Planet Within a Planet: Rotation of the Inner Core of Earth

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The time dependence of the orientation of Earth's inner core relative to the mantle was determined using a recently discovered 10-degree tilt in the axis of symmetry of the inner core's seismic-velocity anisotropy. Two methods of analyzing travel-time variations for rays traversing the inner core, on the basis of 29 years of data from the International Seismological Centre (1964–1992), reveal that the inner core appears to rotate about 3 degrees per year faster than the mantle. An anomalous variation in inner-core orientation from 1969 to 1973 coincides in time with a sudden change ("jerk") in the geomagnetic field.

The inner core of Earth, a crystalline region with a radius of 1220 km, was discovered 60 years ago (*1*). Seismological evidence seems to suggest that the rotation rate of the inner core differs from that of Earth's surface. This differential rotation may occur because of the decoupling effect of the outer core, the 2270-km-thick liquid-metal shell that separates the solid inner core from the rocky mantle. Song and Richards (*2*) found that the difference in arrival times for seismic waves traveling along DF and BC paths through the core (Fig. 1, A and B) changed from the late 1960s to the early 1990s. Under the assumption that this change is caused by the inner core rotating with respect to the mantle, they obtained a rate of 1.1° year^{-1} on the basis of waves traveling from earthquakes in the South Sandwich Islands to the seismic station at College, Alaska. Here, we used independent data from ~2000 stations to solve the inverse problem of determining the time-dependent orientation of the inner core, and also found that there is evidence of a differential rotation.

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If the seismological structure of the core were not tilted or offset relative to the rotation axis, there would be little hope of monitoring the orientation of the inner core with respect to the mantle. Poupinet *et al.* (*3*) first observed that seismic waves travel fastest along the rotation axis (north-south direction) of the inner core. Morelli *et al.* (*4*) and Woodhouse *et al.* (*5*) then showed that the inner core is transversely anisotropic, a result confirmed by more recent work (*6–13*). However, none of these studies could resolve whether the symmetry axis of the inner-core anisotropy was different from Earth's rotation axis. Thus, seismic anisotropy did not seem to help in monitoring the orientation of the inner core, until Su and Dziewonski (*14*) showed that 29 years of data (1964–1992) from the Bulletins of the International Seismological Centre (ISC) are best fit with the inner core having an anisotropy axis (fast direction) tilted by ~10° with respect to Earth's rotation axis (*15*). For the full data set, they obtained a fast-axis orientation with latitude 79.5° ± 1°N and longitude 160° ± 5°E.

Although the specific cause of the anisotropy is still debatable, the pattern is thought to be constant over periods of centuries or more (*16–18*). Thus, any temporal variations observed in the orientation of the (tilted) axis of fast velocity can be attributed to a differential rotation of the inner core with respect to the mantle. Also, the seismologically determined pattern of anisotropy contains higher order structure than discussed here, and the presence of longitude-dependent structure may provide an additional means of quantifying the orientation of the inner core with time.

Here, we repeated Su and Dziewonski's analysis (*14*) but divided the interval from 1964–1992 into six time periods of 5 years each, with a 1-year overlap between the periods 1984–1988 and 1988–1992. To diminish the impact of a sixfold reduction in data (on average) relative to the earlier analysis (*14*), we doubled the distance range of the analysis to 150° to 156°. The remaining threefold net reduction of data density may have led to an increased noise level in the present study relative to the earlier study, particularly in the earlier years when the network of reporting stations was much sparser than it is now. Nonetheless, applying the cylindrical anisotropy stacking (CAS) procedure (*14*) to the six time periods yielded a distinct eastward rotation of the inner-core anisotropy pattern between the earliest and latest time periods (Fig. 2). This conclusion is reached not only from the least-squares fit to the data, but also by tracking the patterns of individual data points.

In our first experiment, the solution of the inverse problem (the orientation of the best axis of symmetry for the inner core) shows an overall increase in longitude over time, with a straight-line least-squares fit yielding a slope of 2.27° ± 0.90° year^{-1} (Fig. 3A). Much of the standard error for this result comes from the apparently anomalous orientations for the period 1969–1973; ignoring this time period yields a better con-

