

Seismological evidence for differential rotation of the Earth's inner core

Xiaodong Song & Paul G. Richards

Lamont-Doherty Earth Observatory, Palisades, New York 10964, USA, and Department of Earth and Environmental Sciences, Columbia University, New York, New York 10027, USA

The travel times of seismic waves that traverse the Earth's inner core show a small but systematic variation over the past three decades. This variation is best explained by a rotation of the inner core that moves the symmetry axis of its known seismic anisotropy. The inferred rotation rate is on the order of 1° per year faster than the daily rotation of the mantle and crust.

THE Earth's liquid iron core was formed very early in the planet's history, after which slow cooling has led to the formation and growth of the solid inner core. The inner core was discovered by seismological methods in 1936¹. Its slow growth to a present radius of about 1,220 km has provided a source of energy as liquid iron freezes at the inner-core boundary. This energy is thought to drive convection in the outer core, which in turn maintains the dynamo action that generates the Earth's magnetic field².

As the inner core lies at the centre of a much larger, liquid outer core (radius 3,480 km) of very low viscosity^{3,4}, it is easy for the inner core to rotate^{5,6}. We present observational evidence for inner-core rotation, at a rate that has taken the inner core through more than a quarter of a complete revolution this century, in an eastward direction. Around the equator of the inner core this rate corresponds to a speed of a few tens of kilometres per year, which is more than 100,000 times faster than the fastest relative motion of the tectonic plates of the lithosphere. Our estimate of the inner-core rotation rate comes from an interpretation of measurements of the difference in travel time between seismic body waves that pass through the inner core, and those that pass only through the outer core. We show that these measurements, made upon seismograms in archives going back to the mid-1960s, have systematically changed with time.

Essential to our interpretation of the travel-time data is the recent discovery that the inner core is anisotropic, with cylindrical symmetry about an axis aligned approximately with the Earth's north-south spin axis. Such anisotropy was first proposed^{7,8} to explain the faster propagation of seismic waves that travel closer to the spin axis⁹ and the anomalous splitting of certain normal modes of the whole Earth that have significant energy in the inner core¹⁰, and has been recently confirmed¹¹⁻¹³ to have a significant amplitude of about 3%. The fact that the inner core is anisotropic is now well established, based on further studies of travel times and normal modes¹⁴⁻²². The cause of the inner-core anisotropy is believed to be preferred orientation of hexagonal close-packed iron (a high-pressure polymorph of iron²³), but how the alignment occurs is still under debate²⁴⁻²⁶. Recent studies^{11,15,18} also indicate that the inner core's axis of anisotropy is tilted by a few degrees from the north-south axis. A rotation about any axis different from the symmetry axis, including the north-south axis, will cause a systematic change in the times of waves transmitted through the inner core from a repeating seismic source, observed at the same fixed station. Moreover, we can use the arrival times of seismic waves transmitted through the outer core as reference points, as the outer core is thought to have negligible fluctuations in seismic velocity associated with its own convection²⁷.

We were led to our search for inner-core motion in part by the work of Glatzmaier and Roberts^{28,29}, who recently reported a three-dimensional, self-consistent numerical model of the geo-

dynamo. Their model has an inner core with a differential rotation that is "as large as a few times 10^{-9} rad s⁻¹ and changes on a timescale of about 500 years"²⁹ and is almost always rotating slightly faster than the mantle and crust. The rate that we report here is comparable with their value, is large enough to be important in many areas of geophysics and geochemistry, and adds a new dimension—time—to traditional seismological studies of the Earth's deep interior structure.

