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Geomagnetic Reversals

Although decreasing rapidly, the earth's magnetic field is probably not now reversing.

Allan Cox

The periods of the earth's magnetic field that are most crucial to an understanding of the geomagnetic dynamo occur at the ultralow-frequency end of the spectrum. Yet, until recently, this part of the spectrum received relatively little attention, mainly because the spectrum extends to periods longer than the productive lifetimes of individual scientists. The longest periods are, indeed, longer than the entire history of scientific observation. The first hint of the existence of periods greater than 100 years came from observations of changes in field intensity. In 1835—the first year for which C. F. Gauss was able to assemble enough worldwide data to analyze the field through the use of spherical harmonic functions — the earth's dipole moment was 8.5×10^{25} gauss cm^3. By 1965, the moment had decreased to 8.0×10^{25} (*1*). The dipole field decreased during this time at a remarkably uniform rate, and it will, if the rate remains constant, pass through a zero point about 2000 years from now and then reverse its polarity (*2*). However, other interpretations are equally consistent with the intensity data—for example, the field might oscillate without changing polarity. Therefore, although there is little question that the geomagnetic spectrum extends to periods considerably greater than the few centuries spanned by the records of magnetic observatories, the nature of the spectrum has long remained uncertain.

The length of the available magnetic record has now been increased by more than six orders of magnitude through the study of the natural magnetism of rocks and baked clay which retain a magnetic memory of the earth's field in the past. This paleomagnetic research indicates that the earth's field has undergone numerous fluctuations of intensity and that it has also undergone many (although less numerous) reversals in polarity. Recent work suggests that these two phenomena may occupy adjacent parts of the geomagnetic spectrum, the division between them being at periods of about 10^4 years. This part of the spectrum, which is very difficult to resolve experimentally, may hold the key to understanding how polarity reversals are related to intensity fluctuations.

Changes in the Earth's Dipole Moment

The technique for measuring ancient geomagnetic field intensity is based on the observation that, when volcanic rocks and pieces of pottery are cooled in weak magnetic fields, they acquire thermal remanent magnetization which is parallel in direction and proportional in intensity to the applied field. If the natural thermal remanence of rocks is magnetically stable, the ancient field acting on a sample when it originally cooled may be found by reheating the sample and cooling it in a known field. The ancient field intensity F_o is then given by

$$F_o = F_a \frac{J_o}{J_a} \quad (1)$$

where J_o is the natural remanence and J_a is the remanence acquired in the known applied field F_a (*3*).

This technique, although simple in concept, is difficult to carry out experimentally because, on being reheated in the laboratory, the ferromagnetic minerals contained in rocks and baked clay commonly undergo chemical changes to form new ferromagnetic minerals. Clearly, the magnetization acquired when the altered samples are cooled in a known field does not provide a measure of the ancient field. However it has proved possible experimentally to identify samples which are chemically stable on heating. These yield values for ancient field intensity which are internally very consistent and appear to be reliable.

In all, 127 paleomagnetic intensity determinations have now been made on samples with ages $0 < t \leqslant 10^4$ years. This work, which was carried out in different laboratories on samples from many parts of the world, was summarized recently by P. J. Smith (*4*), who reduced each ancient field intensity to a virtual dipole moment, defined as the moment of the dipole needed to produce the observed ancient intensity if the earth's field had been entirely dipolar. Virtual dipole moments for samples of the same age from different parts of the world are scattered, with a standard deviation of about 20 percent, because the earth's field consists in part of an irregular nondipole component. This scatter has been reduced in Fig. 1 by averaging virtual dipole moments for different parts of the world. The decrease in dipole moment since 1885, as shown by the slanting bar at left in Fig. 1, is seen to be but the most recent part of a well-defined half cycle with a maximum of 12×10^{25} gauss cm^3, which began about 4000 years ago.

The author is professor of geophysics at Stanford University, Stanford, California, and does part of his research with the U.S. Geological Survey. This article is an expanded version of a paper presented 8 April 1968 at a meeting of the American Geophysical Union in Washington, D.C.

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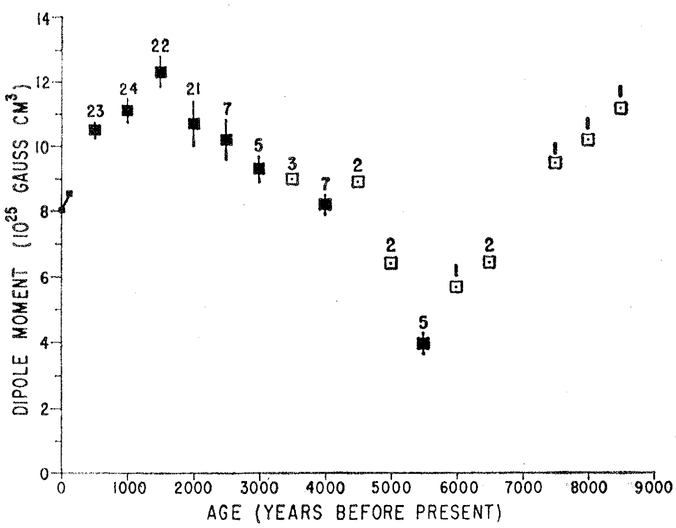


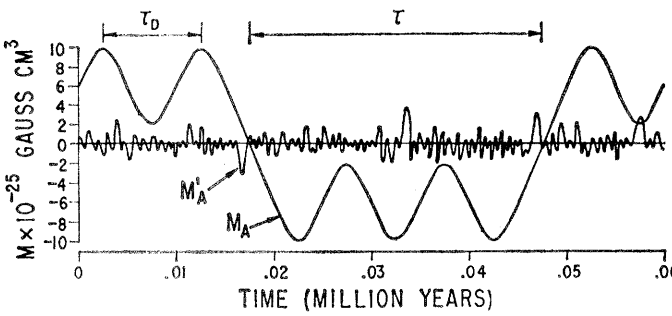


Fig. 1 (left). Variations in geomagnetic dipole moment (*12*). Changes during the past 130 years, as determined from observatory measurements, are shown by the slanting bar at left. Other values were determined paleomagnetically (*4*). The number of data that were averaged is shown above each point, and the standard error of the mean is indicated by the vertical lines (*12*) except for points (open squares) with too few data to provide meaningful statistics. Fig. 2 (above). Model for reversals used to derive the distribution function for reversals. τ_D is the period of the dipole field and τ is the length of a polarity interval. A reversal occurs whenever the quantity M'_A, which is a measure of the nondipole field, becomes sufficiently large relative to the dipole moment M_A (*12*).