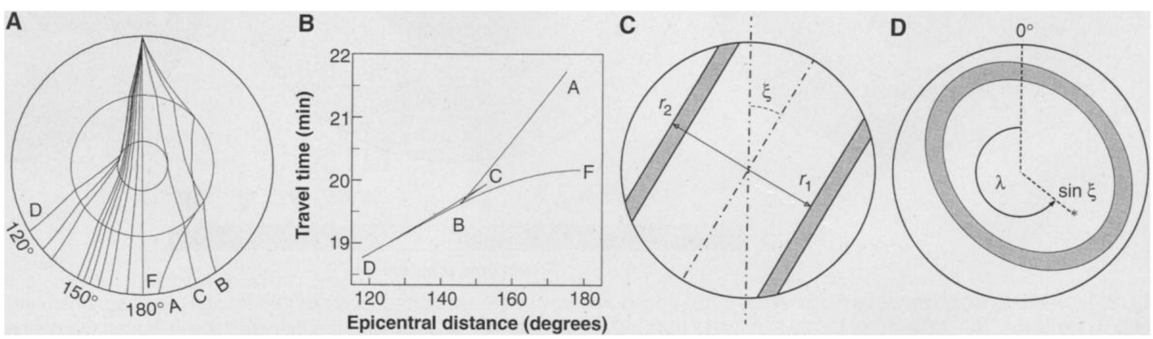

Fig. 1. (**A**) Ray paths for AB, BC, and DF branches of the *PKP* seismic phase. The DF branch, which enters the inner core, is at the left; the AB and BC branches, which are the rays turning in the outer core, are at the right. The BC branch becomes diffracted at ~153° arc distance. (**B**) Travel times for the *PKP* phases as a function of epicentral distance. (**C**) Polar cross section through the inner core, showing the region (shaded), comprising rays with bottoming radii between r_1 and r_2, within which any ray parallel to the dashed line through the center of Earth, inclined at the angle ζ to the axis of rotation, would contribute to a particular cylindrical anisotropy stack. (**D**) Equatorial cross section through the inner core viewed from above the North Pole. The shaded region is the same as in (C). The projection of the intersection of the dashed line with the surface is marked by an asterisk; its longitude is λ and the arc distance from the pole is ζ [adapted from (*14*)].

strained slope of 3.02° ± 0.43° year^{-1}. To investigate the anomaly associated with the period 1969–1973, we obtained 5-year sliding-window averages with a 1-year increment. Although the estimates were not independent, they suggest that there is a 5-year period for which the results are anomalous.

There was no change in the procedure of acquiring or reducing the original seismological observations that could be responsible for this effect; our analysis was performed uniformly for the entire 29 years. We therefore infer that this anomaly is real. It coincides temporally with the geomagnetic "jerk" (a sudden change in the strength of Earth's magnetic field) of 1969–1970 (*19*, *20*). The variation in latitude becomes statistically insignificant (−0.14° ± 0.19° year^{-1}) if results for all periods are used (*21*). Thus, we find no conclusive evidence for a time variation in the latitude (wobble) of the fast-velocity axis, and we assume that the inner-core motion is purely a rotation.

Our second approach is even more general and considers the spatiotemporal variations of the entire seismic-velocity field, not just the optimal axis of symmetry. After an initial experiment (*22*), we allowed for a time-independent component for the expansion of travel-time residuals:

$$\delta t(\vartheta, \varphi, \tau) = \sum_{\ell} \bar{A}_{\ell 0} p_{\ell 0} (\cos \vartheta) + \sum_{\ell, m \neq 0} [\bar{A}_{\ell m} \cos(m\varphi) + \bar{B}_{\ell m} \sin(m\varphi)] p_{\ell m}(\cos \vartheta) + \sum_{\ell, m \neq 0} \{\hat{A}_{\ell m} \cos[m(\varphi - \omega\tau)] + \hat{B}_{\ell m} \sin[m(\varphi - \omega\tau)]\} p_{\ell m}(\cos \vartheta) \quad (1)$$

where ϑ and φ are spherical coordinates, τ is time (in years since 1964), $p_{\ell m}$ are fully normalized associated Legendre polynomials, ℓ is angular degree and m is secular order, $\bar{A}_{\ell m}$ and $\bar{B}_{\ell m}$ are the time-independent components, and $\hat{A}_{\ell m}$ and $\hat{B}_{\ell m}$ are the time-dependent components. The first sum involves only axially symmetric terms (m = 0), for which the time dependence cannot be ascertained, whereas the second sum represents nonaxial terms ($m \neq 0$) that do not vary with time. The first and second sums might represent, for example, a failure to

