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Paleomagnetism and the Nature of the Geodynamo

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Records of direct observations of the earth's magnetic field cover less than a ten-millionth of the known lifetime of the field. Thus our knowledge of several geomagnetic phenomena, critical to our understanding of the geodynamo, must come from the paleomagnetic record. A combination of substantial advances during the past decade or so both in dynamo theory (previously the domain of the mathematician) and in paleomagnetism (previously the domain of the geologist) has led to provocative models of the earth's magnetic field and a better understanding of the geodynamo.

THE EARTH'S MAIN MAGNETIC FIELD IS NOW GENERALLY thought to be created by dynamo action in the earth's fluid outer core, and considerable effort has recently been devoted to obtaining useful constraints for the geodynamo problem. The areas of investigation fall into several broad categories: dynamo theory (including supercomputer simulations); predicting physical properties of the earth's interior; observations of magnetic fields in other planets and the sun; historical observations of the main field; and paleomagnetic observations. In this article we focus on some of the major areas where paleomagnetism is playing an important role in improving our understanding of the origin of the geomagnetic field. As such, it is only possible to provide a restricted view, and we have chosen to emphasize those areas associated with reversals of the field. This approach also dictates that only the briefest of comments be made on the areas other than paleomagnetism.

General Background

Not surprisingly, an analogy has been made (*1*) between the weather, largely determined by thermally driven fluid motions in the atmosphere, and the magnetic field. For example, phenomena such as fluid eddies, planetary (Rossby) waves, and thermal winds are invoked to account for various observations in both regimes. The large variation in the boundary conditions at the earth's surface significantly affects the earth's meteorology; similarly, horizontal variations at the core-mantle boundary, perhaps second only to the diversity at the earth's surface (*2*), almost certainly have a significant effect on the earth's magnetic field. Of course, the earth's two largest fluid bodies have important differences, but the analogy is sufficient to indicate that it is a formidable task to model the geodynamo. Moreover, although the atmosphere can be observed directly, the physical properties of the outer core, its boundary conditions, and its internal motions producing the magnetic field can only be inferred indirectly.

A complete mathematical solution to the geodynamo problem must simultaneously satisfy Ohm's law, the Maxwell, Navier-Stokes, Poisson, continuity, and generalized heat equations, together with the equation of state for the outer core and the appropriate boundary conditions (*3*). Most of these are partial differential equations; they are coupled; and there are strong nonlinearities. Solution would present an enormous problem even if the equation of state and boundary conditions were well known, but they are not. Because of the complexity of the problem, there are numerous partial models, each of which incorporates simplifying assumptions and restrictions and many of which use poorly known parameters.

Dynamo theory (*4*) weathered an early setback in the form of Cowling's theorem (*3, 4*) (which effectively states that axisymmetric magnetic fields of any sort cannot be maintained by dynamo action) and consequent fears that there was a general antidynamo theorem. It is now recognized that the geomagnetic field departs significantly from axial symmetry and that a magnetic field can be generated by almost any fluid motion that is sufficiently vigorous and complicated.

The magnetic induction equation (from Maxwell's equations and Ohm's law) is central to dynamo theory:

$$\frac{\partial \mathbf{H}}{\partial t} = k\nabla^2\mathbf{H} + \nabla \times (\mathbf{U} \times \mathbf{H}) \quad (1)$$

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where **H** is the magnetic field, t is time, **U** is velocity, and k is magnetic diffusivity (the inverse of the product of the free-air permeability and the electrical conductivity). The first term on the right side indicates that in the absence of dynamo action the magnetic field would simply decay with time, and estimates for the free decay (that is, $\mathbf{U} = 0$) time are $\approx 1.5 \times 10^4$ years (*5*). The second term on the right side represents the interaction of the velocity field with the magnetic field (that is, dynamo action) and can cause either breakdown or buildup of the magnetic field. Because there has been a magnetic field for most of the earth's 4.5-billion-year history (*3*), the first term cannot dominate on the average. Briefly, dynamo action is the conversion of mechanical energy associated with the fluid motion to magnetic energy.