Differential travel-time observations

All of the seismic waves discussed here are simple short-period longitudinal waves (sound waves) with a wavelength of 5–20 km. Such waves that have propagated through the outer core are generically labelled PKP (Fig. 1). There are two possible paths, called PKP(AB) and PKP(BC), for waves that turn at the middle and the bottom of the outer core, respectively. The wave transmitted through both the outer and inner core is called PKP(DF). Our basic measurement is the time difference between the PKP(BC) and PKP(DF) arrivals, often abbreviated as the BC – DF time. Figure 1 shows that these two waves travel very closely together in the crust and mantle and in most of the outer core, so that the differential time (BC – DF) is relatively insensitive to uncertainty in source location and to three-dimensional heterogeneities of the crust and mantle along the ray paths. A standard Earth model (the Preliminary Earth Reference Model, or PREM³⁰) is used to form the BC – DF residual (that is, the observed time difference minus the time difference predicted for PREM) for purposes of subsequent analysis. In addition, residuals of the differential times between PKP(AB) and PKP(BC), which pass through the outer core only, are used to examine possible systematic biases in source locations. Residuals of the differential times between PKP(AB) and PKP(DF) are also useful when PKP(BC) is not observable.

We began to explore time-variability in PKP(DF) times using north-south paths from nuclear explosions at Novaya Zemlya

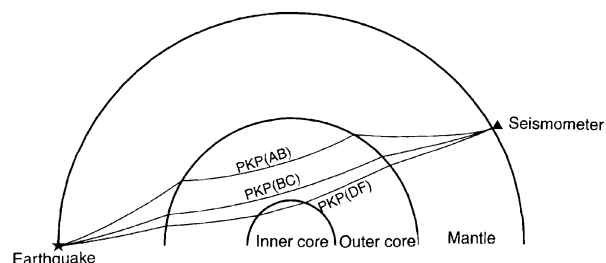


FIG. 1 Ray paths for seismic waves that pass through the Earth's cores.

(which have well constrained locations at around 73° N, 54° E) to seismographic stations in Antarctica. Direct visual comparisons of AB – DF times in the analogue seismograms from different explosions show a decrease by about 0.2 s after about a decade—a difference not much greater than the resolution of the direct visual comparison. To improve the precision of measurement significantly will require scanning and digitization of the analogue data, ideally using the original paper records instead of microfilms (an effort that we have initiated).

We therefore turned to stations with significantly long histories of digital recordings, which permit even more precise measurement of travel times than digitized analogue data. One such station is College, Alaska (station code COL), which has 14 years of digital recordings to date (1982–96), in addition to 18 years of World Wide Standardized Seismographic Network (WWSSN) analogue recordings (1964–82). In our subsequent analysis, the data for the 1960s and 1970s are digitized from the corresponding analogue records. The north–south paths from earthquakes in the South Sandwich Islands (between South America and Antarctica) to COL had provided critical observations in previous anisotropy studies^{11,12}. Figure 2 shows an overlay of PKP waveforms from such paths for two earthquakes that occurred about 15 years apart in almost exactly the same location. The time differences between BC and AB are almost exactly the same, confirming that the two event locations are indeed very close. The DF signal for the 1982 record, however, arrives earlier by about 0.4 s relative to BC than the DF signal for the 1967 record.

We use records such as those shown in Fig. 2 to measure BC – DF and AB – BC differential times by cross-correlation techniques. The BC – DF and AB – BC residuals obtained for the South Sandwich Islands to COL paths for a period of about 28 years (1967–95) are divided into three groups, about 12 years apart (Fig. 3). It can be seen that the BC – DF residuals increase from the 1967–75 period, to the 1980–85 period, and to the 1992–95 period (Fig. 3a), the increase amounting to about 0.3 s over the 28-year period. Possible explanations include a change in the DF (or BC) travel times, or an effect from systematic mislocation of the events. An increase in source–station distance by only 50 km would change the BC – DF time at distances near 151° by 0.3 s, so the time change could potentially be an artefact of the use of different global networks, used to locate the earthquakes in different decades. However, the distribution of the AB – BC residuals (Fig. 3b) for those records which have BC – DF measurements does not appear to have a systematic change with time, suggesting that the observed variation in AB – DF residuals is unlikely to be due to systematic event mislocations or recording instrument biases.