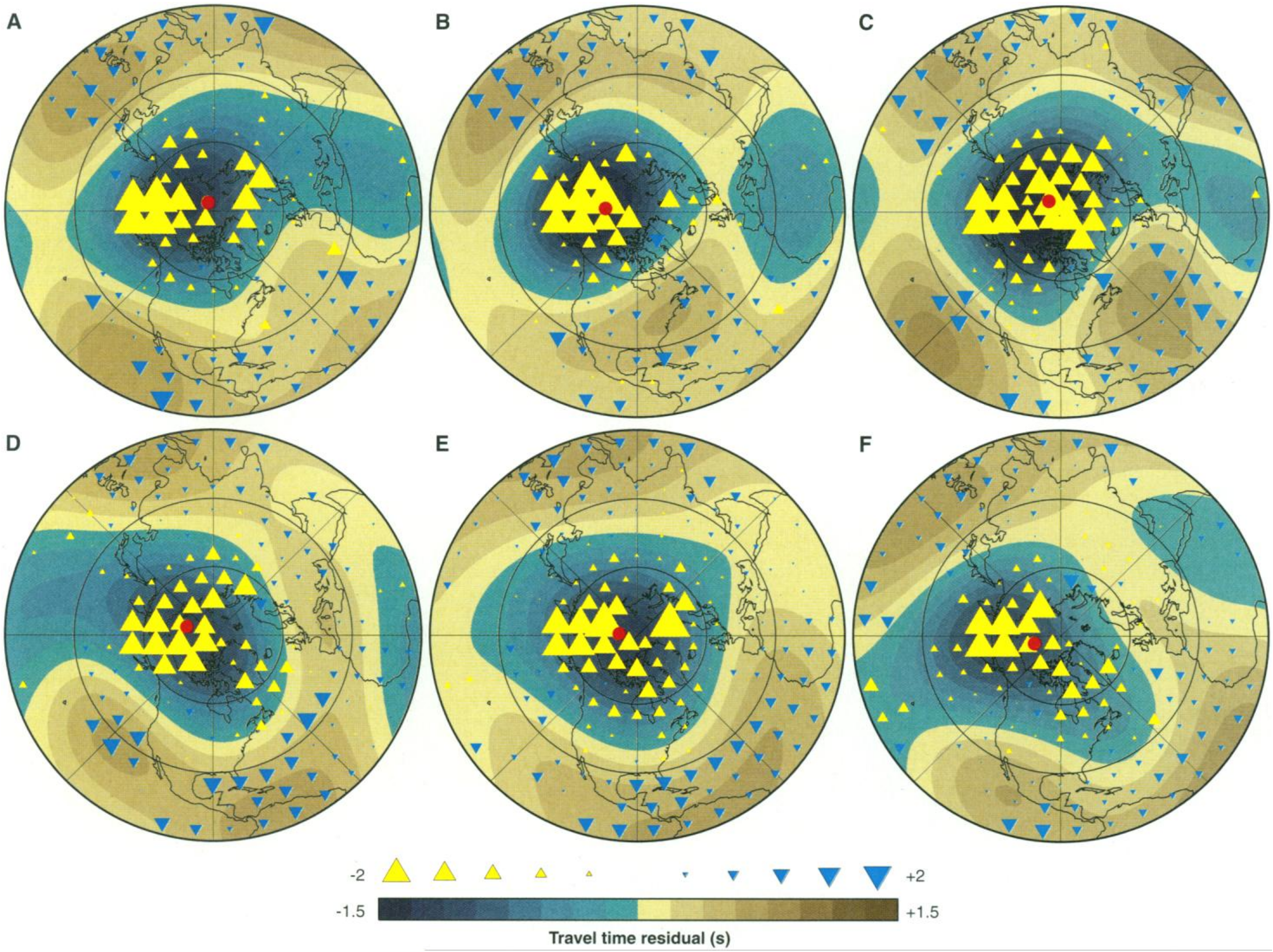

Fig. 2. Travel-time residuals obtained from the CAS procedure for six different time windows: (**A**) 1964–1968 (16,952 arrivals, 163 CAS estimates); (**B**) 1969–1973 (27,788 arrivals, 164 CAS estimates); (**C**) 1974–1978 (38,840 arrivals, 179 CAS estimates); (**D**) 1979–1983 (46,828 arrivals, 178 CAS estimates); (**E**) 1984–1988 (65,290 arrivals, 186 CAS estimates); and (**F**) 1988–1992 (66,518 arrivals, 184 CAS estimates). The residuals were plotted using a polar projection centered on the North Pole. The outermost circle corresponds to the equator, the two inner circles correspond to 60°N and 30°N, and the thin solid line indicates the continents. Yellow and cyan triangles are the negative and positive residuals, respectively. The color contours are the best fitting spherical harmonics up to degree 4. The red dots are the best axis of symmetry of inner-core anisotropy for the corresponding time period. An eastward (counterclockwise) rotation of the pattern between the earliest and latest periods is evident; compare (A) or (C) with (D), (E), and (F).

remove completely the effects of velocity anomalies in the mantle, even though the residuals were corrected for that effect [see (*14*)]. The last term corresponds to the differential rotation of the inner core. The advantage of this approach is that, in contrast with the CAS analysis described above, it allows for structure within the inner core beyond that of uniform transverse anisotropy [for example, (*23*)]. We formulated a nonlinear least-squares problem in which we found the coefficients $\bar{A}_{\ell m}$, $\bar{B}_{\ell m}$, $\hat{A}_{\ell m}$, and $\hat{B}_{\ell m}$ and the frequency ω for which the largest variance reduction is achieved, and obtained ω corresponding to a differential rate of rotation of 3.28° year^{-1}.

Our results agree in sign, and to within a factor of 3 in amplitude, with the findings of Song and Richards (*2*). The two studies were carried out independently, using different data sets and different methods of analysis (*24*). We solved the inverse problem for the optimal location of the axis of symmetry during six time intervals. Each estimate was independent, so that the observed change in longitude of the anisotropy axis with time indicates a differential rotation of the inner core.

