JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 93, NO. B10, PAGES 11,621-11,630, OCTOBER 10, 1988

# INCLINATION ANOMALIES FROM INDIAN OCEAN SEDIMENTS AND THE POSSIBILITY OF A STANDING NONDIPOLE FIELD

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Abstract. We report magnetic inclinations measured in deep-sea sediments of the equatorial Indian Ocean which record the behavior of a nondipole component of the time-averaged geomagnetic field during the Plio-Pleistocene (0-5 Ma). The magnitude of the nondipole effect recorded in these sediments appears to depend on polarity state, with inclinations showing departures from geocentric axial dipole directions which are small (2°) during normal polarity and larger (5°) during reverse polarity times. The overall nondipole effect observed here is consistent with prior spherical harmonic estimates of the paleomagnetic field; the polarity bias found agrees, in both sense and magnitude, with earlier reports of polarity asymmetry in the low-degree zonal harmonic fields. The presence of this asymmetry supports previous suggestions of the existence of a standing component of the nondipole field which does not invert during reversals of the main field. We explore whether the standing field so indicated may have influenced paleomagnetic directions recorded during polarity transitions at other equatorial sites.

## Introduction

Most applications of paleomagnetism assume that the Earth's magnetic field has, on average, the form of a geocentric axial dipole, yet no rigorous theory is available to show why this must be so. Indeed, the present geomagnetic field is quite complicated, being neither geocentric nor axial nor even fully dipolar, and one might well wonder why the geocentric axial dipole (GAD) hypothesis applies to the time-averaged field. The answer comes, in part, from simple plausibility arguments. Because nondipolar features of the geomagnetic field vary most rapidly in time, they may eventually average to zero. Furthermore, the Earth's symmetry about the rotation axis gives no preferred direction for any dipole offset or tilt. It follows then that the time-averaged magnetic field would be dipolar, geocentric, and axially aligned. This sort of theoretical foundation, however, provides only weak support, and confidence in the GAD hypothesis must be built largely on observation.

Paleomagnetic results generally support the GAD hypothesis [e.g., Opdyke and Henry, 1969]; however, small second-order effects have long been noted which show that the time-averaged field deviates slightly from this simple configuration [Wilson, 1970]. Though these effects are quite small, amounting to only a few degrees, the discrepancy is nevertheless significant, and some modification of the GAD hypothesis is required. Wilson showed that the time-averaged field could be better decribed as the field of an axially offset dipole. More recent analyses have refined his observations using spherical harmonics to describe the long-term nondipole field (NDF). These efforts to study the NDF are motivated both by a practical need to better define the paleomagnetic field and by the desire to constrain models of dynamo processes and core boundary conditions. The general agreement between several different studies argues that the form of the time-averaged field, at least for the past several million years, is now reasonably well determined.

The time-averaged field is found to be largely axisymmetric; the axial quadrupole and octupole terms tend to dominate in global spherical harmonic analyses of paleomagnetic data [e.g., Merrill and McElhinny, 1977; Coupland and Van der Voo, 1980; Livermore et al., 1983]. A convenient device for examining these axial NDF effects is the inclination anomaly [Cox, 1975], defined as the difference between observed inclination and geocentric dipole inclination:

$$\Delta I = I(\text{observed}) - I(\text{dipole})$$

For example, the inclination anomaly function for an axial quadrupole contribution is largely symmetric about the equator, while that for an axial octupole contribution is mostly antisymmetric (Figure 1). The quadrupole effect thus has greatest magnitude at the equator and diminishes toward the poles; the octupole has no effect at the equator, has largest magnitude at mid-latitudes, and opposite sign in opposing hemispheres.

The global analyses have shown that the contributions from the axial quadrupole and octupole terms each amount to a few percent of the axial dipole. Less well known, however, is the behavior of the NDF through time. While some results indicate that the magnitude of the NDF is time varying [Wilson and McElhinny, 1974], other findings suggest that the variation is associated with different polarity states [Wilson, 1972; Merrill and McElhinny, 1977]. The analysis of this polarity dependence, however, is complicated by the possibility of slow changes of the NDF with time. The studies which have examined global field for polarity dependence [Wilson, 1970; Merrill and McElhinny, 1977] rely on paleomagnetic data covering just the past few million years, where plate motions are minimal and where data are relatively abundant. As the bulk of the normal polarity data analyzed may be Brunhes age and thus younger than the corresponding reversed polarity data, an apparent polarity asymmetry might be caused by a slow change in the NDF with time.

In order to resolve this ambiguity, one requires estimates of the NDF that have a greater resolution in time than those available from previous studies of global data (which average over 2 m.y. or more). We believe that deep-sea sediments provide just such a detailed record. Here we report results from deep-sea sediments of the Indian Ocean which we have examined in an opening phase of a broader study of nondipole effects recorded in deep-sea sediments. We have concentrated our work on cores from low latitudes because data from these should be most sensitive to the even-harmonic components of the NDF and least sensitive to many potential sources of experimental error, as will be discussed. Our current focus on the Indian Ocean region grew from a desire to supplement the relatively sparse paleomagnetic data available from this part of the globe.

## Experimental Procedure

Deep-sea sediments have been shown to be a fertile source of paleomagnetic data [Opdyke, 1972; Harrison, 1974] and are particularly appropriate to the study of the time-averaged field, since the slow deposition rates (about 1 cm/kYr) combined



Paper number 7B3037
0148-0227/88/007B-3037$05.00

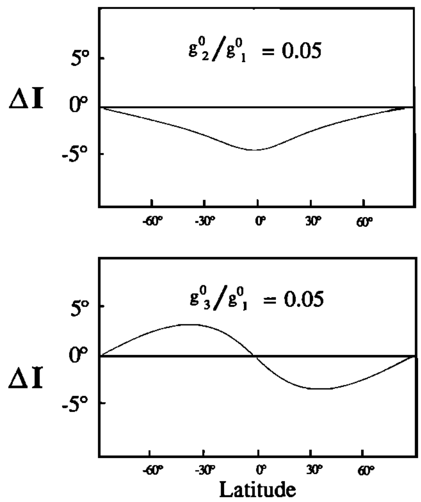


Fig. 1. Comparison of expected inclination anomaly with latitude for low-degree axial fields. (Upper panel) Anomaly for axial quadrupole NDF having a $g0_2/g0_1$ ratio of 0.05. (Lower panel) Anomaly for an axial octupole NDF having a $g0_3/g0_1$ ratio of 0.05.

with the effects of sediment mixing by burrowing benthic organisms suggest that short-term secular variation will tend to be averaged within an individual sample. Moreover, these sediments can be readily dated using a combination of paleontology and magnetic polarity stratigraphy.