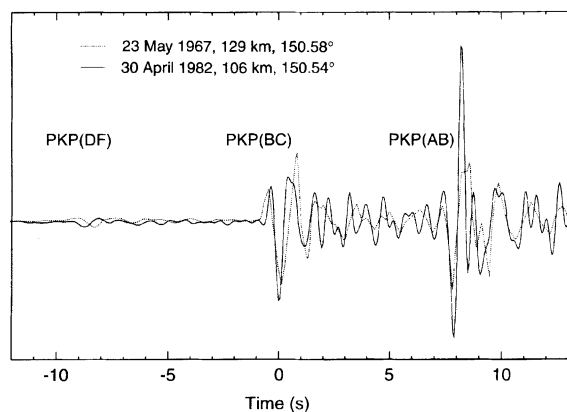


FIG. 2 An overlay of two short-period PKP seismograms from earthquakes that occurred 15 years apart at almost the same location in South Sandwich Islands, as recorded at College, Alaska (COL). Event locations in this study are taken from the Earthquake Data Reports of the US Geological Survey.

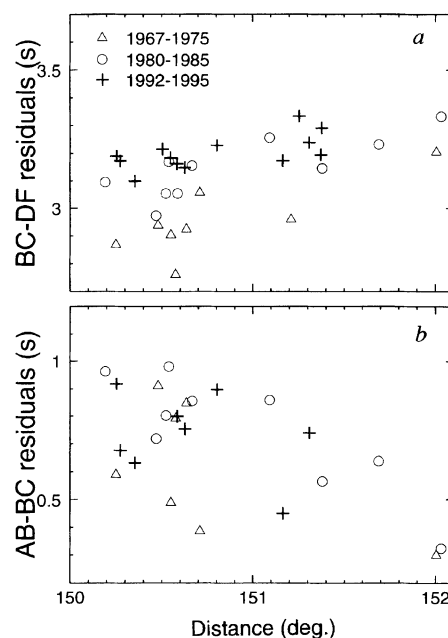


FIG. 3 Residuals for BC – DF (panel a) and AB – BC (panel b) times obtained for earthquakes grouped in three time intervals in the same region (South Sandwich Islands), observed on north–south paths at the same station (COL). Distance is measured by the angle subtended at the Earth's centre by the earthquake source and seismometer locations. The original broadband data for the 1990s are converted to the same short-period instrument response for the earlier years before the measurements. The BC – DF residuals typically increase with time. No such pattern is apparent for the AB – BC residuals, indicating that systematic event mislocations are unlikely to be the cause of time-dependent effects on BC – DF.

Figure 4 shows BC – DF residuals for three event–station pairs that are chosen to have ray directions in the inner core that make very different angles to the equatorial plane. The residuals for different distances and focal depths are corrected to a standard distance and depth by dividing the observed BC – DF residual by the time that the DF ray spends in the inner core for the distance and focal depth concerned, and multiplying by the time the DF ray spends in the inner core for a source at 151° and depth of 0 km. The residuals for South Sandwich Islands to COL paths are progressively larger from the 1960s/1970s to the 1990s, as seen before. But the paths from Kermadec to Kongsberg (KONO) and Bergen (BER), Norway, show smaller residuals for the 1990s than the 1980s. The paths from Tonga to Graefenberg, Germany (GRFO) show no such systematic differences in the data between the 1980s and 1990s. The ray directions in the inner core for the Tonga–GRFO paths are so much closer to the equatorial plane that no effect of inner core anisotropy would be expected. The deviation of the observations at COL and KONO/BER from the time-independent model (dashed curve in Fig. 4) reflects a departure of the observations from a simple constant cylindrical anisotropy of the inner core with the symmetry axis parallel to the north–south axis. We conclude from Figs 3 and 4 that there is the need to consider the effects of more complex anisotropic structure, including a time dependence.

