until the present), Proteorozoic + Archean (0.57–2.50 aeons BP and 2.50–4.00 aeons BP)

Phanerozoic (number of data $n=135$)		
$(0–100)\times10^6$ years ago	$(100–250)\times10^6$ years ago	$(250–570)\times10^6$ years ago
$n=47$	$n=56$	$n=39$
Mean: 0.8962	Mean: 0.5385	1.0408
RMS: ±0.0290	RMS: ±0.0267	RMS: ±0.0442
The whole Phanerozoic		
Mean: 0.7867		
RMS: ±0.0190		
The Phanerozoic without the Mesozoic		
$[(100–250)\times10^6$ years ago]		
$n=79$		
Mean: 0.9735		
RMS: ±0.0258		
Proterozoic + Archean		
$[(0.57–4.00))\times10^9$ years ago]		
$n=83$		
Mean : 0.9379		
RMS: ±0.0541		

[a] Data are expressed in relative units, i.e. the ancient magnetic moment values are divided by $M_o=6.77.1012$ Am^2 (the present magnetic moment).

where subscripts 'o' and 'i' refer to the present epoch and any previous epoch, respectively. Using this rough estimation, we tentatively conclude that M_i has been 15% larger than M_o at the boundary Proterozoic/Phanerozoic, 0.57×10^9 years ago. Similarly, 2.5×10^9 years ago, M_i should have been about 25.7% larger than the present-day value of $M_o=6.77\times10^{22}$ A m^2. To perform these estimations, we took $\omega_o=15°/\text{h}$ for the present spinning rate, then, using Eq. (3), we found that 0.57×10^9 years ago, $\omega_i=17.25°/\text{h}$, and 2.5×10^9 years ago, $\omega_i=18.85°/\text{h}$, respectively.

(2) During the last 4 aeons, the geomagnetic field was always mainly dipolar. The latter circumstance is clearly documented by a number of authors (Schwartz and Symons, 1969; Prevot et al., 1990; Morimoto et al., 1997; Perrin and Sherbakov, 1997).

(3) It becomes evident from an inspection of Table 1, which results from an analysis of our data base, that during the Mesozoic epoch (250–200 million years BP) a geomagnetic intensity minimum occurred. This phenomenon has been known for some time, and details concerning this 'Mesozoic low' can be found e.g. in the paper of Perrin and Sherbakov (1997). During the Mesozoic, the average strength of the geomagnetic dipole amounted to approximately 50% of the value typical for the Paleozoic and the Cenozoic.

Moreover, the despinning history of the Earth exhibits a significant non-linear behaviour in at least two instances, namely (1) around the border between the Proterozoic and the Phanerozoic (this strong change in the despinning rate has already been mentioned), and (2) a minimum of the angular speed between 100 and 50 million years ago.

The mean angular rates for the Phanerozoic and Proterozoic yield for the time derivative of the inertia moment about the polar axis (Varga et al., 1998)

$[dC/dt]_{\text{Phanerozoic}} = -4.114\times10^{18}$ kg m^2 s^{-1}

respectively

$[dC/dt]_{\text{Proterozoic}} = -6.36\times10^{17}$ kg m^2 s^{-1}.

Let us assume now that the acceleration process of the Earth's axial rotation is caused by the evolution of the Earth's interior, which we characterize by the difference

$[dC/dt]_{\text{Proterozoic}} - [dC/dt]_{\text{Phanerozoic}} = 3.478\times10^{18}$ kg m^2 s^{-1}.

Consequently, we assume here that in the course of the Proterozoic, the formation of the inner structure of our planet was reducing the effect of the despinning caused by the tides, while in the course of the Phanerozoic, the formation of the Earth was more or less complete, and the despinning was determined by tidal friction alone. In this way, during the 4 eons spanned by the Archean plus the Proterozoic, the variation of C due to an inner mass redistribution is 4.39×10^{35} kg m^2, i.e. about 5% of the value of the moment of inertia of the core at present, and about 0.5% of the moment of inertia of the present entire Earth. Indeed, Denis et al. (1998) indicate for the moment of inertia of the Earth the value 8.012×10^{37} kg m^2, and for the moment of inertia of the core 9.0416×10^{36} kg m^2. This means that the seemingly large discrepancy between the Proterozoic and Phanerozoic despinning rates can be explained by a moderate mass redistribution within the Earth.