An earlier half cycle is weakly suggested by the few older intensity measurements. The important question of whether changes in field intensity are periodic cannot be answered from the present data, but there can be no question that dipole fluctuations occur, with durations of the order of 10^4 years.

Polarity Reversals

Extension of the known geomagnetic spectrum to even longer periods has come about with the discovery that the earth's field undergoes reversals in polarity. Through paleomagnetic research it has been found that the earth's dipole moment alternates between two antiparallel polarity states, a *normal* state, in which the field at the earth's surface is directed northward, and a *reversed* state in which the field has the opposite direction. In either state the dipole undergoes an irregular wobble about the earth's axis of rotation, with an angular standard deviation of 12 degrees (*5*). The time required to complete a transition between polarity states is estimated to be from 10^3 to 10^4 years (*6–8*); during this time the field does not go to zero but undergoes an intensity decrease of from 60 to 80 percent (*4–9*). Averaged over a long sequence of reversals, the total amount of time spent by the dipole in the reversed state is equal to that spent in the normal state (*10–12*). Moreover, the average intensity of the field in the two states is the same (*4*), suggesting that the two states have equal energy levels. In all of these characteristics the earth's field operates like a remarkably symmetrical oscillator or, more precisely, like a bistable flip-flop circuit.

The greatest element of irregularity in reversals is the length of time between successive changes in polarity. The longest known polarity interval lasted for 5×10^7 years (*13*); intervals with lengths of the order of 10^6 years are common, and short polarity intervals occur, with durations of less than 10^5 years. The latter have proved to be the most difficult to measure. This article deals mainly with recent experimental work on short polarity intervals.

Distribution Function for Polarity Intervals

The results from paleomagnetic studies of reversals may be presented either in the form of a time scale for reversals or, more compactly, in the form of a histogram showing the frequency of occurrence of polarity intervals as a function of their length. The question of what distribution function should be used to describe the observed frequency distribution depends on the more fundamental questions of why reversals occur and what controls the immense variation in the length of time between them.

It is now a generally accepted theory that the earth's field is generated by hydromagnetic processes in the earth's fluid core and that the energy which maintains the field against ohmic dissipation is provided either by thermal convection or by turbulence generated by the earth's precession (*14*). Unfortunately, however, the theory of the homogeneous fluid dynamo is not sufficiently developed to account quantitatively either for fluctuations in dipole intensity or for polarity reversals. Particularly difficult is the problem of reconciling the long times between polarity reversals, which may exceed 10^6 years, with the much shorter time constants of dipole fluctuation (10^4 years) and fluctuations of the nondipole field (3 years to 10^3 years).

In the absence of a complete theory of geomagnetism, some interesting and possibly relevant analogies are provided by the behavior of nonfluid self-excited dynamos. The simplest one that provides a model for geomagnetic intensity fluctuations and polarity reversals consists of two mutually coupled Faraday disk dynamos (*15*). In some solutions for this dynamo the current oscillates about a mean current flowing first in one and then in the opposite direction. The distribution function for the lengths of time between successive reversals in current direction depends on (i) the period of the oscillations and (ii) the number of cycles between successive reversals. Both are sensitive to changes in the physical parameters of the model, some solutions for the dynamo having sharply peaked distributions and others having monotonic decreasing functions, the shortest polarity intervals being the most frequent.

The earth's field is produced by a much more complex fluid dynamo, so that detailed comparisons with particular solutions for the disk dynamos are of little value. Unlike the rigid disks of the dynamo, the earth's core resembles the atmosphere in possessing a large random component of motion due to turbulence. A direct measure of this component is provided by the rapidly varying nondipole field, the average intensity of which is 20 percent of the dipole field. The sensitivity of the timing of reversals in a solid disk dynamo to small changes in the physical conditions of the model suggests that in fluid dynamos the timing of reversals would also be sensitive to a large random element in the pattern of fluid motions and magnetic fields.

The foregoing considerations suggest the following model, from which a distribution function for polarity intervals can be derived (*12*) (Fig. 2). The main geomagnetic dynamo is assumed to be a steady dipole oscillator which undergoes a polarity reversal only when this is triggered by random fluctuations of the much more rapidly varying nondipole field. The probability P that a reversal will occur during one cycle of dipole oscillation is assumed to be the same for all cycles and is determined by the spectrum of nondipole fluctuations and the amplitude of dipole oscillations. The resulting distribution function for variations in the length T of polarity intervals depends only on the probability P and the period T_D of dipole oscillation:

$$f[T_C < T \leqslant T_C + T_D] = P(1 - P)^{T_C/T_D} \quad (2)$$

where T_c is an integral number of dipole periods. For small values of P, this may be approximated by a simple exponential distribution

$$f'(T) = \lambda \exp(-\lambda T)\ dT \quad (3)$$

where $\lambda = P/T_D$ and $f'(T)$ is the probability for the interval dT.

Before comparing this function with experimental data, one should note the extreme sensitivity of the distribution function to the proportion of short events. This sensitivity is much greater than that of the spectrum of a continuously varying signal to a high-frequency component. Because the earth's field has only two polarity states, inserting a short polarity interval near the middle of a long interval not only adds the short interval to the distribution but also removes a long interval and adds two intervals of intermediate length. This effect may be seen (Fig. 3) in the marked change that occurred in the apparent distribution of polarity intervals as several short polarity intervals were discovered during the course of research between the years 1963 and 1966.

Radiometric Time Scale for Reversals

The first quantitative time scale for reversals was achieved in 1963 by measuring the ages (using the potassium-argon technique) and the magnetic polarities of young volcanic rocks (*16*). This work appeared to confirm earlier paleomagnetic results from undated sedimentary rocks which indicated that all polarity intervals were of nearly equal length (Fig. 3, 1963 scale). These polarity intervals were termed "polarity epochs" and given the names of early workers in the field of geomagnetism (*17*).