Nevertheless, complications remain in trying to understand the observations in

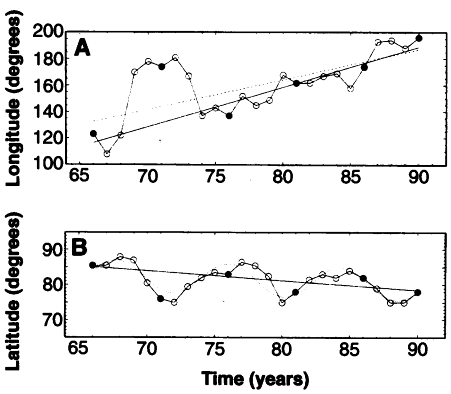

Fig. 3. (**A**) Longitude of the best-fit axis of fastest seismic-wave velocities, illustrating the time-dependent orientation of Earth's inner core. Results averaged over 5-year time intervals at 5-year increments (filled circles) and averages with 1-year increments (open circles) are shown. A linear fit to the independent data (filled circles) after the anomalous value centered on 1971.5 is removed yields a rate of rotation of 3.02° ± 0.43° year^{-1}; this fit is shown with the solid line. The dashed line is the fit if the anomalous value is also used. (**B**) Similar to (A) but for latitude. The solid line is the best straight-line least-squares fit with the anomalous point removed. The dashed line is the best-fit sinusoidal curve [$90° - \zeta = A + B(\tau - 1964) + C \cos \omega(\tau - 1964) + D \sin \omega(\tau - 1964)$] to the data, where A, B, C, and D are constants. The best period is found to be 8.39 years, and it produces ~80% variance reduction. However, this may be an artifact of smoothing, and we do not attach significance to the observed latitude variations.

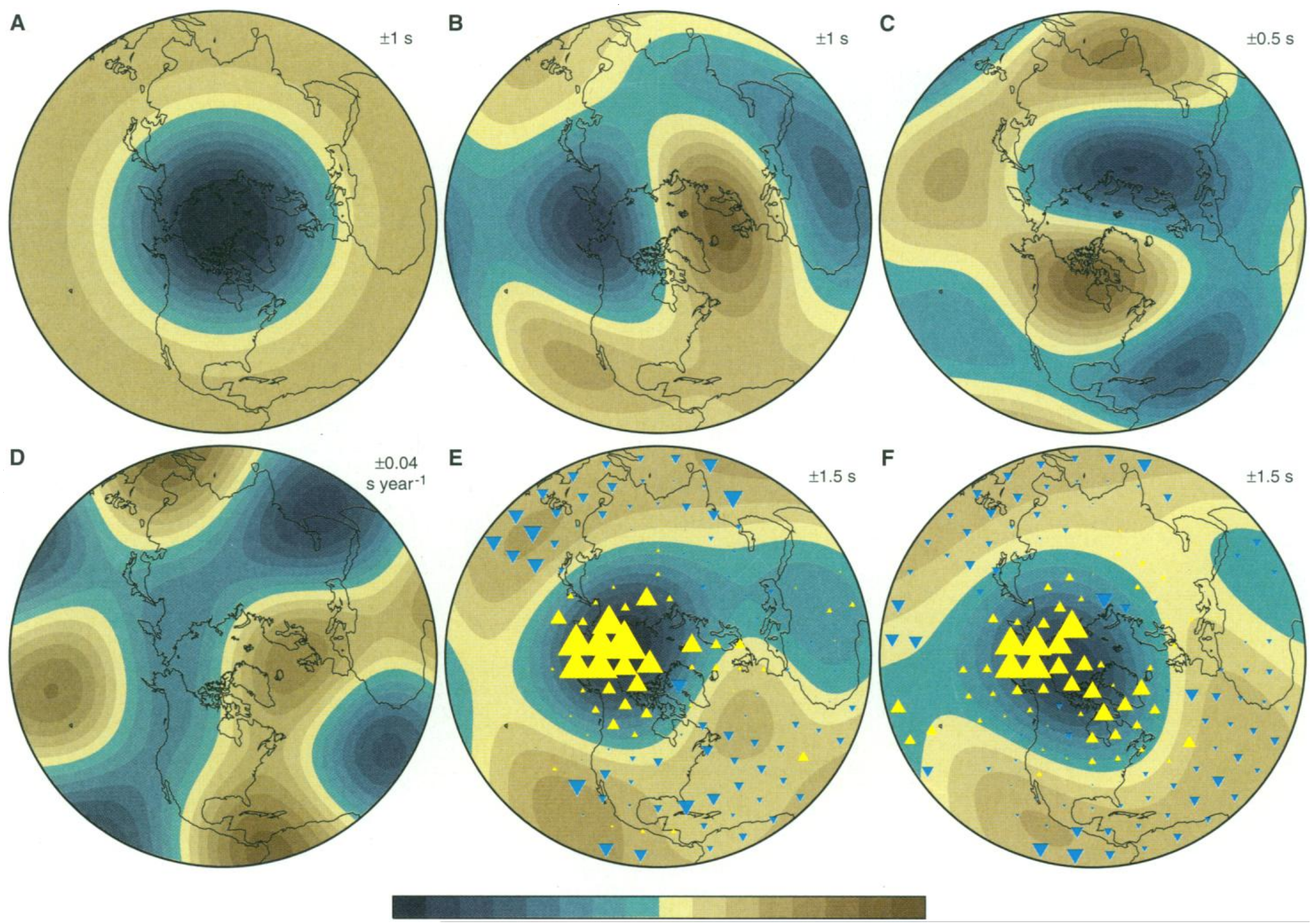

Fig. 4. Spherical harmonic decomposition and synthesis of the observed residual field. (**A**) The axially symmetric part, whose rotation cannot be detected. (**B**) The nonrotating nonaxial part. (**C**) The rotating nonaxial part shown for the year 1964. The rate of rotation is 3.28° year^{-1}; by the current year, 1996, the pattern should have rotated counterclockwise by ~100°. (**D**) Time derivative of the function shown in (C). (**E** and **F**) Residuals synthesized using Eq. 1 for the years 1966.5 and 1990.5; compare with Fig. 2, A and F, respectively. The triangles are the same CAS residuals as in Fig. 2. Scales are indicated at the top right corner of each panel.