The sediments used in this study were taken from the floor of the equatorial Indian Ocean (Figure 2) between 1967 and 1981 using conventional piston coring techniques. With one exception, all sites are within 15° of the equator (locations are listed in Table 1). We initially considered all equatorial Indian Ocean cores from the Lamont collection over 8 m in length which were in good physical condition and for which available biostratigraphic reconnaissance data (D. A. Johnson, personal communication, 1986) indicated pre-Brunhes bottom ages. In addition, we examined three cores from the collection at the Museum Nationale d'Histoire Naturelle [Caulet et al., 1984]. In this analysis we include the results from 23 cores (roughly half of the total number of cores initially sampled) which show stable primary magnetic directions (standard deviations of inclination generally less than 10°) and have magnetic stratigraphies consistent with the biostratigraphic information available. These measurements constitute our primary data set. A second group of six cores, which were examined during previous magnetostratigraphic studies [Opdyke and Glass, 1969; Burckle and Opdyke, 1977], provide supplementary data.

In generating the primary data, we used (on average) some 60 samples taken at intervals of 10 to 50 cm, depending on the length and age of the particular core. Core lengths generally range from about 8 to 18 m; bottom ages range from less than 0.4 Ma to more than 4 Ma (Table 1). As no orientation device was used during coring, absolute declination could not be recovered; only relative declination measurements (with respect to the split face of the core) were possible.

For each core, we subjected at least three, but usually five or more samples to progressive alternating field demagnetization. We chose a suitable demagnetization field level (above which the pilot samples showed a univectorial decay to the origin) to use in blanket treatment of the remaining samples within each particular core. These treatments range from 15 to 40 mT. We took care to avoid orienting all the samples from a core uniformly within the demagnetizing apparatus so that any residual anhysteretic remanent magnetization (ARM) component would not add a consistent bias to our results. Procedures used in acquiring the data in the supplementary set were largely similar, though the magnetometers and demagnetization equipment used during the earlier studies were somewhat inferior, which had necessitated the use of lower demagnetization levels (5 to 15 mT).

An example of the data so obtained is shown in Figure 3. Since this core was stored in 11 sections (of 1.5 m each), there are several declination shifts associated with the numerous physical breaks. These shifts do not record polarity reversals (note the lack of inclination changes at these levels). The cores studied commonly suffer physical breaks or (less obviously) twists and thus can show declination shifts that do not indicate field reversals. We use the presence of 180° declination shifts to confirm the location of polarity reversals indicated by changes of inclination (which are small at equatorial latitudes). The polarity interpretation shown in Figure 3 is supported by a detailed biostratigraphic study of this core [Johnson et al., 1988]. Further examples of paleomagnetic data from cores used for this analysis are also presented in that stratigraphic study.

Knowing chron boundary depths (Table 1) and their associated ages (following the time scale of Berggren et al. [1985]) allowed us to construct simple age models for each core, assuming a constant sedimentation rate between identified boundaries. For analysis of mean inclinations, we excluded those data close to the chron boundaries that showed transitional directions, as well as those that we considered to be part of a subchronozone. For example, we excluded two sample data near 8.5 m in Figure 3 (shown with crosses) because these directions appear to be transitional. Many more points are excluded in this example because they clearly fall within subchronozones. Though some points within the thicker subchronozones may well be representative of the time-averaged field, this was often difficult to determine, and we adopted the more conservative procedure of uniformly excluding all data associated with any of the various Plio-Pleistocene subchrons.

We further removed from the analysis any obviously erratic directions which often (but not always) appeared to be associated with segments of the core prone to physical disturbance. For instance, two samples near 14 m in Figure 3 are so excluded: these points fall close to a core break and likely experienced some physical disturbance. Our editing of such spurious values in the 29 cores removed a total of 72 points, amounting to about 4% of the data.

We analyzed the filtered core inclination data into four groups corresponding to the four most recent polarity chrons: Brunhes (0 - 0.73 Ma), Matuyama (0.73 - 2.47 Ma), Gauss

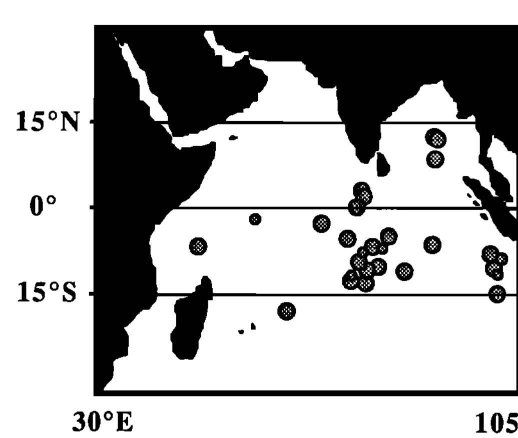


Fig. 2. Indian Ocean core site locations. Larger points indicate newly studied cores (source of primary data); smaller points show previously studied cores (source of supplementary data).