The second problem which needs further considerations is connected with the Mesozoic low of the geomagnetic dipole. Table 2 shows that during the time interval 250–100 million years ago, which comprises the time of the geomagnetic low, the oceanic tidal torque and the angular speed (after removal of the main linear trend) also show an extremum. Figure 94 on page 199 of the book published by Zonenshain and Kuzmin (1997) shows that the mean continental speed values also were lower by 1/3 during the Mesozoic than during the Cenozoic and Paleozoic. Presently, there is no satisfactory interpretation of these coincidences. It should be mentioned, however,

Table 2
Variation of the oceanic tidal torque (for M_2), angular speed (with linear trend removed)[a], mean speed of the continental plates[b] and magnetic dipole moment[c]

$(0–100)\times10^6$ BP ~Cenozoic	$(100–250)\times10^6$ BP ~Mesozoic	$(250–570)\times10^6$ BP ~Paleozoic
Oceanic tidal torque		
-5.00×10^{16} J	-4.27×10^{16} J	-4.77×10^{16}
Anomaly of the angular speed with linear trend removed		
+0.024 h	−0.433 h	−0.124 h
Mean speed of the continental plates		
6 cm a^{-1}	4 cm a^{-1}	6 cm a^{-1}
Geomagnetic dipole moment (in relative units)		
0.896	0.539	1.041

[a] Varga et al. (1998).
[b] Zonenshain and Kuzmin (1997).
[c] Table 1.

that during most of the Mesozoic, all continental plates were gathered within the single super-continent *Pangaea*.

The latter circumstance led probably (1) to the reduction of the mean plate motion rates, and (2) to the diminution of the tidal displacements. The correlation of these two surface phenomena with the observed reduction of the average dipole strength during the Mesozoic constitutes an exciting problem, unless it is merely a coincidence. It is worth mentioning that the action of the ocean load, which was anomalous during the considered time interval, generated displacements of the core–mantle boundary which were of about the same amplitude as the displacements of the Earth's surface (Groten and Molodensky, 1999). The former could possibly lead to significant variations of the topography of the core–mantle boundary, and in this way perhaps influence the magnitude of the dipole component of the geomagnetic field 250–100 million years ago.

5. Characteristic time for tidal despinning of the earth

For interpreting events of the Earth's tidal history, it is useful to consider the characteristic time of the luni-solar despinning. This tidal time scale is defined by Hubbard (1984, p. 103) as follows:

$$\tau_{\mathrm{tidal}} = 2\pi Mc^6 / \left[3k_s GM_{\mathrm{m(s)}}^2 a^3 T_{\mathrm{o}}\delta\right] \tag{7}$$

where M and $M_{\mathrm{m(s)}}$ are the masses of the Earth and the Moon (or the Sun), respectively, k_{s} is a secular Love number ($k_{\mathrm{s}}=0.94$), G denotes the gravitational constant, and T_{o} is the primordial rotation period of the Earth. The numerical value of the latter, estimated by extrapolation of formula (3) in which we put $\tau=4.6$, amounts to 17.0 h. The phase shift angle δ is expressed in terms of the quality factor Q, which has most probably kept throughout geological history a value comprised between 50 and 200, by the relation $\delta=\mathrm{atan}\ (1/Q)$. Using either of those two values for Q, we find for the lunar tides either $\tau_{\mathrm{tidal}} \approx 2.5\times10^{10}$ years or $\tau_{\mathrm{tidal}} \approx 6.2\times10^9$ years. For the solar tides, τ_{tidal} becomes roughly 10^{11} years.

Two remarks should be made here: (1) If the tidal history is dealt with by means of the elastic Love number $k=0.30$, i.e. in the way Hubbard (1984) proceeded, we find thatτ_{tidal} is about 1.1×10^{11} years for lunar tides and 5.5×10^{11} years for solar tides. The latter values seem to be much too large for characterizing the Earth's tidal evolution in an adequate way. Here we use the secular Love number $k_{\mathrm{s}} \approx 0.94$ instead of the elastic Love number $k \approx 0.30$. (2) If we use the same formulation for investigating the tidal evolution of the Moon in response to the tides caused on the Moon by planet Earth, we obtain $\tau_{\mathrm{tidal}} \approx 1\times10^7$ years. This rather short characteristic tidal time scale of 10 million years shows that the tidal evolution of the Moon had been completed soon after the formation of the Earth–Moon system. All the time since then, the Moon has revolved in a synchronous orbit about the Earth.

6. Conclusions

There are two main data bases at our disposal for interpreting long-term variations of the length of day throughout most of the geological history of the Earth.