detail. We find that the second sum on the right side of Eq. 1 is required to explain the data (22), which implies that the nonaxial time-invariant anomaly (Fig. 4B) is about twice as large as the time-variable term (Fig. 4C) and resembles some aspects of the pattern in the CAS stacks (Fig. 2). One possibility is that we have made an incomplete correction for the heterogeneity of the mantle. However, the same effect is independently evident in Song and Richards' work [figure 4 of (2)], in which the effect of mantle anomalies should be nearly totally eliminated because they considered only differences in travel times. Yet the systematic residuals that they found relative to the prediction of the radially uniform model of transverse anisotropy are larger than the time variation itself, agreeing in sign and approximately in magnitude with the values in our Fig. 4B (25). Clearly, the origin of this time-invariant part of the solution requires further study. If it is real, the assumption that the seismic velocities themselves, or the local orientation of the axis of symmetry, are constant in time may be invalid.

If we assume that the temporal variation of the residuals is attributable to rigid rotation, our results yield a differential rotation of $\sim 3° \text{ year}^{-1}$ over the past three decades, implying that the inner core rotates in a period shorter than that of the mantle by ~2 s per day. How could the inner core be rotating faster than Earth's surface? At least two classes of explanation suggest themselves, emphasizing either mechanical or electromagnetic force balances within the core, and they lead to mutually compatible conclusions. The first recognizes that the rotation of the mantle is slowing down with time, as a result of external tidal forces (26). If the known secular increase in the length of day, ~2 ms per century (27), is extrapolated backward, the inner core can be construed to be rotating with the same period the mantle had $\sim 10^5$ years ago (28). Leaving out the electromagnetic forces (20, 29), a lag time of 10^5 years implies a viscosity $\eta \sim 10^{-4}$ Pa·s (30).

Such a low value of η is consistent with the conclusion, on the basis of previous estimates, that viscous forces are likely to be of relatively minor importance in determining the orientation of the inner core (29, 31, 32). This leads us to focus on the effects of electromagnetic forces. The dynamo calculations of Glatzmaier and Roberts (31) predict a differential rotation between the inner core and mantle of the sign and magnitude inferred from the seismological data (33). The coincidence in time between the geomagnetic jerk of 1969–1970 and the rapid change in inner-core orientation that we find around 1970 (Fig. 3A) is itself suggestive of inner-core rotation being influenced by the magnetic field.

The eastward differential rotation of the inner core is roughly one order of magnitude faster than the classical westward drift of the nondipole geomagnetic field (20). This is of interest because the magnetic field lines are expected to diffuse into the inner core (34, 35). The consequence of the differential rotation, were it to continue unhindered, would therefore be to wind the field lines around the rotation axis roughly every century. Clearly, any winding of this kind would result in an increased electromagnetic coupling with the inner core, which would no doubt influence the differential rotation, the geomagnetic field, or both, over time scales of centuries to millennia. The possible correlation between an anomaly in the orientation of the inner core and the geomagnetic jerk of 1969–1970 suggests the importance of establishing a systematic observational program to monitor the motions of the inner core. By analogy with modern observations of global weather by satellites, seismology can serve to monitor the "meteorology" of Earth's outer core and geodynamo (36). The analogy is particularly appropriate in that the geomagnetic field observed at Earth's surface represents only the radial part of the field emerging out of the top of the core, reflecting flow just below the top of the outer core (29). In contrast, the orientation of the inner core reflects conditions at the bottom of the outer core, so that the prospect now exists of being able to monitor both the top and bottom boundary conditions for the region within which Earth's magnetic field is created.