TABLE 1. Magnetic Chron Boundary Depths for the 29 Cores Studied

| Core | North Latitude | East Longitude | Length, cm | Water Depth, m | Boundary Depths, cm BRU-MAT | MAT-GAU | GAU-GIL | Bottom Age, Ma (Upper Limit) |
|---|---|---|---|---|---|---|---|---|
| | | | | *Primary Set* | | | | |
| MD81-369 | -10.05 | 79.80 | 1772 | 5293 | 135 | 855 | 1250 | < 4.40 |
| MD81-375 | -12.78 | 77.77 | 1750 | 5279 | - | 675 | 1185 | < 4.40 |
| RC12-320 | -6.60 | 47.80 | 980 | 4784 | - | - | - | < 0.73 |
| RC12-331 | -2.50 | 69.87 | 846 | 3941 | 643 | - | - | < 1.66 |
| RC12-333 | 0.80 | 76.17 | 1032 | 4233 | - | - | - | < 0.73 |
| RC12-334 | 2.40 | 77.27 | 1013 | 4217 | - | - | - | < 0.73 |
| RC12-339 | 9.13 | 90.03 | 824 | 3010 | - | - | - | < 0.73 |
| RC12-340 | 12.70 | 90.02 | 690 | 3012 | - | - | - | < 0.73 |
| RC12-341 | 13.05 | 89.58 | 1099 | 2988 | - | - | - | < 0.73 |
| RC14-019 | -17.57 | 63.55 | 1620 | 3568 | 691 | - | - | < 1.66 |
| RC14-023 | -9.17 | 76.67 | 1175 | 5376 | 164 | 995 | - | < 3.08 |
| RC14-024 | -6.63 | 79.43 | 1215 | 5183 | 280 | - | - | < 2.47 |
| RC14-046 | -7.82 | 100.00 | 1415 | 5566 | 972 | - | - | < 1.66 |
| VM19-156 | -14.63 | 101.33 | 1204 | 5363 | 981 | - | - | < 0.91 |
| VM19-203 | -9.47 | 43.32 | 1324 | 3651 | - | - | - | < 0.73 |
| VM28-355 | -10.45 | 100.52 | 1248 | 5066 | - | - | - | < 0.73 |
| VM29-030 | 3.08 | 76.25 | 1320 | 3651 | - | - | - | < 0.73 |
| VM29-034 | -5.35 | 74.40 | 1020 | 4762 | 237 | 594 | 830 | < 4.24 |
| VM29-040 | -10.48 | 78.05 | 1788 | 5325 | 348 | 1109 | 1541 | < 3.97 |
| VM29-043 | -12.33 | 75.08 | 1682 | 5150 | 443 | 1085 | 1460 | < 3.97 |
| VM33-054 | -11.02 | 84.68 | 960 | 4907 | 334 | - | - | < 2.47 |
| VM33-055 | -4.73 | 81.70 | 964 | 4891 | 275 | 595 | 840 | < 3.97 |
| VM34-053 | -6.12 | 89.58 | 556 | 3808 | - | - | - | < 0.73 |
| | | | | *Supplementary Set* | | | | |
| RC12-327 | -1.73 | 57.83 | 1598 | 4446 | 629 | 1383 | - | < 2.92 |
| RC14-022 | -11.47 | 75.15 | 1698 | 5276 | 607 | - | - | < 2.47 |
| VM19-153 | -8.85 | 102.12 | 1232 | 5433 | 526 | - | - | < 1.66 |
| VM19-154 | -11.41 | 101.40 | 1951 | 4964 | - | - | - | < 0.73 |
| VM19-171 | -7.07 | 80.77 | 1138 | 5053 | 355 | 695 | 895 | < 4.24 |
| VM29-039 | -7.70 | 77.38 | 1165 | 5082 | 106 | 1052 | - | < 2.92 |

Depth values reported are the average of the sample levels which clearly bracket the reversal. Basal ages are estimated by association with the magnetic reversal nearest the bottom of the core. Cores marked <0.73 contain no reversals; however, a Brunhes age is indicated by the normal polarity inclinations and/or biostratigraphic constraints.

(2.47 - 3.40 Ma), and Gilbert (3.40 - 5.35 Ma). We computed within-core averages for each of the four age/polarity groups when seven or more samples were available for each chron (corresponding to the usual minimum for paleomagnetic sites). In principle, the knowledge of relative declination should allow Fisher averaging; however, the presence of any break or twist would erroneously steepen the Fisher mean. Consequently, we adopted a maximum likelihood estimation technique [McFadden and Reid, 1982; P. L. McFadden, personal communication, 1986] to estimate average inclination using the inclination values alone. Because of the low inclinations recorded in these equatorial sites, the maximum likelihood estimate of inclination is in all cases quite close to a simple arithmetic average. For the core shown in Figure 3, the inclinations estimated with the maximum likelihood method differ from simple arithmetic averages by less than 1/2°. This small difference is typical, and thus the errors associated with this averaging procedure are negligible, being only a fraction of a degree.

## Results

### Introduction

To cast results as inclination anomalies, one must subtract the expected geocentric axial dipole inclination for each site. Here we have determined expected inclination in two ways: in one scheme, we simply used the present core latitude; in an alternative approach, we computed site latitude by compensating for plate motions. For each of the four age/polarity groups, we restored the paleoposition of each core using a hotspot based absolute motion model (AM1-2 of Minster and Jordan [1978]) by projecting sites back for the average age of the data points included. We present inclination anomalies computed with the plate motion correction in Table 2 (inclination anomalies without plate motion correction can be simply computed, using the dipole formula with site latitudes). Note that we invert the sign of reverse polarity inclination anomalies to convert them to normal polarity equivalents. Thus for either polarity, a negative inclination anomaly would, for example, correspond to the observed inclination being shallower than predicted in the northern hemisphere and steeper than predicted in the southern hemisphere.

Both the primary and supplementary sets show a dichotomy of anomaly estimates when the data are averaged by polarity (Table 3). We computed these results by first averaging the two common polarity estimates (Brunhes with Gauss and Matuyama with Gilbert) for each core before averaging between cores. Both primary and supplementary data sets show a larger magnitude anomaly for reversed