As more data were obtained (*18*), some of the ages and polarities were found to be inconsistent with the simple pattern of epochs. This inconsistency led in 1964 (*19*) to the discovery of intra-epoch "polarity events" with unexpectedly short durations of about 10^5 years (Fig. 3). The first polarity events to be recognized were the Olduvai event ($t = 1.9 \times 10^6$ years ago) and the Mammoth event ($t = 3.1 \times 10^6$ years ago), named after the sites of their discovery in Tanzania and California. A paleomagnetic record of polarity events has been found in rocks with similar ages from many parts of the world, demonstrating that the events, like the epochs, are the result of rapid switching of the main dipole field.

With the discovery of polarity events it became apparent that complete resolution of the fine structure of reversals would require an immense number of paleomagnetic and radiometric data. Ideally, such data should be obtained from formations whose ages are uniformly spaced at intervals no greater than 10^4 years; this would require at

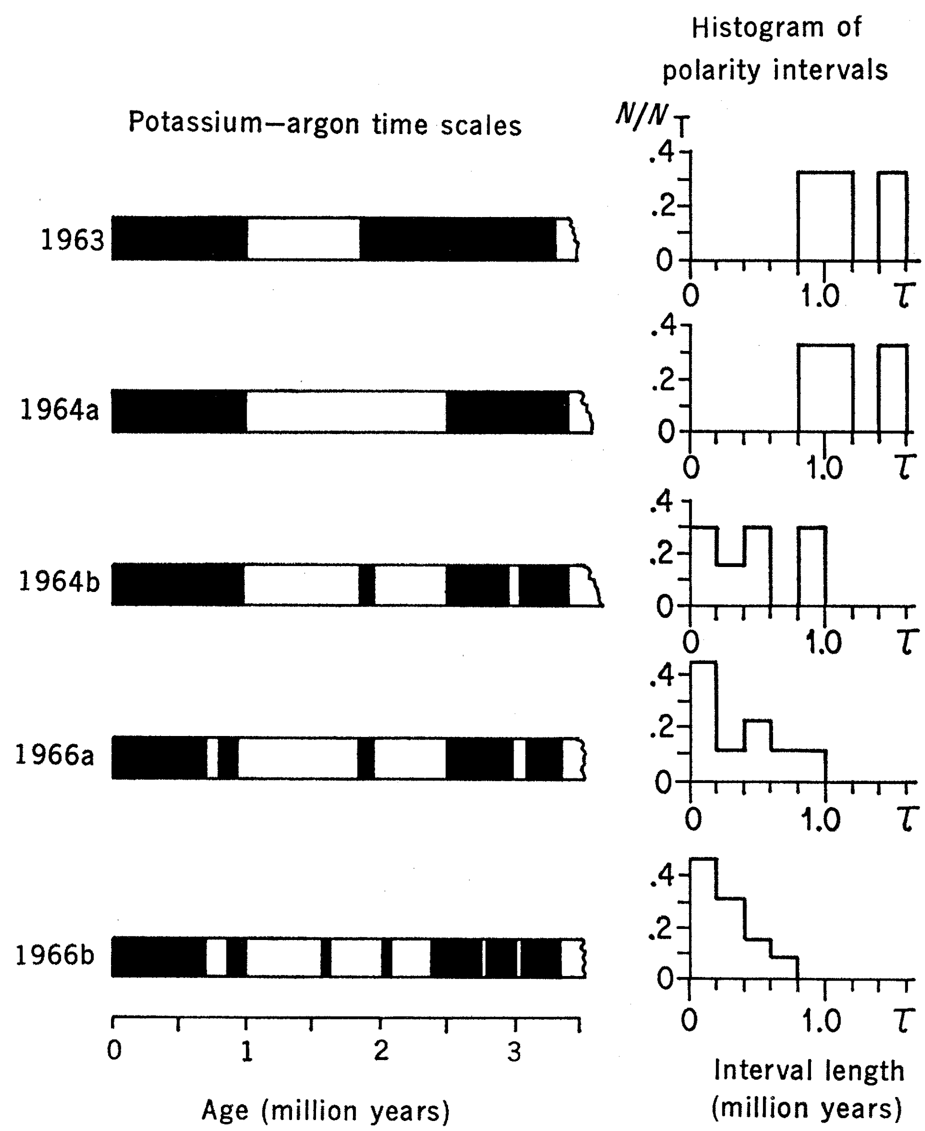
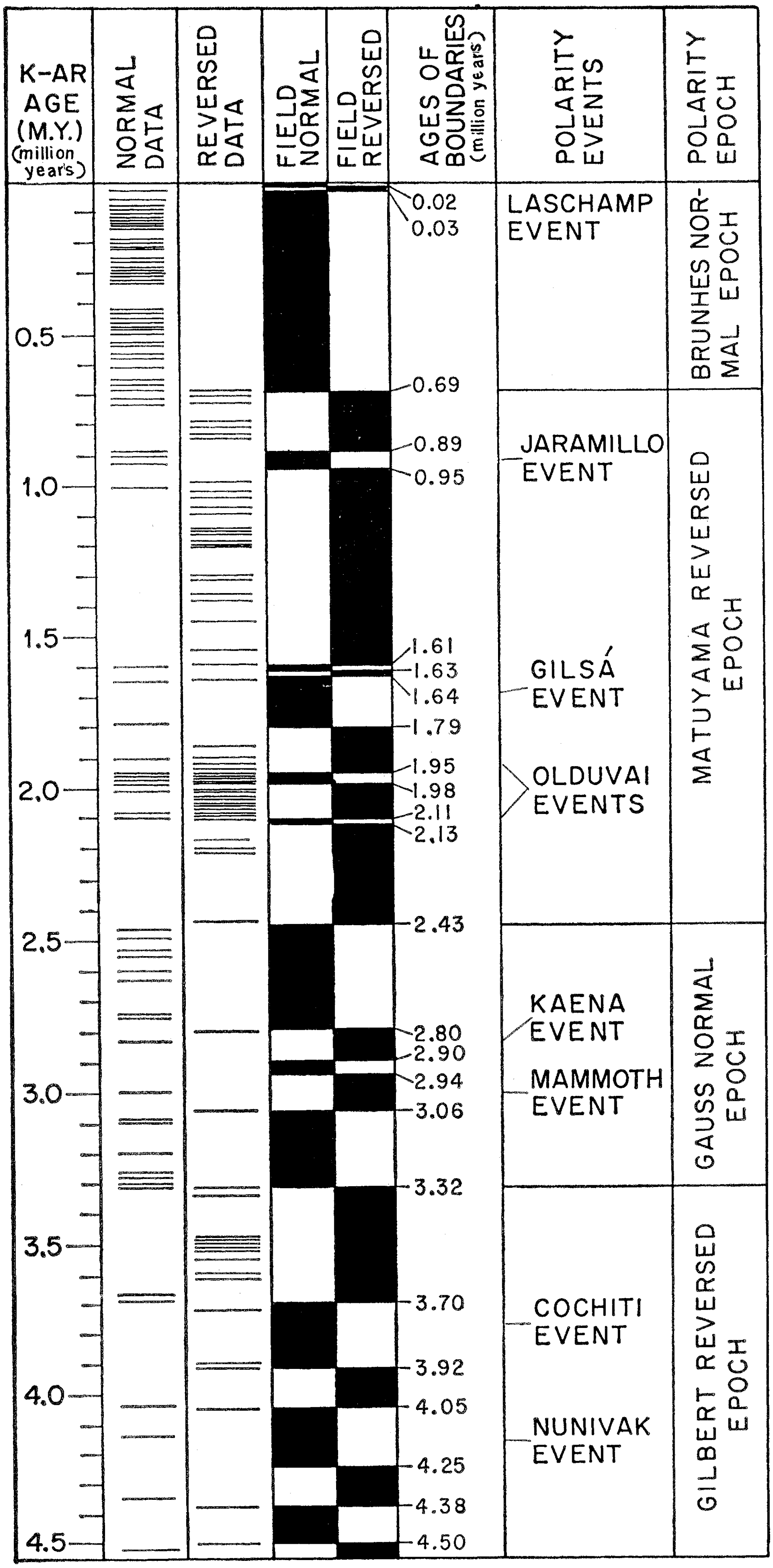