REFERENCES AND NOTES

1. I. Lehmann, *Bur. Cent. Seismol. Int.* **A14**, 3 (1936).
2. X.-D. Song and P. G. Richards, *Nature* **382**, 221 (1996).
3. G. Poupinet, R. Pillet, A. Souriau, *ibid.* **305**, 204 (1983).
4. A. Morelli, A. M. Dziewonski, J. H. Woodhouse, *Geophys. Res. Lett.* **13**, 1545 (1986).
5. J. H. Woodhouse, D. Giardini, X.-D. Li, *ibid.*, p. 1549.
6. P. M. Shearer, K. M. Toy, J. A. Orcutt, *Nature* **333**, 228 (1988).
7. P. M. Shearer, *J. Geophys. Res.* **99**, 19647 (1994).
8. ——— and K. M. Toy, *ibid.* **96**, 2233 (1991).
9. K. C. Creager, *Nature* **356**, 309 (1992).
10. J. Tromp, *ibid.* **366**, 678 (1993).
11. X.-D. Song and D. V. Helmberger, *Geophys. Res. Lett.* **20**, 2591 (1993).
12. ———, *J. Geophys. Res.* **100**, 9805 (1995).
13. L. Vinnik, B. Romanowicz, L. Berger, *Geophys. Res. Lett.* **21**, 1671 (1994).
14. W.-J. Su and A. M. Dziewonski [*J. Geophys. Res.* **100**, 9831 (1995)] used the CAS method to solve the inverse problem of determining the orientation of the axis of symmetry for the inner core in a manner similar to that proposed by Shearer and Toy (8). If a body is characterized by transverse anisotropy in velocity, then the pattern of travel-time residuals obtained by summing the anomalies for all rays bottoming at the same depth and parallel to a line through the center of Earth having the colatitude ζ (with respect to the axis of rotation) and longitude λ (with respect to the Greenwich meridian) should show axial symmetry with respect to the coordinates ζ_0 and λ_0. Polar-plane and equatorial-plane sections of a sphere are shown in Fig. 1, C and D, respectively; all rays passing through the shaded region are used to form the CAS average for the ray angles (ζ, λ). More specifically, the global pattern of residuals should be described by the spherical harmonics Y_2^0 and Y_4^0, appropriate for transverse anisotropy, in a coordinate system with the pole at (ζ_0, λ_0). Thus, the axis of symmetry can be determined by finding the location of the pole that yields the greatest variance reduction by a least-squares fit of the residuals to these two harmonics. The minimum variance, σ^2, of n CAS averages, δt_i, is found by means of a parameter search with respect to the zonal spherical harmonics Y_2^0 and Y_4^0 in a coordinate system with a pole at ζ_0 and λ_0
$$\sigma^2 = \sum [\delta t_i - A_2 Y_2^0(\vartheta_i - \xi_0) - A_4 Y_4^0(\vartheta_i - \xi_0)]^2 \quad (2)$$
where ϑ_i and φ_i are the coordinates of the ith CAS average. The CAS approach enhances the effect of anisotropy with respect to, for example, random reading errors or any signal that might result from lateral heterogeneity in the inner core. The *PKIKP* or DF branch, which defines the rays entering the inner core (Fig. 1A), is used to measure the inner-core anisotropy. Because of the interference of travel times belonging to different branches of the core phases (Fig. 1, A and B), the DF branch arrivals cannot be reliably measured in the full distance range of interest. At large distances (approaching 180°), there are few measurements because of the small area at Earth's surface where observations can be made.
15. *Bulletin of the International Seismological Centre: Catalogue of Events and Associated Observations, vols. 1–29* (years 1964–1992) (International Seismological Centre, Thatcham, Berkshire, England, 1967–1994).
16. R. Jeanloz and H. R. Wenk, *Geophys. Res. Lett.* **15**, 72 (1988).
17. S. Karato, *Science* **262**, 1708 (1993).
18. L. Stixrude and R. E. Cohen, *ibid.* **267**, 1972 (1995).
19. V. Courtillot, J. Ducruix, J. L. Le Mouël, *C. R. Acad. Sci. Paris Ser. D* **287**, 1095 (1978).
20. R. T. Merrill and M. W. McElhinny, *The Earth's Magnetic Field* (Academic Press, Orlando, FL, 1983).
21. The open circles in Fig. 3B show an oscillation with a period of ~8 years, and perhaps a linear trend in latitude. If this were significant, it would correspond to a wobble of the axis of symmetry with an amplitude of ~4.5°. A nonlinear least-squares fit yields a period of 8.39 years and a linear slope of $-0.13° \text{ year}^{-1}$. However, the appearance of periodic variations may be an artifact of smoothing data that happen to form a high-low pattern. A straight-line fit to the data represented by filled circles, but with the second point removed, yields a slope of $-0.28 \pm 0.12° \text{ year}^{-1}$.
22. If the only temporal variation of inner-core seismic structure is attributable to differential rotation with respect to Earth's surface or mantle, then the travel-time residuals should follow the equation
$$\delta t(\vartheta, \varphi, \tau) = \sum_{\ell, m} \{A_{\ell m} \cos[m(\varphi - \omega\tau)] + B_{\ell m} \sin[m(\varphi - \omega\tau)]\} p_{\ell m}(\cos \vartheta) \quad (3)$$
where τ is time (in years since 1964.0) and $p_{\ell m}$ are fully normalized associated Legendre polynomials. We formulate a nonlinear least-squares problem in which we find the coefficients $A_{\ell m}$ and $B_{\ell m}$ and the frequency ω for which the best variance reduction is achieved. Using the coefficients derived from the data shown in Fig. 2, we find that this formulation is not appropriate: The maximum variance reduction is obtained for $\omega = 0$. This means that, in addition to the variable part (only $m \neq 0$ terms are involved),

there must be also be a part of the rotation (off-)axis velocity pattern that does not depend on time.