Studies of the earth's interior have yielded critical information regarding the core structure and material properties (*6*). The outer liquid core, extending from a depth of 2881 to 5150 km, consists primarily of Fe alloyed with small amounts of Ni and lighter elements (S, O, and H are the most commonly suggested, less dense elements; these elements are needed to match observed seismic velocities with velocities in materials measured in the laboratory at core pressures). The inner core–outer core boundary is believed to be at the melting temperature of the inner core, and hence the inner core is believed to be mostly solid (although its outer part may be a mush), extending to the earth's center (at a depth of 6371 km). As the earth cools, growth of the inner core by freezing preferentially extracts Fe from the outer core. As a consequence, liquid enriched in the less dense elements is released from the inner core–outer core boundary. The chemical buoyancy of this liquid is currently thought to be the most likely energy source for driving convection and, hence, the geodynamo (*7*). Because neither the density contrast across the inner core–outer core boundary nor the chemistry of the core is well determined (*8*), there is still some uncertainty concerning the primary energy source.

The temperatures throughout the core are also not well known, primarily because of errors associated with high-pressure measurements of Fe and uncertainties about which lighter elements are present. Most investigators would place the temperature of the core-mantle boundary between 3500° and 5000°C.

The viscosity of the fluid in the outer core is critical to the fluid's behavior. Poirier (*9*) has recently reviewed the published experimental data and concluded that the viscosity, at least near the inner core–outer core boundary, is probably close to that of liquid Fe at atmospheric pressure. Thus it would appear that the viscosity is (perhaps surprisingly) only a few times that of water at the earth's surface. This suggests that the fluid motions in the core might be very complex and provides some support for advocates of the statistical mechanical approach embedded in so-called mean-field electrodynamic dynamo models.

The relevance of studies of the magnetic fields of the sun and planets is a matter of judgment. Each object in the solar system is different from the other objects in some way, and therefore one can always question whether a property observed for one planet has the same origin as a similar property on another planet. This situation is further complicated because on the earth we can observe only the poloidal field (the part of the field that has a radial component); the mantle acts as a shield for the toroidal field (the part of the field that does not have a radial component). The strength of the earth's toroidal field is unknown, but in the currently popular strong-field models (ones in which the magnetic field in the core significantly affects the fluid velocity) the toroidal field strongly dominates the poloidal field in the core. Thus it may be that we can observe at most a relatively small part of the field; certainly we cannot observe all of it. Notwithstanding such problems, observation of the solar magnetic field does not provide three clear pieces of relevant information. First, it is possible to generate a long-lived magnetic field with a self-sustaining dynamo. Second, a self-sustaining dynamo in which the toroidal field dominates the poloidal magnetic field is quite feasible. Third, the quasi-periodic (every 11 years) reversal of the solar magnetic field shows that such a dynamo is capable of internally generated reversals.

Geomagnetic Observations

Since the early part of the past century direct measurements of the intensity and direction of the geomagnetic field have been made at a sufficient number of locations on the earth's surface to allow geomagnetists to characterize the field, and some of its short-term variations, using spherical harmonics. The spherical harmonic representation is a convenient mathematical description that, nonuniquely and artificially, places all "sources" at the earth's center. Roughly 90% of the magnetic field at the earth's surface can be described by a geocentric dipole (essentially a bar magnet) tilted 11° with respect to the rotation axis. If the field were entirely dipolar, the geomagnetic poles (the extension of the best fitting dipole) and the magnetic poles (where the inclination of the magnetic field is vertical) would coincide, and they would be 11° away from the geographic poles. However, roughly 10% of the magnetic field is nondipolar at the earth's surface. Indeed, if the dipolar part of the field is subtracted from the main field, the result is a complex field consisting of several magnetic foci (where the field is vertical). Bullard *et al.* (*10*) showed that, in a gross sense, this nondipolar field has been drifting westward at a rate of about 0.18° per year. Bloxham and Gubbins (*11*) traced mariners' past measurements of the field to construct maps that characterize the magnetic field back into the early 18th century. They emphasized that for dynamo theory it is more important to have knowledge of the magnetic field at the core-mantle boundary than at the earth's surface: a difficult task, given the problems in achieving a reliable downward continuation of the field from the earth's surface, where it is measured. Although the dipole term is still the largest single term at the core-mantle boundary, the total nondipole field dominates the dipole field there. This was known before the analyses of Bloxham and Gubbins, but they also presented good evidence that the apparent westward drift of the nondipole field varies considerably from one location to another and that, at least during the past ~300 years, most of the drift at the core-mantle boundary occurred in the Atlantic hemisphere, as opposed to the Pacific hemisphere.