Fig. 3. Successive versions of the radiometric time scale for reversals, showing how the discovery of polarity events changed the apparent distribution of polarity intervals (*16, 18, 19*). In the corresponding histograms, N_T is the total number of polarity intervals and N is the number in each class interval of the histogram.



least 450 precisely determined radiometric dates for the interval $0 < t \leqslant 4.5$ million years. The task of making this many age determinations is formidable, and that of locating volcanic formations with the required ages is even more formidable. The age of a volcanic formation is rarely known to within a factor of 2 prior to determination of the radiometric age. Thus the best sampling scheme that can be hoped for is a relatively inefficient one in which the ages are randomly rather than uniformly distributed. Even this is difficult to achieve because of the episodic character of volcanic activity.

For the interval $0 < t \leqslant 4.5$ million years there are now 150 radiometric ages and polarity determinations which meet reasonable standards of reliability and precision (*20*). The main contributions from the data acquired since 1964 have been more accurate determination of the ages of polarity changes (*8*) and the identification of several hitherto unknown events, summarized in Fig. 4. Attempts to extend the radiometric time scale for reversals back beyond 4.5 million years have been unsuccessful because the errors in the radiometric ages of the older rocks are too large. A dating error of 5 percent in a 5-million-year-old sample is 2.5×10^5 years, which is larger than many polarity intervals (*8*, *11*, *21*).

Have all the polarity events more recent than 4.5 million years ago been discovered? All of the longer ones appear to have been, to judge both from the density of the present age determinations and from the convergence of the more recent versions of the reversal time scale. However, there are gaps of 10^5 years or more in the present data, and the recent discovery by Bonhommet and Babkine (*22*) of a previously unsuspected reversed event near the end of the Brunhes normal-polarity epoch demonstrates that additional events may exist in gaps even shorter than 10^5 years.

Fig. 4. Time scale for geomagnetic reversals. Each short horizontal line shows the age as determined by potassium-argon dating and the magnetic polarity (normal or reversed) of one volcanic cooling unit. Included are all published data (*37*) which meet reasonable standards of reliability and precision (*20*). Normal-polarity intervals are shown by the solid portions of the "field normal" column, and reversed-polarity intervals, by the solid portions of the "field reversed" column. The duration of events is based in part on paleomagnetic data from sediments (*7*, *28*) and magnetic profiles (*11*, *23*, *24*, *27*).

Midoceanic Magnetic Anomalies

Additional information about reversals is provided by the magnetic anomalies over the midoceanic ridges (*11, 23, 24*). These anomalies are produced by igneous rocks which become magnetized as they solidify and cool in a narrow zone along the ridge axis. As new material forms, the previously magnetized material spreads to either side. If the rate of spreading is the same on both sides of the ridge, the result is a bilaterally symmetrical pattern of normally and reversely magnetized strips with widths proportional to the lengths of the corresponding polarity intervals. The magnetic anomalies do not in themselves determine an independent reversal time scale because the ages of the igneous rocks are usually not known. However, after being calibrated against known points on the radiometric time scale for reversals, the profiles provide a nearly continuous record of polarity intervals. Of special interest are events that may exist within the gaps in the radiometric data.

The value for the minimum duration of a detectable event depends on the width of the strip of crust that was formed during the time of the event, the distance of the strip from the magnetometers at the sea surface (usually about 3 kilometers), and the level of background noise due to irregularities in the formation and magnetization of the crust. The strong dependence of the minimum value for duration on the rate of crustal spreading, v, may be seen from variations in the size of the anomaly due to the Jaramillo event ($T = 5 \times 10^4$ years). This anomaly is quite large on the profiles across the East Pacific Rise ($v = 4$ to 5 centimeters per year) and the Juan de Fuca Ridge ($v = 3$ centimeters per year) but is not visible on profiles across the Reykjanes Ridge ($v = 1$ centimeter per year) and is at about the limit of resolution on the profile across the Indian Ocean ($v = 2$ centimeters per year) (*11, 23, 24*). The minimum strip width ($v \times T$) detectable at the sea surface is therefore about 1 kilometer, and the shortest detectable event is about 2×10^4, 5×10^4, or 10^5 years long, depending on whether the rate of lateral spreading is 5, 2, or 1 centimeter per year (*25*).