23. B. Romanowicz, X.-D. Li, J. Durek, *Science* **274**, 963 (1996).
24. Song and Richards (*2*) relied on high-quality observations, but these were relatively few in number and they investigated only three source-receiver paths to obtain the time variations. In comparison, we used noisier ISC data, but because each CAS value was derived on average from 230 observations, the standard error of the mean should be ~0.1 s. The spatial uniformity of the CAS averages in Fig. 3 testifies to their accuracy. Because of the dense data coverage, we can simultaneously determine both the spatial and temporal variations of the residuals, inverting for both the magnitude and orientation of the anisotropy as a function of time. The analysis in (*2*) required Song and Richards to assume both the geometry (constant model of anisotropy and tilt of the symmetry axis) and process (rigid rotation about an axis coinciding with the mantle rotation axis) to interpret their observations. Still, data for the second path (Kermadec to Kongsberg) give time variations inconsistent with the model. The factor of 3 difference between the rotation rates obtained in (*2*) and in the present study is at least partly attributable to the model of anisotropy used by Song and Richards for the inner core. The magnitude of anisotropy they assumed is about twice that of model IAC4B (*14*), for example, and had they used this model, they would have obtained a rotation rate of ~2° year^{-1}. We inverted for the magnitude of anisotropy in the present study.
25. The time-dependent solution (Fig. 4C) is dominated by the harmonics Y_2^1 and Y_4^1, whose sum has a maximum at ~60°N. Thus, observations at corresponding values of the angle ζ, the angle that a ray forms with the axis of symmetry, should be the best for monitoring the rotation of the inner core. This is confirmed in Fig. 4D, which shows the time derivative of Fig. 4C (the relatively large effects predicted for low latitudes may be contaminated by noise, because the CAS coverage below 30° has had large gaps during the early time intervals; Fig. 2). The predicted solutions for the earliest and latest time periods compare well with the data (Fig. 4, E and F) as well as with the spherical-harmonic expansion for the same time periods (Fig. 2, A and F).
26. K. Lambeck, *The Earth's Variable Rotation: Geophysical Causes and Consequences* (Cambridge Univ. Press, New York, 1980); *Geophysical Geodesy: The Slow Deformations of the Earth* (Oxford Univ. Press, New York, 1988).
27. This angular acceleration value, -5×10^{-22} rad s^{-2}, is an average for the past 2 millennia and has varied by <50% over this time period (*26*). Because of the need for a large backward extrapolation in time, however, its uncertainty for the present application is difficult to evaluate.
28. A naïve analysis would attempt to relate the lag time $\tau = 10^5$ years (3×10^{12} s) between the inner core and mantle with the viscous coupling time $t_\nu = d^2/\nu$, where d is the thickness (2.25×10^6 m) and ν is the kinematic viscosity of the outer core. The result, $\nu \sim 2$ m^2 s^{-1} [viscosity $\eta = \nu\rho \sim 2 \times 10^4$ Pa·s, using a density of $\rho = 1.1 \times 10^4$ kg m^{-3}; A. M. Dziewonski and D. L. Anderson, *Phys. Earth Planet. Inter.* **25**, 297 (1981)], is orders of magnitude greater than current estimates for the viscosity of the outer core (*37–39*). This approach, however, disregards the essentially two-dimensional character of rotating flows (as expressed by the Taylor-Proudman theorem); momentum need only be transmitted across a thin (Ekman) boundary layer rather than the full thickness of the fluid layer (*29, 40, 41*). The characteristic time for momentum transfer is thereby shortened by a factor of $E^{1/2}$, where the Ekman number $E = \nu/d^2\Omega \sim 3 \times 10^{-16}$, yielding a rotation rate $\Omega = 7.3 \times 10^{-5}$ rad s^{-1} (compare D. L. Book and J. A. Valdivia, *J. Plasma Phys.*, in press).
29. D. Gubbins and P. H. Roberts, in *Geomagnetism*, J. A. Jacobs, Ed. (Academic Press, Orlando, FL, 1987), vol. 2, pp. 1–23.
30. Equating the lag time $\tau \approx 10^5$ years (3×10^{12} s) with the "spin-up" time $t_E = d/(\nu\Omega)^{1/2}$, where d is the thickness of the outer core and Ω is Earth's rotation rate, yields a value for the kinematic viscosity $\nu \sim 10^{-8}$ m^2 s^{-1} for the outer-core fluid [viscosity $\eta = \nu\rho \sim 10^{-4}$ Pa·s (*29, 40–42*)]. Taking into account the Lorentz forces associated with the magnetic field leads to a magnetic spin-up time $\tau_M = \tau_E/(\alpha\sqrt{2})$ in the case of strong fields ($\alpha > 2$) (*42*). Here, the magnetic interaction parameter or Elsasser number, $\alpha^2 = B^2\sigma/(2\rho\Omega) \approx 3.7 \times 10^5 B^2$ in SI units (where B is the magnetic field) gives the ratio of Lorentz to Coriolis forces (*20, 29*) [the electrical conductivity of the outer-core fluid, $\sigma = 6\ (\pm 3) \times 10^5$ S m^{-1}, is relatively well known from high-pressure experiments (*38*)]. Because the total strength of the magnetic field deep inside the core is not well known, the magnitude of the Elsasser number α^2 is also