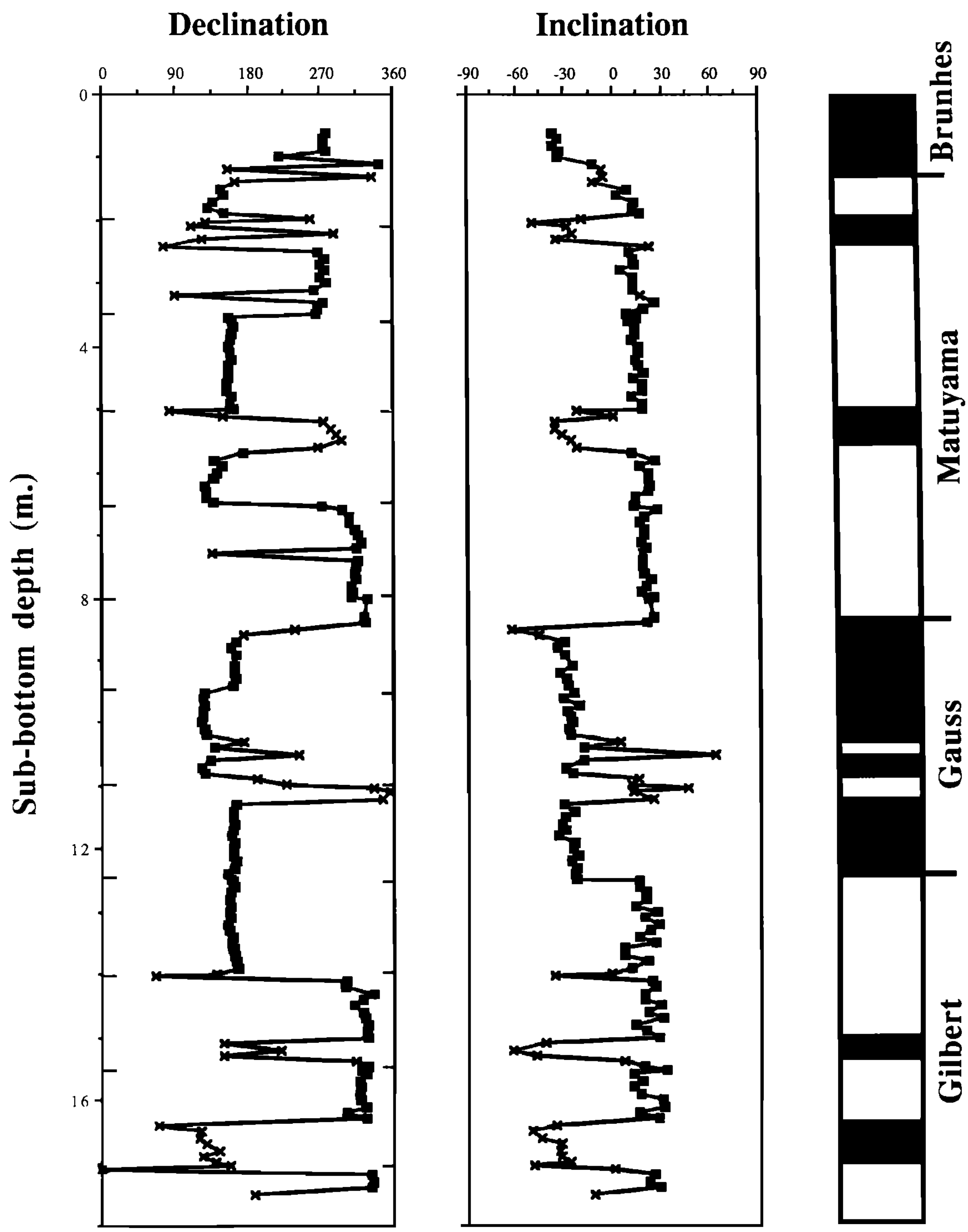


Fig. 3. Paleomagnetic data from core MD81-369. Points marked with crosses are those excluded from the analysis. Declination is given relative to the split face of the core. Inner tick marks on declination panel indicate position of core breaks.

polarity (-5 to -6°) compared with that for normal polarity (-1 to -2°). Since no systematic differences between the primary and supplementary sets are apparent, we carry the distinction no further and treat the results from all 29 cores equally. The combined data yield a normal polarity inclination anomaly of -2.1 ± 0.9° and a reverse polarity anomaly of -5.1 ± 1.5° with plate motions removed (1-sigma limits given). The inclination anomalies without a plate motion correction are somewhat larger: -2.4 ± 0.8° for normal polarity and -6.6 ± 1.5° for reverse.

Because the Indian plate is moving rapidly northward, the plate motion correction reduces the magnitude of the anomaly estimates, though it does not change the difference between normal and reverse substantially. We chose to adopt the more conservative corrected values as our preferred estimates. Having no reason to suspect gross errors in the determination of recent plate motion, we believe these corrected estimates should be more accurate.

Averaging the data by chron gives a greater resolution in time, but the statistical errors associated with the estimates tend

TABLE 2. Inclination and Inclination Anomaly Estimates for the Four Most Recent Chrons

| | Brunhes | | | | Matuyama | | | | Gauss | | | | Gilbert | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Core | N | I | A95 | ΔI | N | I | A95 | ΔI | N | I | A95 | ΔI | N | I | A95 | ΔI |
| *Primary Data* | | | | | | | | | | | | | | | | |
| MD81-369 | - | - | - | - | 58 | 16.5 | 1.6 | 4.4 | 32 | -27.5 | 1.8 | -5.5 | 38 | 20.0 | 2.8 | 2.7 |
| MD81-375 | - | - | - | - | 55 | 23.3 | 2.6 | 2.6 | 38 | -31.1 | 2.1 | -4.5 | 41 | 26.0 | 3.1 | 1.3 |
| RC12-320 | 27 | -9.0 | 4.7 | 4.0 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC12-331 | 50 | -9.6 | 2.1 | -4.3 | 14 | 17.2 | 5.6 | -11.6 | - | - | - | - | - | - | - | - |
| RC12-333 | 20 | 1.8 | 3.9 | 0.4 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC12-334 | 18 | 7.9 | 8.3 | 3.4 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC12-339 | 15 | 21.2 | 7.7 | 3.6 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC12-340 | 10 | 20.8 | 9.3 | -3.5 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC12-341 | 15 | 22.7 | 5.8 | -2.2 | - | - | - | - | - | - | - | - | - | - | - | - |
| RC14-019 | 29 | -40.1 | 2.5 | -7.7 | 24 | 51.4 | 3.2 | -18.8 | - | - | - | - | - | - | - | - |
| RC14-023 | - | - | - | - | 34 | 27.8 | 3.0 | -8.5 | 8 | -29.5 | 8.0 | -9.4 | - | - | - | - |
| RC14-024 | 8 | -15.9 | 3.9 | -2.4 | 37 | 14.1 | 3.7 | 0.4 | - | - | - | - | - | - | - | - |
| RC14-046 | 38 | -25.6 | 2.7 | -9.7 | 13 | 21.8 | 4.9 | -5.5 | - | - | - | - | - | - | - | - |
| VM19-156 | 18 | -29.2 | 2.8 | -1.3 | - | - | - | - | - | - | - | - | - | - | - | - |
| VM19-203 | 23 | -20.2 | 3.2 | -1.7 | - | - | - | - | - | - | - | - | - | - | - | - |
| VM28-355 | 22 | -29.9 | 4.9 | -9.4 | - | - | - | - | - | - | - | - | - | - | - | - |
| VM29-030 | 67 | 7.9 | 4.2 | 2.0 | - | - | - | - | - | - | - | - | - | - | - | - |
| VM29-034 | 11 | -7.0 | 3.6 | 4.0 | 23 | 14.5 | 6.5 | -2.6 | 18 | -2.3 | 4.1 | 10.7 | 9 | 7.9 | 17.9 | 5.7 |
| VM29-040 | 13 | -15.9 | 5.2 | 4.7 | 35 | 32.3 | 2.0 | -10.5 | 21 | -36.5 | 5.1 | -13.8 | 12 | 34.5 | 6.5 | -11.2 |
| VM29-043 | 20 | -32.6 | 5.2 | -8.6 | 45 | 25.0 | 3.1 | -0.1 | 27 | -29.7 | 5.7 | -3.9 | 18 | 25.4 | 4.2 | 0.8 |
| VM33-054 | 12 | -20.4 | 6.2 | 1.3 | 24 | 26.2 | 3.5 | -3.6 | - | - | - | - | - | - | - | - |
| VM33-055 | 8 | -10.3 | 6.1 | -0.5 | - | - | - | - | 7 | -9.1 | 8.4 | 2.9 | - | - | - | - |
| VM34-053 | 24 | -19.5 | 3.9 | -6.9 | - | - | - | - | - | - | - | - | - | - | - | - |
| *Supplementary data* | | | | | | | | | | | | | | | | |
| RC12-327 | 57 | 2.1 | 2.4 | 5.6 | 51 | 6.8 | 4.1 | -2.9 | - | - | - | - | - | - | - | - |
| RC14-022 | 59 | -24.0 | 5.3 | -1.6 | 92 | 29.2 | 3.3 | -6.0 | - | - | - | - | - | - | - | - |
| VM19-153 | 69 | -23.3 | 1.8 | -5.5 | 33 | 21.0 | 4.0 | -2.7 | - | - | - | - | - | - | - | - |
| VM19-154 | 29 | -23.0 | 3.1 | -1.0 | - | - | - | - | - | - | - | - | - | - | - | - |
| VM19-171 | 31 | -16.2 | 4.1 | -1.9 | 27 | 20.6 | 2.7 | -4.9 | 19 | -22.4 | 4.4 | -5.8 | 19 | 31.6 | 5.9 | -14.4 |
| VM29-039 | 11 | -20.1 | 9.8 | -3.8 | 63 | 26.1 | 3.2 | -9.6 | 12 | -8.9 | 5.5 | 8.4 | - | - | - | - |