The first set of data contains information derived from paleontological and paleosedimentological data about long-term variations of the Earth's spin. From these data we infer that the Earth's rotational history has been essentially non linear. Indeed, the ratio of the despinning rate during the Proterozoic with respect to the despinning rate during the Phanerozoic is approximately 7.5. Moreover, the data indicate that there occurred a statistically significant relative minimum in a time interval which essentially coincides with the Mesozoic, when the continents had merged into the supercontinent *Pangaea*.

The second set of data contains paleomagnetic information concerning the temporal variations of the relative magnetic dipole moment during the last 3.0 eons, an interval of time which spans most of the Earth's lifetime. However, the variation of the strength of the geomagnetic field throughout geological time is known only poorly, because there exist some fundamental experimental and methodological difficulties to obtain meaningful and accurate values of the geomagnetic paleointensities. Nevertheless, in spite of these obvious uncertainties which characterize the geomagnetic paleointensity data, we believe that we may safely draw the following conclusions:

Although the scatter of the data showing the strength of the Archean and Proterozoic geomagnetic paleodipole is considerable, there is no obvious pattern which could lead us to conclude that the dipole field of these epochs would, on the average, have differed significantly from the present dipole field. However, it is almost certain that a pronounced relative minimum of the magnetic dipole intensity occurred during the Mesozoic (roughly at the same epoch when a relative minimum occurred also in the despinning rate). When trying to correlate the two data sets, it is worthwhile to consider that, according to most authors, the separation of the core from the mantle occurred during the Hadean, i.e. within the first 600 million years after the Earth's formation, and that the differentiation processes within the Earth's mantle have probably not stopped yet (Carlson, 1994). It seems plausible that differentiation and fractionation of mantle material had been more effective during the Proterozoic than later on. However, some kind of an upper limit can be put on the efficiency characterizing mass redistribution within the Earth by noting that the corresponding total change of the polar moment of inertia, derived from the respective despinning rates during Proterozoic and Phanerozoic, was only of the order of 1%.

Considering both the similarities and the differences exhibited by the two data sets, we obtain the following synoptic view of the nonlinear patterns occurring in the observed paleorotation and paleointensity curves:

Similarities	Differences
• The mean values of the despinning rates in the Proterozoic and in the Mesozoic are similar.	• The anomaly of the despinning rate during the Mesozoic is correlated with a minimum of the geomagnetic field intensity. This coincidence suggests that the anomaly of the rotation speed during the middle Phanerozoic influences to extent, or is related to, the processes occurring within the Earth's interior (in particular, the phenomena which have their origin in the core). For the Proterozoic nonlinearity, no long-term variation of the geomagnetic paleointensity has been detected. Thus, it is difficult to relate this anomaly to processes within the Earth's core.

Similarities	Differences
• During the Mesozoic, the existence of the *Pangaea* significantly influenced the tidal torque and therefore the rate of the Earth's despinning. According to paleogeographic reconstructions, a similar supercontinent existed during a large part of the Proterozoic.	• The duration of the low despinning rate of the Proterozoic is comparable to the characteristic tidal despinning time of the Earth whereas the anomaly occurring between 250 and 100 million years ago (Mesozoic anomaly) is not.

On this basis we may conclude that despinning anomalies of the Proterozoic and of the Mesozoic have probably different, yet unknown, causes. It is worth mentioning again that a relatively small mass redistribution within the mantle can influence (e.g. accelerate) greatly the Earth's rotation in the course of the Proterozoic. If we except the Mesozoic, the despinning history during the Phanerozoic can adequately be explained considering only the oceanic tidal torque. Therefore, we infer that during the last 0.5×10^9 years, the inner structure of our planet remained more or less the same. On the other hand, the coincidence in the Mesozoic of anomalies concerning the paleomagnetic intensity, the tidal torque, the despinning rate, and the plate speed suggests that in the time interval extending roughly between 250 and 100 million years ago, possibly some mechanism involving external sources acted in the Earth's core.

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Appendix. The paleointensity data set used in this study

The symbols denoting the different columns have the following meanings:

τ age of the rock specimen in 10^6 a (determinded radiometrically or stratigrafically estimated) on which megnetic paleointensity measurements have been performed

$\Delta\tau$ the estimated error bound of the age determination expressed in 10^6 a
F ancient field strength (normalised with respect to the present day value)
ΔF the corresponding inferred error bound

References are abbreviated as follows:
Mc 68 = McElhinny and Evans (1968); K 68 = Kobayashi (1968); SS 69 = Schwartz and Symons 1969); H 87 = Hale, 1987; P 90 = Prévot et al., 1990; PD 92 = Prévot and Perrin, 1992; S 95 = Solodovnikov, 1995; PS 97 = Perrin and Scherbakov, 1997; M 97 = Morimoto et al., 1997; T 99 = Truhin et al., 1999.