On the profiles one can distinguish small peaks due to polarity events from those due to magnetic noise only by determining which peaks are consistent from profile to profile. A difficulty in doing this arises from the fact that variations in the rate of lateral spreading along one profile may, over long distances, produce cumulative displacements equal to half the wavelength of anomaly peaks, producing a large loss of signal on cross-correlation. In looking for short events it is desirable to tie the magnetic profiles as closely as possible to well-determined points on the radiometric time scale. Displacements due to variable rates of spreading may then be minimized by interpolating between closely spaced points.

Table 1. Ages of the three normal intervals (YG, IG, and OG) in the Gauss normal epoch, as found by interpolation between the Gilbert-to-Gauss boundary (3.32 million years ago) and the Gauss-to-Matuyama boundary (2.43 million years ago) (*36*).

Interval	Number of profiles	Age found by interpolation (million years)	Standard deviation	Standard error	Age of interval midpoint, from K-Ar dating (million years)
YG	13	2.64	0.03	0.01	2.62
IG	8	2.94	.10	.04	2.9
OG	13	3.19	.03	.01	3.19

Figure 5 shows the total magnetic field anomaly along the Eltanin 19 profile, which crosses the East Pacific Rise at latitude 52°S and longitude 118°W. As is typical of midoceanic ridges, the positive anomaly over the central zone (M-B to M-B in Fig. 5) is complex, for reasons not yet understood. Elsewhere, many of the polarity transitions can be correlated unambiguously with those of the radiometrically determined time scale. For example, the transition from the Matuyama reversed epoch to the Brunhes normal epoch can be easily recognized (M-B in Fig. 5), as can the transition from the Gauss to the Matuyama epochs (G-M) and the transition from the Gilbert reversed to the Gauss normal epoch (G-G). At high latitudes the steepest gradients in the magnetic profiles occur almost exactly above the boundaries between the normally and the reversely magnetized strips, so that distances between polarity transitions and events may be read directly from the magnetic profiles. It may be seen that the anomalies to the northwest of the rise are more widely spaced; this indicates that the rate of spreading was greater in the direction of the Pacific than toward Antarctica. Therefore the interpolations must be made separately for the two sides of the rise.

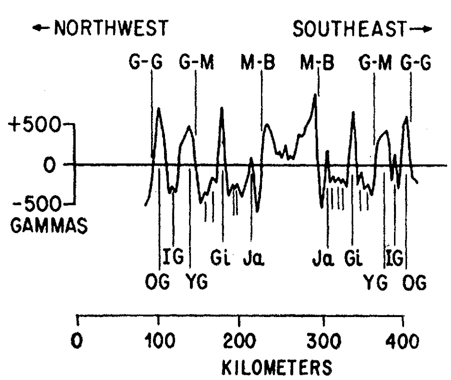


Fig. 5. Magnetic profiles across the East Pacific Rise at latitude 52°S, longitude 118°W (*11, 24*). (G-G, G-M, M-B) Boundaries between the Gilbert, Gauss, Matuyama, and Brunhes polarity epochs. (OG, IG, YG, Gi, Ja) The larger and more consistent positive anomaly peaks. Other small positive peaks are shown with unlabeled lines.

The age of the boundary between the Gauss and Matuyama epochs was first found, as a check on the method, by interpolating between the Gilbert-to-Gauss boundary (3.32 million years ago) and the Matuyama-to-Brunhes boundary (0.69 million years ago), both of which have well-determined radiometric ages. Fifteen half profiles were used, all from high latitudes in the South Pacific, North Pacific, and Indian oceans. The mean interpolated age is 2.41 ± 0.03 million years (standard error). This age is consistent with the two radiometrically determined ages of 2.43 million years (reversed magnetization) and 2.45 million years (normal magnetization) which bracket the boundary. When both radiometric and interpolated results are taken into account, the best estimate of the age of the boundary is 2.43 million years.

Three positive anomaly peaks corresponding to the three normal polarity intervals (YG, IG, and OG in Fig. 5) of the Gauss normal epoch appear on those profiles for which spreading rates are rapid. The two reversed events, the Mammoth and the Kaena, which separate the normal intervals have been recognized by radiometric dating (*18, 19*) and from magnetic profiles (*11, 23, 24*). The ages of the midpoints of the two larger anomalies, found by interpolating 13 profiles (Table 1) between the boundaries of the Gauss

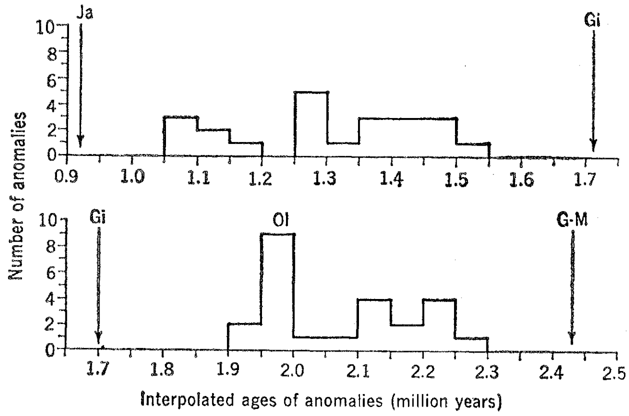


Fig. 6. Histogram of ages of small peaks on magnetic profiles found by interpolating between the Jaramillo event (0.92 million years ago), the Gilsá event (1.71 million years ago), and the Gauss-Matuyama boundary (2.43 million years ago). The most consistent anomaly peak corresponds to the Olduvai event (Ol).

epoch, are in good agreement with the radiometric time scale. The short central anomaly peak has not been dated radiometrically but is bracketed by ages of 3.0 and 2.8 million years obtained for samples from the adjacent reversed events, in agreement with the interpolated age of the normal event of 2.94 ± 0.04 million years.