N, number of samples included in estimate; I, inclination computed with maximum likelihood estimation procedure; A95, alpha-95 radius of confidence based on maximum likelihood estimate of the Fisher precision parameter; ΔI, inclination anomaly calculated by subtracting geocentric axial dipole inclination (computed with a plate motion correction) from the observed inclination. Reversed polarity ΔI values (Matuyama and Gilbert) are inverted to give normal polarity equivalents.

TABLE 3. Comparison of Results Averaged by Polarity From Primary and Supplementary Data Sets

| Data Set | Polarity | N | Latitude, deg, S | ΔI, deg |
|---|---|---|---|---|
| Primary | N | 23 | 5 | -2.4 ± 1.0 |
| | R | 11 | 10 | -4.6 ± 2.1 |
| Supplementary | N | 6 | 8 | -0.6 ± 1.7 |
| | R | 5 | 8 | -6.2 ± 1.5 |
| Combined | N | 29 | 6 | -2.1 ± 0.9 |
| | R | 16 | 9 | -5.1 ± 1.5 |

N is number of cores. Plate motion correction is applied. One sigma errors on inclination anomaly (ΔI) are shown.

to increase (Figure 4). The Brunhes anomaly value of -1.7 ± 0.9° (N=26, mean latitude 5° S) is still relatively well determined, as is that of -5.0 ± 1.5° for the Matuyama (N=16, mean latitude 10° S); the Gauss estimate of -2.2 ± 2.7° (N=9, mean latitude 10° S) and that for the Gilbert of -2.5 ± 3.4° (N=6, mean latitude 11° S), however, are considerably less precise because of the smaller number of cores available with these older sediments.

In considering the statistical significance of these results, we cannot use the usual two-dimensional Fisher statistics, but instead assume a one-dimensional normally distributed population and employ Student's t statistic. The overall normal polarity anomaly is significant at 98% confidence; the overall reversed anomaly is significant at 99% confidence. Furthermore, the normal polarity estimate is significantly different from the reversed polarity value at 90% confidence. In considering the individual chrons, we find that the Brunhes and Matuyama anomalies are significant (at 90% and 99% levels, respectively), but the Gauss and Gilbert anomalies, having much larger errors because fewer cores are available, are not.

Most directly at issue here is whether the differences between adjacent age/polarity intervals follow the polarity dependence indicated by the overall results. In general they do: the normal polarity Brunhes anomaly is significantly smaller (in magnitude) than the reversed polarity Matuyama at 95% confidence. The normal polarity Gauss anomaly is also significantly smaller than the Matuyama but only at a relatively relaxed 80% confidence level. The Gauss and Gilbert averages are quite similar and do not, of course, imply a significant difference.

## Interpretation of Results

The cores studied clearly show inclinations which systematically deviate from expected dipole directions. The inclination anomalies computed are small but significant and consistently negative. Moreover, comparison of normal and reverse polarity averages indicates a statistically significant difference (at 90% confidence) with the magnitude of the inclination anomaly being larger during reversed polarity times. These results are consistent, both in the magnitude of the effect and in the sense of polarity dependence, with previous spherical harmonic analyses of global paleomagnetic data [Merrill and McElhinny, 1977; Coupland and Van der Voo, 1980; Livermore et al., 1983]. However, because the effects are admittedly rather small and, for example, the normal and reverse averages are not uniformly well separated (e.g., Gauss versus Gilbert), it is important to consider alternative mechanisms that could give rise to these anomalies.

There are, of course, many possible error sources associated with each paleomagnetic measurement: slight misorientation of samples, induced ARM during demagnetization, physical disturbance of the sediments during coring, etc. All such errors, however, should have no preferred orientation and thus contribute only random noise to each of the core averages. This assertion is supported by the close agreement of the results from the supplementary data set with those from the primary (i.e., higher quality) set.