Figure 5 summarizes our observations at COL as a function of the calendar time at which the South Sandwich earthquakes occurred—and also interprets them with a time-dependent model. We observe a gradual increase in the residuals from early 1967 to early 1995, suggesting that the DF ray during this period was taking a progressively faster path through the inner core. Such faster paths can be achieved by a mechanism that decreases the angle between the path within the inner core and the axis of anisotropic symmetry. As the ray path is essentially fixed, we propose that the symmetry axis is moving, owing to a rotation

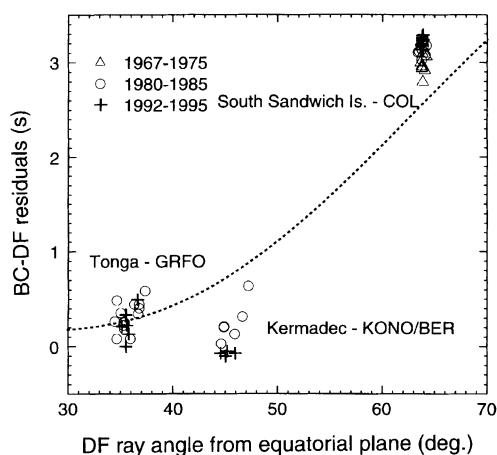


FIG. 4 Residuals for BC – DF times as a function of the angle between the inner-core part of the ray path and the equatorial plane. Ray paths to three different seismographic stations are used (COL—College, Alaska; GRFO—Graefenberg, Germany; KONO/BER—Kongsberg and Bergen, Norway), each using a separate set of earthquakes in the same source region (Tonga to GRFO; Kermadec to KONO/BER; South Sandwich Islands to COL). The dashed line is the prediction for the anisotropy model of Song and Helmberger¹², which has the axis of symmetry of the inner core aligned with the north–south axis. Systematic shifts in residuals over time are absent from the near-equatorial path, and are in opposite directions for the other two paths. Such systematic shifts indicate the need to consider time-dependent effects on the fixed paths through the inner core.

of the inner core. The dashed line in Fig. 5 shows our best estimate of the inner-core rotation required to explain the observed change of residuals over time, as discussed below. In this estimate, the inner core rotates about the north–south axis like the rest of the planet, but at a rate that is 1.1° per year faster. Because the symmetry axis of the anisotropy is not aligned north–south, it moves to the east as illustrated in Fig. 6.

Inferred inner-core motions

We have presented evidence that the inner core is moving, presumably by some type of rotation. But with the limited number of observations made so far, we can get only an approximate estimate of the rotation rate and orientation.

The inner core's moment of inertia, although only 0.07% of the Earth's total moment of inertia, is about 500 times larger than the moment of inertia of the atmosphere, whose internal motions can effect small changes in the Earth's length of day³¹. Observed changes in length of day, over a timescale of years, have been adequately explained without postulating transfer of angular momentum to or from the inner core. We therefore assume that the rotation axis and the rotation rate of the inner core do not change on the timescale of our observations.

At least eight parameters are needed to describe an appropriate model of the time variations in travel-time residuals: the latitude and longitude of the axis of symmetry at a particular time, a direction-dependent relative velocity perturbation from PREM in the form $a + b \cos^2 \xi + c \cos^4 \xi$ (where ξ is the angle between the ray direction in the inner core and the anisotropy symmetry axis), the latitude and longitude of the fixed rotation axis, and finally the fixed rotation rate.