Item	τ	F	$\Delta\tau$	ΔF	Ref.
Phanerozoic					
1	0.0	0.95			T99
2	0.0	1.00			T99
3	0.0	1.05			T99
4	0.0	1.30			T99
5	0.4	0.88			T99
6	0.5	0.81			T99
7	1.0	1.45	0.99	0.71	P90
8	1.5	1.10			T99
9	3.5	1.30	1.50	0.50	P90
10	6.0	1.30		0.37	PS97
11	9.5	1.05			T99
12	10.0	0.94		0.50	T99
13	13.7	0.30	8.75	0.60	P90
14	14.0	1.00		0.20	T99
15	15.0	0.90		0.29	PS97
16	20.0	0.87		0.40	T99
17	20.0	0.75		0.35	T99
18	30.0	0.75		0.75	T99
19	31.0	0.92		0.42	199
20	36.0	0.56	10.0		PS97
21	45.0	0.55		0.05	T99
22	49.0	1.28		0.06	T99
23	50.0	1.08		0.03	T99
24	50.0	1.29		0.04	T99
25	65.0	0.66		0.48	P97
26	65.0	1.10			T99
27	67	1.26	2	0.30	P90
28	75	0.85		0.17	T99

Item	τ	F	$\Delta\tau$	ΔF	Ref.
29	75	0.62		0.21	T99
30	75	0.95		0.16	T99
31	81	0.81		0.09	T99
32	82	1.03	4	0.05	P90
33	83	0.50		0.06	T99
34	86	0.83		0.21	T99
35	86	0.77			T99
36	86	0.90		0.06	T99
37	86	0.77			T99
38	86	0.60			T99
39	88	0.58		0.06	T99
40	90	0.52		0.08	T99
41	101	0.29			T99
42	101	0.52			T99
43	118	0.50			T99
44	120	0.50		0.05	T99
45	120	0.46			T99
46	124	0.44	18		PS97
47	126	0.41	5	0.15	P90
48	126	0.82		0.02	SS69
49	126	0.17		0.07	SS69
50	130	0.40		0.06	T99
51	130	0.30		0.06	T99
53	135	0.38		0.07	T99
54	135	0.32		0.05	T99
55	135	0.24		0.05	T99
56	135	0.37		0.15	PS97
57	138	0.23	28	0.08	SS69
58	140	0.32			T99
59	140	0.50	3	0.12	P90
60	143	0.40			T99
61	146	0.44	3	0.10	P90
62	148	0.30		0.07	T99
63	148	0.38		0.07	T99
64	148	0.42			T99
65	152	0.40		0.06	T99
66	155	0.48		0.17	PS97
67	163	0.37	8	0.10	P90
68	166	0.35		0.07	PS97
69	169	0.30		0.07	T99
70	174	0.29		0.04	T99

Item	τ	F	$\Delta\tau$	ΔF	Ref.
71	178	0.38		0.05	I99
72	178	0.41		0.05	I99
73	180	0.66	3	0.05	P90
74	184	0.66		0.27	PS97
75	188	1.10		0.04	T99
76	189	0.39		0.05	T99
77	189	0.54		0.06	T99
78	189	0.45		0.04	T99
79	189	1.55	10	0.15	P90
80	190	1.62	20		M97
81	190	1.33	20		M97
82	191	0.62		0.05	T99
83	195	0.55			PS97
84	200	1.48			PP92
85	206	0.56		0.07	PS97
86	219	0.40		0.05	T99
87	219	0.50			T99
88	219	0.46			T99
89	223	0.40		0.04	T99
90	225	1.32	30	0.20	SS69
91	230	0.42		0.10	T99
92	240	1.55	50		M97
93	240	0.48		0.08	T99
94	245	0.56		0.06	PS97
95	249	1.25			T99
96	249	0.60		0.18	T99
97	250	1.00	16	0.15	P90
98	250	1.00	18	0.14	FF92
99	255	0.74		0.24	PS97
100	260	1.92	30		M97
101	268	0.61		0.15	FF92
102	270	0.53			T99
103	270	0.45		0.08	T99
104	275	1.17		0.28	PS97
105	280	1.08	5	0.11	PP92
106	280	1.18		0.08	T99
107	283	1.58		0.10	T99
108	283	1.21		0.07	T99
109	283	1.58		0.05	T99
110	284	0.75		0.04	I99
111	285	1.00		0.17	PS97