There are three sources of error, however, which are potentially more serious, as they might uniformly affect all samples within a single core and hence bias the core mean inclinations. The first of these is the nonvertical penetration of the core barrel. Experience with core orienting devices [McCoy and Von Herzen, 1971; Seyb et al., 1977] shows that the orientation of piston cores may well be several degrees from plumb. Errors from such tilting could range up to the amount of tilt and thus might be of comparable size to the small NDF effects we attempt to measure. A second error may stem from the incomplete removal of a low-coercivity component (either a viscous component or perhaps one associated with the drying process) acquired while the cores were stored. Though this sort of "shelf" remanence is not obvious in any of the cores studied here, it has been observed in other deep-sea cores [Johnson et al., 1975; Witte and Kent, 1988]. The third potential error source is the effect of nearby and highly magnetized oceanic crust. Magnetic anomalies of several thousand nanoteslas are commonly encountered in deep tow magnetometer surveys near ridge axes [e.g., Macdonald, 1977]. Though these are likely to be the largest values attained anywhere on the ocean floor, a moderately large local field may be quite common; one of perhaps 2000-nT intensity could perturb magnetic directions near the equator (where the total intensity is only about 30,000 nT) by some 4°.

There is, however, no reason to expect that these potential errors will give a consistent bias to all the cores, so we may address these three problems by assuming that all such effects are random between different sites and storage positions. Accordingly, we have given each core mean equal weight in computing averages and base confidence limits on the number of independent cores so included.

More critical to this analysis are those errors which might affect all cores similarly and thus add a systematic bias to the overall results. Two systematic errors may be present in the data set: a shallowing of inclination and a bias toward the present-day field direction. Our results, however, argue that neither error is particularly severe.

Systematic shallowing could be produced by the well-known detrital inclination error. This inclination error follows the form

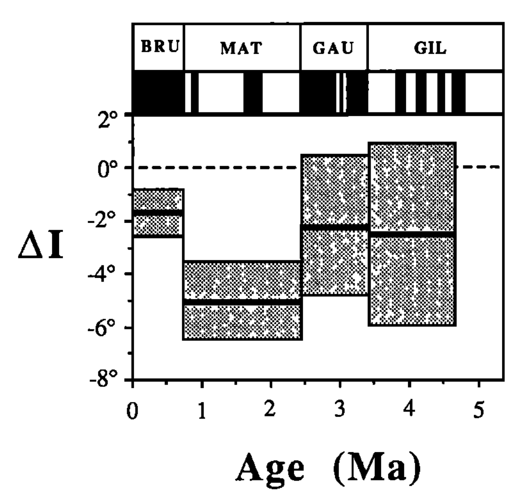


Fig. 4. Inclination anomaly averages for Brunhes (BRU), Matuyama (MAT), Gauss (GAU), and Gilbert (GIL) chrons with 1 standard error limits.

$$\tan I \text{ (observed)} = f \tan I \text{ (field)}$$

where the factor f can be considered, for example, the fraction of sedimentary magnetic particles that do not rotate into the horizontal upon deposition [King, 1955]. Such an error (Figure 5) would produce an inclination anomaly which has much the same form as one generated by a positive octupole term (i.e., having the same sign as the geocentric dipole term), with shallowed inclinations in both northern and southern hemispheres [Merrill and McElhinny, 1983]. Spherical harmonic analysis of inclination data from deep-sea sediments [Livermore et al., 1983] shows only a small octupole effect, $g^0_3$ being 1.5% of the dipole. This result effectively imposes an upper limit on the magnitude of any inclination error. If all of the octupole effect is caused by inclination error, then a value of f near 0.96 is appropriate for deep-sea sediments. An inclination error of this type would shallow inclinations at the locations studied here (average core latitude is 6° S) by less than 1°. Since the negative inclination anomalies found here generally correspond to inclinations that are steepened (compared with the expected dipole direction), an inclination error could not have caused the overall effect; rather, it would tend to minimize it, but only by a fraction of a degree.

An alternate mechanism for systematic shallowing of inclination has been proposed by Creer [1983], who points out that a unit vector analysis of directions will inherently give rise to a spurious $g^0_3$ component. This problem, however, is not particularly applicable to this study because inclinations are generally low and because we use only deep-sea sediments where most, if not all, of the time averaging is accomplished in situ.

A different source of systematic bias may come from the present-day field which in the study region is inclined relatively steeply upward at about -30°. Under alternating field treatment, the sediments shed a soft component of magnetization which often appears to be aligned with this present-day field direction; however, we cannot be assured that the demagnetization procedure completely removed this viscous overprint in every sample. Nevertheless, the presence of an unremoved present-day field component would not account for the results we have obtained. While a present-day field component might cause a steepening in the normal polarity (negative) inclinations, it could not account for the even greater steepening of the reverse (positive) polarity directions. We conclude that viscous overprints do not seem to be important and could not, in any event, explain the sense of polarity asymmetry displayed in these data.

## Discussion

### Introduction

We interpret these inclination anomalies as being a manifestation of the NDF. The overall effect observed here is consistent with the dominantly axisymmetric field models found in prior spherical harmonic analyses of fully oriented paleomagnetic data.

In order to compare our Indian Ocean results with previous spherical harmonic estimates, we may suppose the inclination effect is caused predominantly by an axial quadrupole field ($g^0_2$) since an octupole field ($g^0_3$) would have little effect at equatorial latitudes (as was shown above for inclination error) and thus cannot contribute significantly. We can thus estimate the magnitude of $g^0_2$ relative to the dipole ($g^0_1$); this gives a quadrupole to dipole ratio of about 2% during normal polarity and 6% during reversed polarity times. (The equivalent quadrupole is negative for normal polarity, positive for reversed polarity.)

These values compare reasonably well with previous estimates of the polarity dependence of $g^0_2$ based on global data [Merrill and McElhinny, 1977] which call for a 5%

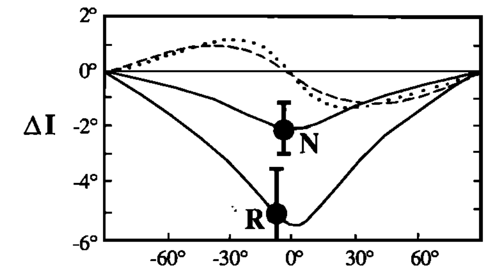


Fig. 5. Comparison of inclination anomaly produced by an axial octupole field for $g^0_3/g^0_1 = 0.015$ (dashed line) with the effect of detrital inclination error for F = 0.96 (dotted line). Also shown are normal (N) and reverse (R) polarity inclination anomalies determined in this study plotted with the best fitting quadrupole anomaly curves: $g^0_2$(Norm.)=2.5%; $g^0_2$ (Rev.)=6.2% (1 sigma standard error on data shown).

quadrupole during normal polarity and an 8% quadrupole during reverse. Our results are not completely independent, as there is partial overlap between the core data used in the previous spherical harmonic analysis and the supporting data included here. However, the results from our primary data set (which is completely independent from the previous analysis) are essentially the same.