Two positive anomalies (Ja and Gi in Fig. 5) appear in the Matuyama epoch on almost all magnetic profiles. The smaller and younger anomaly (Ja) has usually been identified with the Jaramillo normal event; the older and larger anomaly, with the Olduvai event (*11*, *23*, *24*). The age of 0.93 million years (Table 2) found for the younger anomaly by interpolating between the boundaries of the Matuyama reversed epoch agrees well with the radiometric age of 0.92 million years. However the interpolated age of 1.70 million years for the larger anomaly does not agree with the ages obtained for the Olduvai normal event by potassium-argon dating, most of which are from 1.95 to 2.00 million years. It agrees more nearly with the two ages of 1.60 and 1.65 million years for normally magnetized lava flows from Iceland and Alaska (*26*), which constitute the evidence for the existence of a normal event termed the Gilsá (*26*). Vine (*27*) concluded independently, from detailed study of the Eltanin 19 profile, that the main anomaly in the Matuyama reversed epoch lasted from 1.80 to 1.64 million years ago. Previously Ninkovich *et al.* (*7*) had concluded from their paleomagnetic study of marine sediments that the main normal event in the Matuyama reversed epoch began 1.79 million years ago and ended 1.65 million years ago, but they correlated this event with the Olduvai event, as it had been defined previously from radiometric dating (*19*). It now appears that the main normal-polarity event in the Matuyama epoch is not the Olduvai but rather the Gilsá event. Its younger boundary is determined from radiometric dating to be at 1.61 million years ago. Its older boundary, although poorly determined radiometrically, appears from the magnetic anomalies and from the paleomagnetism of sediment cores (*7*) to be at about 1.79 million years ago.

To determine whether the profiles contain evidence for shorter events, ages were found by interpolation for all the small positive peaks in the Matuyama reversed epoch (Fig. 5) on 17 half profiles from high latitudes. The most consistent anomaly appears on 11 half profiles at positions having interpolated ages between 1.94 and 2.00 million years (Fig. 6). The radiometric ages associated with the Olduvai event are almost all in this same range, a fact which indicates that the two are correlative. From the size of the anomaly the duration of the event is estimated to be 3×10^4 years, an estimate which agrees with Vine's (*27*) interpretation of the Eltanin 19 profile. On the other hand, in paleomagnetic studies of deep-sea sediments no consistent evidence has been found for the existence of both the Gilsá and the Olduvai events (*28*). Apparently the processes by which sediments become magnetized are sufficiently irregular and integrative in nature as to make it difficult to consistently resolve events as short as 3×10^4 years.

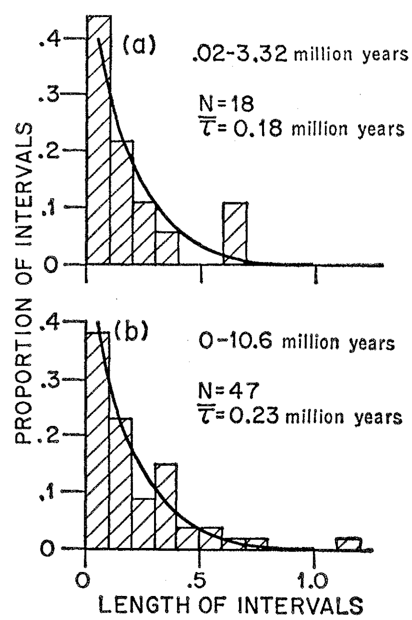


Fig. 7. Histograms of lengths of polarity intervals, (a) from the reversal time scale of Fig. 4, based on ages obtained by potassium-argon dating, and (b) from the reversal time scale of Vine (*27*) based on the Eltanin 19 magnetic profile. (Solid curves) Distribution function of Eq. 2 for $P = 0.05$. N is number of intervals.

Why have so many normally magnetized rocks been found from the short Olduvai event and so few from the much longer Gilsá event? The answer appears to lie in the uneven distribution of the ages of both the normally and the reversely magnetized samples. Few samples of either polarity have ages in the range between 1.6 and 1.8 million years, whereas many samples of both polarities have ages between 1.9 and 2.1 million years. The overlap in the ages of normally and reversely magnetized samples with ages between 1.9 and 2.1 million years is due to the fact that the duration of the Olduvai event is shorter than many of the discrepancies due to dating errors. The high ratio of reversely to normally magnetized samples is in accord with the conclusion that the Olduvai normal event was short.

Rocks with ages of from 2.2 to 1.9 million years used for these studies are from volcanic formations from Africa, Alaska, the western United States, Cocos Island, Australia, Reunion Island in the Indian Ocean, and Iceland—a range which suggests that an interval of unusually intense volcanic activity occurred 2 million years ago. It was also at about this time that a marked evolutionary change occurred in marine microorganisms, including Radiolaria (*29*) and Foraminifera (*30*), marking the beginning of the Pleistocene. Such faunal changes have been noted near several polarity transitions and, in particular, at the time of or slightly before the Olduvai event (*29*, *30*), lending support to earlier suggestions (*31*) that, when the earth's field decayed during a reversal, the increase in radiation would have been large enough to produce a sudden increase in the rate of evolution. However, recent quantitative studies indicate that the shielding provided by the earth's atmosphere is so great that the increase in radiation on complete collapse of the field would be no more than 12 percent (*32*). Such an increase is comparable with the variations that occur during a normal sunspot cycle, or with the normal variation between equatorial and polar regions. The accompanying increase in mortality rate and in the rate of spontaneous mutation is too small to account for the extinction of a species. An alternative explanation is that the faunal extinctions at the beginning of the Pleistocene are related to volcanic activity which occurred at that time, a possible causal link being increased absorption and reflection of sunlight by volcanic dust in the atmosphere.