Item	τ	F	$\Delta\tau$	ΔF	Ref.
112	285	0.56		0.12	SS69
113	285	1.13		0.10	SS69
114	290	1.00		0.06	T99
115	290	1.20		0.06	T99
116	292	1.18		0.12	T99
117	296	1.25		0.15	T99
118	300	1.77	35		M97
119	300	1.49		0.03	SS69
120	302	1.18		0.08	T99
121	304	0.70		0.11	T99
122	308	1.00		0.05	T99
123	314	1.23		0.14	T99
124	314	1.10		0.06	T99
125	314	1.37		0.10	T99
126	317	1.10		0.07	T99
127	318	1.20		0.18	T99
128	325	0.85		0.05	T99
129	325	0.66		0.05	T99
130	332	1.08		0.08	T99
131	390	1.55	40		M97
132	390	0.59	45		M97
133	430	0.15	25		M97
134	434	0.14	75	0.07	SS69
135	500	0.44	40		M97
Protorezoic					
1	570	0.50	35	0.16	SS69
2	580	0.59	35		M97
3	625	0.86		0.35	H87
4	750	0.66	100		M97
5	751	0.56	75	0.12	SS69
6	800	0.81	150		M97
7	800	1.03	90		M97
8	800	0.58	60	0.08	SS69
9	805	0.89		0.07	SS69
10	817	0.67	70	0.03	SS69
11	817	0.47	70	0.03	SS69
12	817	0.80	70	0.02	SS69
13	817	0.94	70	0.09	SS69
14	863	1.45	115	0.12	SS69
15	899	0.66	90	0.11	SS69
16	900	0.81	170		M97

Item	τ	F	$\Delta\tau$	ΔF	Ref.
17	900	0.74	80		M97
18	903	0.75	140	0.05	SS69
19	1000	1.26	60		M97
20	1015	1.11	60	0.07	SS69
21	1095	0.51		0.05	SS69
22	1100	1.92	100		M97
23	1100	2.07	100		M97
24	1100	1.77	100		M97
25	1100	0.74	100		M97
26	1100	0.89	100		M97
27	1100	2.29	100		M97
28	1100	1.11	100		M97
29	1100	1.53		0.48	H87
30	1100	1.55	100		M97
31	1100	1.03	100		M97
32	1100	1.62	100		M97
33	1155	0.94		0.12	K68
34	1200	0.74		0.11	SS69
35	1240	1.56	50	0.16	SS69
36	1240	1.12	50	0.12	SS69
37	1300	0.89	110		M97
38	1300	0.81		0.10	K68
39	1395	1.26		0.05	SS69
40	1400	1.48	40		M97
41	1400	0.52	50	0.07	SS69
42	1400	0.59	50		M97
43	1410	0.53		0.12	SS69
44	1620	0.65		0.07	SS69
45	1620	0.70		0.07	SS69
46	1625	0.88		0.09	SS69
47	1625	1.00		0.13	SS69
48	1630	0.98	180	0.12	SS69
49	1630	0.89	40		M97
50	1630	1.11	200		M97
51	1700	1.33	20		M97
52	1700	0.44	15		M97
53	1705	0.42	140	0.13	SS69
54	1720	0.15	30		M97
55	1740	1.56	200		SS69
56	1750	1.85	220		M97
57	1800	0.58	60	0.08	SS69

Item	τ	F	$\Delta\tau$	ΔF	Ref.
58	1800	0.37	80		M97
59	1800	0.66	60		M97
60	1825	0.71		0.12	K68
61	1825	0.36		0.08	SS69
62	2020	0.15	110		M97
63	2060	0.10		0.04	SS69
64	2095	1.05		0.17	SS69
65	2095	1.03		0.07	SS69
66	2100	1.63		0.42	H87
67	2100	1.18	110		M97
68	2150	0.81	60		M97
69	2320	0.62		0.14	SS69
70	2330	0.74	110		M97
71	2460	1.85	120		M97
72	2485	1.59		0.10	SS69
73	2485	1.16		0.14	S68
Archean					
1	2600	0.74			H87
2	2600	1.00			H87
3	2600	1.61		0.37	H87
4	2700	1.55		0.04	Mc68
5	2750	0.32		0.12	H87
6	2750	0.22	70		M97
7	2750	0.39		0.10	H87
8	2800	0.68		0.16	H87
9	2810	0.90		0.15	H87
10	3450	0.30	20		M97

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