The dipole field also changes with time but at least during recent times at a much slower rate than the nondipole field. During the past 150 years there has been a westward drift of about 0.05° per year in longitude but no progressive motion in latitude of the overall dipole orientation. The moment of the geocentric axial dipole has been decreasing at about 0.05% per year. This decrease should not necessarily be taken as an indication of an impending reversal, however, for archeomagnetic results indicate that for the past ~3000 years the moment has been higher than the long-term average.

Paleomagnetism

In the final analysis, any geodynamo model must be tested against observation. Detailed geomagnetic observations have yielded a wealth of valuable information, but the historical record is barely three centuries long, whereas the geomagnetic field has existed for at least 3 billion years (*3*). Thus it is hardly surprising that many critical geomagnetic phenomena have time constants at least an order of magnitude longer than the records of direct observations (for example, the dominant period in most secular variation records is of order 10^3 years) and that some phenomena, such as reversals, have not even occurred within the historical record. Therefore, it falls to paleomagnetism to establish the existence of such phenomena, to elucidate their details, and to play a critical role both in stimulating and in testing geodynamo models.

Unlike most geophysical techniques, such as seismology, that provide only instantaneous pictures of the earth as it is now, paleomagnetism provides information on the history and evolution of the earth in general and of the earth's magnetic field in particular. Obtaining reliable results that are useful in the context of modeling the geodynamo is not, however, an easy matter.

When a rock forms, it typically acquires a primary magnetization that is parallel to the external field at the time of formation. Providing the rock can be dated and providing the primary magnetization can be distinguished from secondary magnetizations (magnetizations acquired after the rock formed), valuable information can be obtained on the direction and (in some samples) on the intensity of the ancient magnetic field. The nature of the record varies considerably depending on the recording medium. For example, in a typical sample of marine sediment, the result represents some average value of the magnetic field over a time span of a few hundred to several thousand years, depending on the sedimentation rate. In contrast, a thin lava flow may acquire all its primary magnetization in less than a year. As with many field-based sciences, there are numerous sampling problems (for example, the declination information is lost in most deep-sea cores). In addition, there can be substantial rock magnetic problems that lead to significant errors. This problem can be severe, and it is often difficult to document (*12*). An interesting example comes from a carefully performed study of a 15.5-million-year-old polarity transition in basalt flows in Oregon (*13*). If the observations are to be believed as actual representations of the field behavior, then polarity transitions exhibit astonishing properties, including that the earth's magnetic field changed direction by a few tens of degrees in less than a year and underwent rapid changes in intensity. These changes are difficult to accept, because mantle shielding should smooth out such rapid changes and because the changes seem to require unrealistically high fluid velocities in the core. It is easier to believe that there was some as yet undetected complication in the paleomagnetic recorder, such as a chemical remanent magnetization overprint, than to accept that the data are accurate recordings of the earth's magnetic field. We must await the outcome of further study of these flows (which is continuing) before determining the relevance of these observations.

In pursuing the quest of a realistic geodynamo model, one must recognize that the quality of the paleomagnetic record will vary considerably and that the data will be unevenly distributed both in space and in time, will occasionally be in error, and will typically represent some filtered picture of the magnetic field. Success will therefore depend on the availability of very many paleomagnetic data and careful statistical analyses.

Polarity Reversals

An idea which initially appeared to be outlandish, but which has subsequently been incorporated into scientific dogma (*14*), is that the magnetic field reverses polarity (that is, the geomagnetic north and south poles change places). The present field is referred to as having normal polarity, and the opposite polarity state is referred to as having reverse polarity. The reversal chronology is reasonably well established for the past 100 million years and less well established for the preceding 100 million years, and there is only limited and spotty information for times before this. Even in the well-established Cenozoic (the past 66 million years) record, however, there are almost certainly instances when two reversals were sufficiently close (a few to several thousand years) that the short period of opposite polarity they bracket (a polarity interval) has not been identified.