An anomaly slightly older than the Olduvai [the X anomaly of Heirtzler *et al.* (*11*)] appears on some profiles. The interpolated ages, although less well grouped than those of the Olduvai anomaly, provide weak evidence that a short event may exist. Support for this is provided by two radiometric ages slightly greater than those of the main group of normally magnetized Olduvai samples (Fig. 4), but again the evidence is rather weak. There are weak suggestions of possible events at 1.26 and 2.24 million years ago, but whether they are real cannot be determined radiometrically because both occur where there are gaps in the data.

In summary, all of the events with durations $T \geqslant 3 \times 10^4$ years that have been found from radiometric dating have also been identified on the magnetic profiles. The radiometric ages of the events and the ages derived through interpolation from the profiles are in excellent agreement. The profiles further establish that there are no additional, undetected events longer than 3×10^4 years, even where there are gaps longer than this between known radiometric ages. Events shorter than 3×10^4 years are below the level of experimental noise and have not been resolved convincingly from the magnetic profiles.

Table 2. Age of the Ja and Gi events, as found by interpolation between the Gauss-to-Matuyama boundary (2.43 million years ago) and the Matuyama-to-Brunhes boundary (0.69 million years ago) (*36*).

Interval	Number of profiles	Age found by interpolation (million years)	Standard deviation	Standard error	Age from K-Ar dating*
Ja	17	0.93	0.05	0.01	0.92
Gi	17	1.70	.10	.02	1.60–1.86

* See Fig. 4 for range of uncertainty.

Sedimentary Record of Short Events

The paleomagnetism of marine sediments provides a third possible source of information about short events (*28–30*). In general, the record of polarity epochs in sediments agrees with the radiometric time scale. However, the record of short events is obscured by noise due to stratigraphic gaps, variations in the rate of deposition, and delays in the time between deposition and magnetization of the sediments. The amount of delay varies, depending on variations in the amount of reworking by organisms, on the rate of compaction, and on authigenic chemical changes in the ferromagnetic minerals. In addition, there is a loss of information about short polarity intervals because the magnetizing processes in sediments are integrative over time intervals comparable with the durations of the shorter events. As a result, the research on sediments has not produced consistent evidence for the existence of events shorter than those detectable by other methods ($T = 3 \times 10^4$ years)—with one notable exception: Ninkovich *et al.* (*7*) report rather convincing evidence for two short polarity intervals ($T \sim 10^4$ years) at the end of the Gilsá event. Thus, while the paleomagnetic research on sediments has provided valuable information about the duration of the longer events and has helped extend the time scale back into the Gilbert reversed epoch, it has not resolved the question of the frequency of very short events.

Discussion

Over the interval $0.02 < t \leqslant 3.32$ million years ago (the part of the reversal time scale for which the number of short events is most completely known), a good fit to the observed frequency of polarity intervals (Fig. 7) is obtained from Eq. 2 when P is set

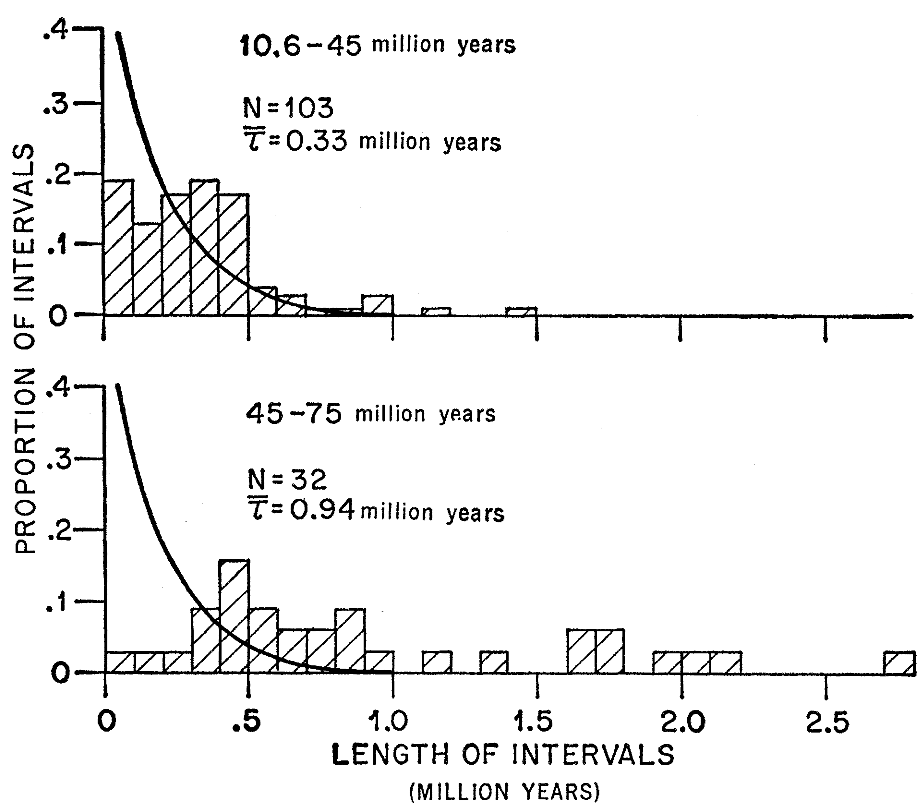


Fig. 8. Histograms of lengths of polarity intervals. Both histograms are from the time scale for intervals of Heirtzler *et al.* (*11*) based on magnetic profiles. Solid curves are as in Fig. 7.

equal to 0.055. The observed number of very small events is consistent with the number predicted by the model. Reversal times scales going back several tens of millions of years have been obtained from magnetic profiles on the assumption that the rate of sea-floor spreading was constant (*11*, *23*, *27*). The best fit of Eq. 2 to the data of Vine (*27*) for the past 10.6 million years is obtained with $P = 0.043$. The observed proportion of very short events ($T < 5 \times 10^4$ years) is slightly smaller than the proportion found for the interval $0.02 < t \leqslant 3.32$ million years; this suggests that, as had been suspected, some of the shortest events were not resolved on the magnetic profiles.

It is concluded from the analysis discussed here that, on the average, 20 fluctuations in intensity occur between successive polarity reversals. This conclusion, although based on a model in which dipole intensity fluctuations are periodic with a period of 10^4 years, would also be valid if the intensity changes were nonperiodic, with an average time between minima of 10^4 years. The corresponding probability that a geomagnetic reversal will result from the decrease in dipole moment currently in progress is 5 percent.