We carried out a grid search for the latitude and longitude of the anisotropy symmetry axis to obtain values of a , b and c and that best fitted observed travel-time residuals without rotation. Using the BC – DF measurements of Song and Helmberger¹² and the new measurements of this study with a combined time span of the past 30 years, the time-invariant best fit to the anisotropy is given by

$$\Delta v/v = 0.0042 - 0.0256 \cos^2 \xi + 0.0517 \cos^4 \xi \quad (1)$$

with symmetry axis crossing the Northern Hemisphere at 82.0° N,

175.0° E. This best fit achieves a variance reduction from the original measurement (with respect to the isotropic PREM) of 93%, not significantly better than the variance reduction of 89% achieved without allowing for a tilt of the symmetry axis. Nevertheless, our estimate of the latitude and longitude of the axis is remarkably close to previous best-fitting symmetry axes. It is only 3.5° away from 79.5° N, 160° E of Su and Dziewonski¹⁸, estimated using PKP(DF) times of the International Seismological Centre Bulletins for years 1964–90; and 8.5° away from 80° N, 120° E of McSweeney *et al.*¹⁹, estimated using differential PKP times for events in the 1980s and 1990s, as in this study. Note the range of these latitude estimates is less than 3° .

The BC – DF measurements can be used to place a minimum bound on the rotation rate, even when the orientation of the symmetry axis of anisotropy is unknown. We computed this bound by calculating the rotation rate needed to match the 0.31-s BC – DF time difference over 28 years at COL (that is, the trend of the data in Fig. 5) for every possible orientation in a grid search, and then selected the smallest. The parameters of anisotropy obtained via the time-invariant analysis were used for this calculation. When the orientation of the rotation axis is also considered unknown, the minimum bound is 0.15° per year. The minimum bound increases if we assume that the differential rotation of the inner core is around the north–south axis. With this rotation axis, the minimum rate increases as the tilt of the symmetry axis decreases, from 0.45° per year for a 15° tilt to 1.30° per year for a 5° tilt. If the BC – DF time differences at KONO are used, the minimum rates are even higher. However, the trend is less well constrained owing to fewer samples.

If we limit the tilt of the symmetry axis to between 8° and 11° away from the north–south axis, as indicated by the above time-invariant analysis, but allow the longitudinal location to vary, the least-squares solution for rotation about the north–south axis to fit the time-varying residuals at COL and KONO puts the symmetry axis at 79.2° N, 169° E at the beginning of 1996 with a rotation rate of 1.1° per year. We regard this as our best estimate, and it gives a variance reduction for COL data compared to the best-fitting time-invariant model (that is without inner-core rotation) amounting to 64%. Assuming the tilt of the symmetry axis in the best-fitting model, the allowed minimum and maximum rotation rates are estimated to be 0.4° per year and 1.8° per year, respectively, from the minimum and maximum slopes of the COL residuals (Fig. 5) constrained by the two-standard-deviation error bars. The variance reduction for the KONO data is insignificant but note that they cover only 13 years (instead of 28 years at COL), and the earlier data are quite scattered. Nevertheless, our pre-

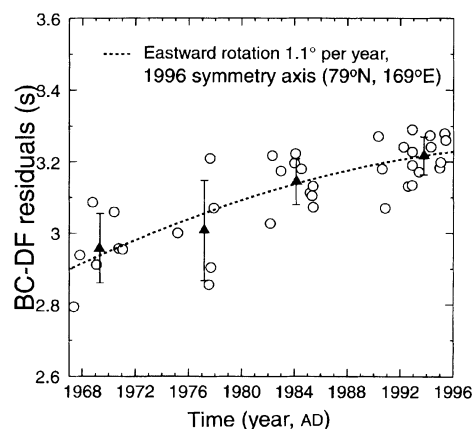


FIG. 5 Residuals for BC – DF times for South Sandwich Islands to COL paths as a function of earthquake time of occurrence. The observed residuals are indicated by open circles. The mean (filled triangles) and \pm one standard deviation (error bar) are shown for the residuals over about four-year periods. The predicted times (dashed curve) are based on our best estimate of inner-core rotation about a north–south axis.

ferred model (Fig. 6) of inner-core motion has its symmetry axis moving away from the Kermadec–Norway paths, consistent with the observed decrease in BC – DF residuals for these paths (Fig. 4). The fact that the preferred model does not significantly improve the fit to the KONO data suggests a need for more data samples gathered for the Kermadec–Norway paths and other paths that are sensitive to the motion of the anisotropic symmetry axis.