The major difference between our results and those found by Merrill and McElhinny is the magnitude of the NDF effect, our results showing smaller anomalies overall. The disagreement might be caused by the differing effects of inclination error on the two data sets. As discussed, an artificial shallowing would in the ideal case just appear as an octupole field, but in practice it might add to the quadrupole estimate if the data are skewed toward nonequatorial, northern hemisphere sites (as may be the case with a global analysis). Conversely, any shallowing by inclination error in this study would tend to reduce the quadrupole estimate, the data being drawn largely from low southern latitudes.

A further cause for the difference between our results and those of Merrill and McElhinny may be the effects of plate motion. In the previous global study, plate motion could be expected to average out, but in this regional study it would not (especially since many of the sites are on a rapidly moving plate). Thus any inaccuracies in the absolute plate motion correction used here might give rise to the small discrepancies between these results.

The difference between normal and reversed polarity anomalies could be caused by some change of the NDF with time that is fundamentally unrelated to polarity; however, the averages for the four age/polarity groups (Figure 4) generally argue against this. The Brunhes anomaly has relatively small magnitude (-1.7°), as does the Gauss (-2.6°), while the Matuyama anomaly has a distinctly larger magnitude (-5.0°). The Gilbert estimate (-2.5°), though not as large as the Matuyama, has a 3.5° statistical error and thus is not incompatible with the proposed polarity effect which would predict -5° for this interval. The presence of a polarity effect over the interval covered in this study (0-5 Ma) does not, however, deny the possibility of longer-term changes in the NDF with time, as has been suggested in the global study of Coupland and Van der Voo [1980] and in the regional study of Epp et al. [1983].

The pattern of inclination anomaly with time thus appears to be tied to polarity state, though there are inconsistencies in the data: for example, an average anomaly for the lower third of the Brunhes (-6.4 ± 1.7°) has a larger magnitude than that for the lower third of the Matuyama (-1.7 ± 1.6°), contradicting the sense of the proposed polarity asymmetry. Our data, however, are not adequate to show whether variation of the NDF could result from changes during constant polarity

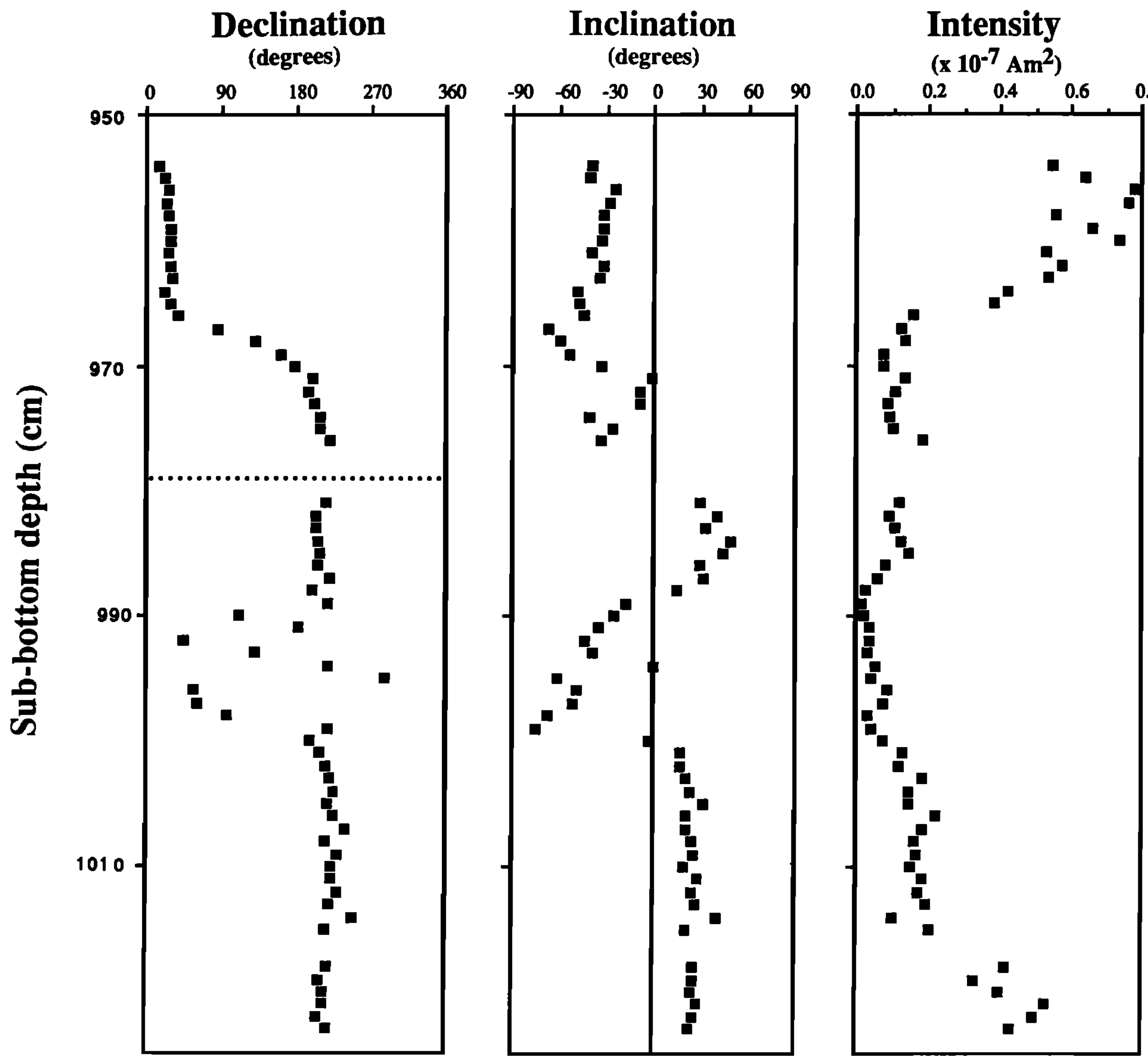


Fig. 6. Paleomagnetic record for the Matuyama to Brunhes transition in core RC14-046. Declination rotated 40° relative to the split face of the core. Note that intensity is plotted as total moment (i.e., not normalized by sample weight or ARM intensity) and thus should be considered only a crude indicator of field intensity.

intervals. In any case, the overall results cannot be easily ascribed to any simple linear trend with time and seem most simply explained by a polarity dependence, with the NDF being larger during reversed polarity intervals.