If sea-floor spreading has occurred at a constant rate, the marine magnetic profiles may be interpreted to yield a reversal time scale going back 75 million years (*11*). The apparent average duration of the polarity intervals was greater during the time $10.6 < t \leqslant 45$ million years than during the past 10.6 million years, and during the time $45 < t \leqslant 75$ million years the average length was still greater (Fig. 8). Part of the difference may be due to variations in spreading rate (*11*, *33*). However, this would account only for an apparent change in the probability P and not for the observed change in the shape of the distribution. The latter change may be more apparent than real, however, if, as is quite possible, a small number of additional short events occurred which have not yet been detected. Numerical experiments indicate that about 30 randomly distributed short events would change the observed distribution to one fitted by Eq. 2 when $P = .02$.

As one goes farther back in time, an average length of 10^7 years has been reported for polarity intervals in the early Paleozoic (*34*), and the Kiaman reversed-polarity interval during the late Paleozoic was 5×10^7 years long (*13*). Clearly, the average length of polarity intervals and hence the value of P have changed during the earth's history, reflecting changes in the physical conditions which control the geomagnetic dynamo. The possible importance to magnetohydrodynamic processes of changes in the properties of the core-mantle interface has been pointed out by Hide (*35*) and by Irving (*13*). These changes may result from large-scale tectonic processes, such as polar wandering accompanied by mass transport in the lower mantle. Hence it is conceivable that individual changes in polarity are produced by individual tectonic events. The present analysis suggests the alternative possibility that the timing of reversals is controlled on two levels. The *average* length of polarity intervals is controlled by conditions at the core-mantle interface and hence may be related to tectonic processes in the mantle, whereas the length of individual polarity intervals is determined by random processes in the fluid core.

Observationally, the main difference between the two interpretations is in the cutoff anticipated for short periods. In the present model, the cutoff is 10^4 years, the assumed period of fluctuations in the intensity of the dipole field. In Hide's model the cutoff is estimated to be from 10^5 to 10^6 years, depending on the strain rate of the lower mantle and the minimum displacement of the core-mantle interface needed to produce a change in polarity. With the discovery of polarity events at least as short as 3×10^4 years, it appears increasingly likely that the timing of individual reversals is controlled by processes occurring in the fluid core.

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Hallucinogens of Plant Origin

Interdisciplinary studies of plants sacred in primitive cultures yield results of academic and practical interest.

Richard Evans Schultes

An outstanding mark of this century will certainly be the growth in use, abuse, and misuse, in sophisticated cultures, of hallucinogenic substances of vegetal and synthetic origin (*1*).

Primitive cultures, where sickness and death are usually ascribed to a supernatural cause, have long accorded psychoactive plants a high, even sacred, rank in their magic, medical, and religious practices, because their ethnopharmacology often values the psychic effects of "medicine" more than the physiological. Ethnobotanical studies have recently advanced our understanding of known hallucinogenic plants and have uncovered new ones, some of which have yielded compounds (Fig. 1) of extraordinary chemical and pharmacological interest, even of promise in modern medicine (*1–5*).

The pace of research into hallucinogens in dying or disappearing primitive cultures; the success of studies into the plants and their constituents; and the confusion generated by casual or frivolous interests in some segments of our society—all seem to justify an ethnobotanical summary based upon the premise that, even though only an interdisciplinary consideration can adequately cope with this fast-growing field, the starting point for understanding naturally occurring psychoactive substances must be an appreciation of the identification and aboriginal significance of the plants involved. My own ethnobotanical research which, since 1936, has taken me into remote areas of the New World to study native narcotics, has convinced me that there exists an appreciable number of hallucinogenic plants still unknown to science, and that we can no longer afford to ignore reports of aboriginal uses merely because they fall beyond the limits of our credence. Primitive cultures are fast disappearing and, with them, native knowledge of plant properties that could help us along paths of academic and practical achievement.

In view of the number of plant species, variously estimated at from 400,000 to 800,000, those that have been used as hallucinogens are few; probably no more than 60 species of cryptogams and phanerogams. Only 20 may be considered important. Even more interesting is the unexplained concentration of the majority of the hallucinogens in the New World. In both hemispheres, there are plants with hallucinogenic properties which apparently have never been employed as narcotics, but even if these were added to those that man has bent to his purpose, the number is very small. Since most hallucinogens owe their activity to alkaloids and at least 5000 higher plants are alkaloidal, the small number of hallucinogens is even more astounding. Glycosides, resinoids, essential oils, and other organic constituents may also be responsible, so the limited number of hallucinogens must be considered challengingly provocative (*6*).

Hallucinogens occur nearly throughout the plant kingdom. Although most are spermatophytes, some of the biologically, chemically, and sociologically most fascinating are cryptogams.

Agaricaceae

The presence of toxic constituents in so many basidiomycetes led to the early discovery of hallucinogenic properties in the mushrooms. It has even been postulated that mushrooms were anciently and widely valued in primitive religions; that the very concept of deity arose from their effects; and that their present disjunct ritualistic use is relict.

The hallucinogenic use of the fly agaric (*Amanita muscaria*) by primitive tribesmen in Siberia came to the attention of Europeans in the 18th century. This fungus—widespread in north-temperate parts of both hemispheres—has long been recognized as toxic; its name refers to the European custom of employing it to poison flies. In recent times, its use as an inebriant has been known in only two centers: extreme western Siberia, among Finno-Ugrian peoples, the Ostyak and Vogul; and extreme northeastern Siberia, among the Chukchi, Koryak, and Kamchadal. Tradition established the use of fly agaric by witch doctors of the Lapps of Inari in Europe and of the Yakagir of northernmost Siberia. Formerly, the narcotic employment of *Amanita muscaria* was apparently more widespread, and it has even been

The author is executive director and curator of economic botany, Botanical Museum of Harvard University, Cambridge, Massachusetts. He is also lecturer in economic botany, Biology Department, Harvard University.