The main assumption underlying our best estimate of the rotation rate, namely that the rotation axis is north–south, will be examined when additional data are brought to bear. However, we can already be fairly sure from seismological observations that the rotation axis is not likely to be near the Equator, as such a rotation would cause highly nonlinear changes of travel times for north–south ray paths when the symmetry axis is rotated towards the Equator. Such changes are not seen in the shifts of travel times at COL in Fig. 5. Also in support of an axis far from the Equator is the fact that going further back, to the mid-1960s, it is still true (as today) that only paths close to the north–south spin axis show fast anomalies^{12,16,18,21}.

Implications of detectable inner-core motions

Much tighter constraints on the rate of inner-core rotation and on the orientation of the rotation axes and symmetry axes will be possible as more recordings of north–south PKP waves are gathered and better event locations are obtained (through, for example, joint hypocentre determination^{32,33}, a simultaneous calculation of all hypocentres for a group of earthquakes, which gives better relative event locations). Systematic shifts in PKP(DF) times are best observed where there is the greatest change in $\Delta v/v$ with respect to ξ in equation (1), that is, $10^\circ \leq \xi \leq 45^\circ$. Also, the effect of inner-core rotation is likely to be amplified for waves that traverse the innermost inner core because the paths within the anisotropy are longer. Travel-time anomalies in such waves, twice as big as the anomalies for South Sandwich Islands to COL paths, have been observed^{17,21}. In addition to the differential travel-time data, it is possible that the absolute travel-time data of Su and Dziewonski¹⁸ and the normal-mode analyses of Tromp^{13,22} can be reinterpreted to include possible inner-core motions. Techniques based on reflected waves may also be applicable, if lateral variability of the inner core boundary resulting in variable reflectance can be found, analogous to medical uses of ultrasound reflections to monitor organs that move inside the human body.

Numerous projects are suggested by our conclusion that inner-core motion is observable and can be quantified. New estimates of viscosity in the outer core may become available that in turn may provide information on melting conditions and temperature through the outer core, and refined estimates of the heat flux across the core–mantle boundary. New constraints may be placed on the dynamo motions underlying the geomagnetic field, leading to refinement of models that can reproduce secular variation of the field and geomagnetic reversals. New questions also arise regarding the distribution of angular momentum throughout the



FIG. 6 A representation of our conclusion that the anisotropic symmetry axis of the inner core is moving owing to inner-core rotation. As the solid core and its symmetry axis rotates, there is a change in the angle between the symmetry axis and the inner-core leg of a ray for a fixed source and fixed station, resulting in the systematic shift that we see in the total time taken to travel along the ray. We show the motion of the symmetry axis since AD 1900 according to our best estimate of a 1.1° per year eastward rotation about the north–south axis.

Earth. For example, does the inner-core rotation axis track the north–south spin axis in the precession of the equinoxes? Does the rotation of the inner core result in observable changes in the gravity field?

The organization and maintenance of seismogram archives will need to be modified in order to facilitate studies of core motions. Older data should no longer be superseded by newer, but instead are vital to the accurate determination of rates of change. In this regard it is unfortunate that important sets of digital and analogue data—some of the best in the world, for our purposes—have deteriorated and/or have been destroyed. Support may be needed to recover data from certain archives that have fallen into disuse, or which were not designed for use by the research community.

Gubbins *et al.*³⁴ suggested that “All discussion of modes of core convection will remain speculative until some direct or indirect observational evidence is found.” We believe that observational evidence is now available. The Earth’s dynamo is in part composed of a rotating inner core. □

Received 15 March; accepted 25 June 1996.