Possibility of a Standing Nondipole Field

Merrill and McElhinny [1977] discussed the polarity asymmetry they observed in the time-averaged field by decomposing the NDF into standing and reversing parts. During normal polarity times, the standing NDF would oppose the reversing NDF, giving a relatively small overall effect. Conversely, during reverse polarity times the standing and reversing components would add constructively, giving a larger effect. Our results are generally consistent with a polarity dependence, and thus they support such a separation of the NDF. Our equatorial data, however, cannot show which of the even harmonic terms may be contributing. For example, our results are consistent with a 4% reversing $g^0_2$ and a 2% standing $g^0_2$, but they are equally consistent with the same reversing $g^0_2$ plus a 2% standing $g^0_4$ field (of negative sign). Both would give virtually the same $\Delta I$ near the equator. In any case, the sense of polarity asymmetry implies that a standing field should have a downward directed component at equatorial latitudes.

The separation into standing and reversing fields may be merely a convenient description of two inherently different modes of the geodynamo. But more significantly, it may indicate two physically independent sources of the time-averaged field: one source periodically reversing (associated with the generation of the well-established geomagnetic polarity time scale), while the other, much smaller, source persists, at least over the span of a few million years.

Merrill et al. [1979] suggested that a true standing field might manifest itself during polarity transitions when the reversing main field has least influence. The nondipolar transitional field might then reflect something of the proposed standing NDF. Though Merrill et al. tentatively concluded that existing transition records did not show any standing field, newer transition data are generally more supportive.

Many of these transitions have been successfully described using the zonal harmonic model of Williams and Fuller

[1981], which partitions the energy lost by the dipole field into the low-degree axial nondipole terms. In such modeling, the NDF terms maintain one sign throughout and act much as a standing field, at least over the course of the transition. There is, however, no consensus on the proper choice of energy partitioning in such modeling. For example, Williams and Fuller [1981] adopted a scheme which included a positive $g^0_2$ term to fit Brunhes-Matuyama transitions in the northern hemisphere, while Clement and Kent [1984] were equally successful in modeling a lower Jaramillo transition from the southern hemisphere using a negative $g^0_2$.

Despite this uncertainty in the relative importance of the different spherical harmonic terms, the general success of the zonal harmonic models suggests that a true standing NDF might well influence transitional directions. In particular, records from low latitude sites would be sensitive only to the even terms, such as $g^0_2$ and $g^0_4$, and may show whether the standing field is downward directed near the equator, as would be expected if the transitions sense the same standing field as is indicated by the time-averaged field results.

Two inclination records of the Brunhes-Matuyama transition from east equatorial Pacific cores [Freed, 1977] modeled by Williams and Fuller [1981] show steep downward directions. Another equatorial Pacific core studied by Clement et al. [1982] shows a similar downward steepening during the transitional intensity low. Moreover, of the eight records of various transitions from equatorial Pacific cores studied by Theyer et al. [1985], four show transitional fields with relatively steep downward inclinations; the remaining four records do not display particularly steep inclinations in either direction.

The data from these Pacific cores thus suggest that transitional fields near the equator might well be influenced by a standing NDF similar to that indicated by polarity asymmetry in the time-averaged field. We have examined one Matuyama to Brunhes transition from an Indian Ocean core in some detail for an unrelated study, but include it here (Figure 6), as it is relevant to the investigation of a standing NDF. This transition appears quite complicated, with the inclination values changing sign thrice through the transitional intensity low. As the steepest inclination values tend to be negative, this record is not consistent with the presence of a downward directed standing field.

Though this example disagrees with the majority of other equatorial transition records reported, the discrepancy is perhaps not entirely unexpected. Transitional field observations are especially prone to many of the error sources discussed above. For instance, local magnetic fields (of remanent origin) would be proportionally much larger during a field transition when the main field intensity is low and may even dominate the directions recorded in some deep-sea cores. We obviously need many more equatorial records to fully gauge the merit of this intriguing connection between transitional and time-averaged fields.

## Summary

Anomalous inclination values recorded in Plio-Pleistocene sediments from the equatorial Indian Ocean indicate significant departures from dipole directions which we interpret as a manifestation of long-term, nondipole components of the time-averaged field. Though we cannot determine which of the NDF terms contribute using data from one region alone, we may reasonably surmise that the inclination anomalies observed are associated predominantly with the low-degree even zonal harmonics ($g^0_2$ and $g^0_4$). This would be consistent both with previous global analyses and with Cox's [1975] model that suggests the time-averaged NDF might result from continual westward drift of the instantaneous field.

In agreement with previous studies, the magnitude of the NDF effect found here appears to depend on polarity and, in general, has a larger value during reversed polarity times. Though some of the data examined appear to deviate from this pattern, no simple monotonic change in the NDF with time is indicated, and a polarity dependence appears to be the least complicated explanation for the variability of our inclination data.

The polarity dependence can be cast as the superposition of standing and reversing NDF terms, suggesting that an independent mechanism for generating these two fields may operate. A standing field that is directed downward at low-latitude sites would account for the differences found here in time-averaged field directions between polarity states. Such a standing field would also explain the preponderance of steep positive inclinations in transition records from equatorial latitudes. The proposed standing nondipole field could thus link these observations that examine the geomagnetic field at time scales which, spanning three orders of magnitude, are vastly different.

Acknowledgments. We wish to thank David Johnson for sharing his biostratigraphic results and helping us to identify many pre-Brunhes sediment cores. We also thank Jean-Pierre Caulet for his help and support during our sampling of Marion Dufresne cores. Further thanks are due to Brad Clement, Rich Jarrard, Ron Merrill and John Miller for providing many constructive comments. The work was supported by NSF Oceanography grant OCE84-00805. Lamont-Doherty Geological contribution #4319.

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(Received September 7, 1987;
revised January 13, 1988;
accepted January 19, 1988.)