uncertain. Downward extrapolation of the present-day field observed at Earth's surface yields a value of $B \sim 10^{-3}$ T for the radial inductance at the base of the mantle, but it is known that this is only part of the field existing deep inside the core; various theoretical estimates bound the characteristic value of the field inside the core between ~10^{-1} and 10^{-4} T (*20, 29, 31*). The Elsasser number is therefore within the range ~10^{-3} to 10^3 (*20, 29*). For the "weak-field" case, which uses the "observed" value $B \sim 10^{-3}$ T, $\alpha^2 \sim 0.4$. If we equate τ_M with the lag time $\tau \approx 10^5$ years identified above for the inner-core rotation, we obtain a value $\nu B^2 \approx 10^{-14}$ T^2 m^2 s^{-1} for the product of kinematic viscosity and squared magnetic field in the outer core. Using the observed magnetic field at the bottom of the mantle (weak-field value $B \sim 10^{-3}$ T) again yields a low value of viscosity, $\nu \sim 10^{-8}$ m^2 s^{-1}, for the outer-core fluid. Indeed, values of kinematic viscosity above the range 10^{-7} to 10^{-6} m^2 s^{-1} imply unacceptably small values for the magnetic field in the core (*20, 29, 31*).
31. G. A. Glatzmaier and P. H. Roberts, *Nature* **377**, 203 (1995); *Phys. Earth Planet. Inter.* **91**, 63 (1995).
32. D. Gubbins, *J. Geophys. Res.* **86**, 11695 (1981).
33. In the Glatzmaier-Roberts model (*31*), the eastward rotation of the inner core relative to the mantle is maintained by magnetic coupling between the solid inner core and the fluid just above it, which is flowing eastward because of a thermal-wind effect (G. A. Glatzmaier, personal communication). As in their original simulation, their latest results show an eastward rotation of the inner core that varies in magnitude between 2° and 3° year^{-1} relative to the mantle.
34. S. R. C. Malin and B. M. Hodder, *Nature* **296**, 726 (1982); M. G. McLeod, *J. Geophys. Res.* **90**, 4597 (1985).
35. The magnetic diffusivity of the core metal, $\eta_m = (\mu_0\sigma)^{-1} \approx 1$ m^2 s^{-1} (where μ_0 is the magnetic permeability of free space and σ is the elastical conductivity of the outer-core fluid), is sufficiently high that field lines diffuse ~6×10^4 m into the inner core within a century (*20, 29, 38*); see also R. Hollerbach and C. A. Jones, *Nature* **365**, 541 (1993).
36. Scaling analysis between the core and atmosphere shows that a few years' seismological monitoring of the inner-core motion corresponds to hourly observations of the weather.
37. C. R. Gwinn, T. A. Herring, I. I. Shapiro, *J. Geophys. Res.* **91**, 4755 (1986).
38. R. Jeanloz, *Annu. Rev. Earth Planet. Sci.* **18**, 357 (1990).
39. E. Wasserman, L. Stixrude, R. E. Cohen, *Eos* **77**, S268 (1996).
40. H. Bondi and R. A. Lyttleton, *Proc. Camb. Philos. Soc.* **44**, 345 (1948).
41. H. P. Greenspan, *The Theory of Rotating Fluids* (Cambridge Univ. Press, New York, 1968).
42. E. R. Benton and A. Clark Jr., *Annu. Rev. Fluid Mech.* **6**, 257 (1974); see also D. E. Loper, *Phys. Fluids* **13**, 2995 (1970); R. G. Davis, *Geophys. Res. Lett.* **21**, 1815 (1994).
43. We thank G. Glatzmaier, D. Loper, M. Manga, P. Olson, P. H. Roberts, B. Romanowicz, L. Stixrude, J. Tromp, and G. Ekström for helpful discussions, as well as X.-D. Song and P. G. Richards for sending us a copy of their paper before publication and for subsequent discussions. This work was initiated while A.M.D. was a visiting professor in the Miller Institute for Basic Research in Science (March 1995) and was supported by NSF and NASA.

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Rotation and Magnetism of Earth's Inner Core

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Three-dimensional numerical simulations of the geodynamo suggest that a super-rotation of Earth's solid inner core relative to the mantle is maintained by magnetic coupling between the inner core and an eastward thermal wind in the fluid outer core. This mechanism, which is analogous to a synchronous motor, also plays a fundamental role in the generation of Earth's magnetic field.

Three-dimensional (3D) numerical simulations of the geodynamo, the mechanism in Earth's core that generates the geomagnetic field, showed that a magnetic field with an intensity, structure, and time dependence similar to that of Earth's can be maintained by a convective model (*1, 2*). The convection, which takes place in the fluid outer core surrounding the solid inner core, twists and shears magnetic field, continually generating new magnetic field to replace that which diffuses away. The model we describe here (*2*) assumes the mass, dimensions, and basic rotation rate of Earth's core, an estimate of the heat flow out of the core, and, as far as possible, realistic material properties (*3*).

Convection is driven by thermal and compositional buoyancy sources that develop at the inner core boundary as the Earth cools and iron alloy solidifies onto the inner core (*4*). For simplicity, we modeled the core as a binary alloy (*5*). Our boundary conditions at the inner core boundary constrain the local fluxes of the latent heat and

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