- Lehmann, I. *Publ. Bur. Cent. Assoc. Int. Seismol.* **A14**, 87–115 (1936).
- Braginskii, S. I. *Dokl. Akad. Nauk SSSR* **149**, 8–10 (1963).
- Gans, R. F. *J. geophys. Res.* **77**, 360–366 (1972).
- Poirier, J. P. *Geophys. J.* **92**, 99–105 (1988).
- Steenbeck, M. & Helmig, G. *Geophys. J. R. astr. Soc.* **41**, 237–244 (1975).
- Gubbins, D. J. *Geophys. Res.* **86**, 11695–11699 (1981).
- Morelli, A., Dziewonski, A. M. & Woodhouse, J. H. *Geophys. Res. Lett.* **13**, 1545–1548 (1986).
- Woodhouse, J. H., Giardini, D. & Li, X.-D. *Geophys. Res. Lett.* **13**, 1549–1552 (1986).
- Poupinet, G., Pillet, R. & Souriau, A. *Nature* **305**, 204–206 (1983).
- Masters, G. & Gilbert, F. *Geophys. Res. Lett.* **8**, 569–571 (1981).
- Creager, K. C. *Nature* **356**, 309–314 (1992).
- Song, X. D. & Helmberger, D. V. *Geophys. Res. Lett.* **20**, 2591–2594 (1993).
- Tromp, J. *Nature* **366**, 678–681 (1993).
- Shearer, P. M., Toy, K. M. & Orcutt, J. A. *Nature* **333**, 228–232 (1988).
- Shearer, P. M. & Toy, K. M. *J. geophys. Res.* **96**, 2233–2247 (1991).
- Shearer, P. M. *J. geophys. Res.* **99**, 19647–19659 (1994).
- Vinnik, L., Romanowicz, B. & Breger, L. *Geophys. Res. Lett.* **21**, 1671–1674 (1994).
- Su, W. J. & Dziewonski, A. M. *J. geophys. Res.* **100**, 9831–9852 (1995).
- McSweeney, T. J., Creager, K. C. & Merrill, R. T. *Phys. Earth planet. Inter.* (submitted).
- Song, X. D. & Helmberger, D. V. *J. geophys. Res.* **100**, 9805–9816 (1995).

- Song, X. D. *J. geophys. Res.* (in the press).
- Tromp, J. *GSA Today* **5**(7), 137–151 (1995).
- Takahashi, T. & Bassett, W. A. *Science* **145**, 483–486 (1964).
- Jeanloz, R. & Wenk, H. R. *Geophys. Res. Lett.* **15**, 72–75 (1988).
- Karato, S. *Science* **262**, 1708–1711 (1993).
- Stixrude, L. & Cohen, R. E. *Science* **267**, 1972–1975 (1995).
- Stevenson, D. J. *Geophys. J. R. astr. Soc.* **88**, 311–319 (1987).
- Glatzmaier, G. A. & Roberts, P. H. *Nature* **377**, 203–209 (1995).
- Glatzmaier, G. A. & Roberts, P. H. *Phys. Earth planet. Inter.* **91**, 63–75 (1995).
- Dziewonski, A. M. & Anderson, D. L. *Phys. Earth planet. Inter.* **25**, 297–356 (1981).
- Lambeck, K. *The Earth’s Variable Rotation: Geophysical Causes and Consequences* 275–282 (Cambridge Univ. Press, London, 1980).
- Douglas, A. *Nature* **215**, 47–48 (1967).
- Dewey, J. W. & Algermissen, S. T. *Bull. seism. Soc. Am.* **64**, 1033–1048 (1974).
- Gubbins, D. C., Thomson, C. J. & Whaler, K. A. *Geophys. J. R. astr. Soc.* **68**, 241–251 (1982).

ACKNOWLEDGEMENTS. We thank the IRIS DMC for easy access to data; the US NSF for long-term funding of the WWSN archive at Lamont; J. Dewey and R. Engdahl for help with event relocations; and W. Menke, D. Kent and A. Lerner-Lam for comments on early drafts. This work was supported by the US NSF.

CORRESPONDENCE should be addressed to X.S. (e-mail: xsong@LDEO.columbia.edu).