Valuable information on processes acting deep within the earth's interior has already been obtained from analyses of the reversal chronology record. Unlike the quasi-periodic reversals of the sun, the time between geomagnetic reversals appears to have a strong stochastic component (*15, 16*), and the process is essentially Poisson (*16*). Although the present reversal rate is quite high (about 4.5 reversals per million years), this has not always been the case; the Cretaceous normal superchron [from ~83 to ~118 Ma (million years ago)] represents one of two well-documented long quiet intervals during which the reversal process was apparently in abeyance and no reversals occurred. From ~83 to ~12 Ma, the mean reversal rate increased steadily, but it now may be decreasing again (*16*); the overall increase is robust and is not sensitive to the details of the chronology, which will almost certainly be modified with future work (*17*). The increase in reversal rate after the Cretaceous superchron is nearly mirrored by a steady decrease before the superchron (see Fig. 1). If the time from the recent peak in reversal rate (about 12 Ma) to the center of the Cretaceous superchron (about 100 Ma) is doubled, a total interval of roughly 200 million years is obtained. The Kiaman reverse superchron (roughly 50 million years) occurred about 200 million years before the Cretaceous superchron. This observation might suggest a quasi-periodic process, but at present the data are inadequate to demonstrate the presence or absence of a superchron at roughly 500 Ma. There has been some speculation that a third quiet zone, the Jurassic quiet zone, may fall between the Cretaceous and Kiaman superchrons. However, recent paleomagnetic and magnetic anomaly evidence now suggests that it is likely an artifact (*18*).

There have been several claims of a small-amplitude periodicity (period about 15 or 30 million years, depending on the method of analysis) superimposed on the mainly stochastic reversal record, but no robust statistical analysis has verified this. A particular problem with such verification is that minor revisions in the reversal chronology record can significantly affect apparent periodicities.

Because of the low viscosity of the outer core, the roughly 100-million-year variation may reasonably reflect the time scale for changes in the boundary conditions at the core-mantle boundary (*16*). Variations in the temperature gradient across the outer core

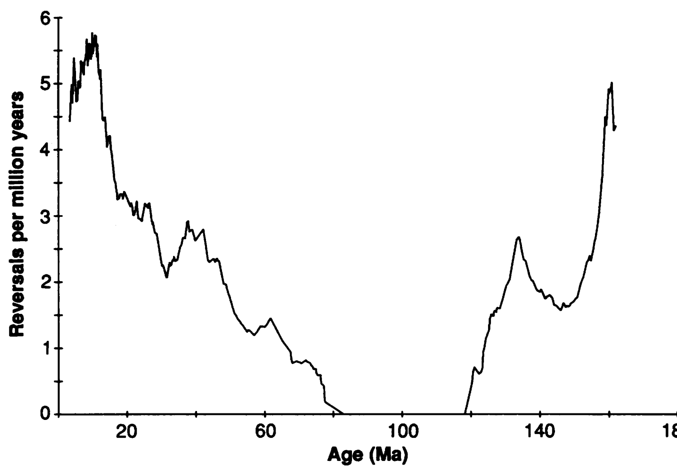

Fig. 1. Estimated mean reversal rate from the present back to 165 Ma.

can critically alter core convection (and therefore the magnetic field). Stacey and Loper (*19*) pointed out that the thermal history of the core is primarily determined by the cooling of the mantle. In particular, the rate at which heat is transferred from the core to the mantle is determined by the temperature gradient at the base of the mantle, a gradient that is still not well known but that has usually been estimated to be ~3° to ~10°C per kilometer (*20*). It is also possible that the lateral variations in the temperature gradient at the base of the mantle, variations expected from the lateral heterogeneity (in topography and/or chemistry) present there (*2*), change on a 100-million-year time scale in such a way as to produce a significant alteration in the core's convection.

The time scale of 100 million years is also approximately the characteristic time scale for mantle convection (there is too much uncertainty in various parameters to distinguish between layered mantle convection and whole-mantle convection). Consequently, heat transfer in the mantle may also have a long-term effect on the reversal rate (*16*). Because tectonic processes at the earth's surface are directly related to heat transfer in the mantle, one can speculate on the possibility of correlations (with appropriate lags) between changes in reversal rate and other geological phenomena such as true polar wandering and volcanism, as has been recently done in a provocative study by Courtillot and Besse (*21*).

The distribution of the lengths of polarity intervals during the Cenozoic is consistent with a nonstationary gamma process (*16*). In particular, it is consistent with a nonstationary Poisson process, which is a special case of a gamma process in that there is no memory of how long it has been since the previous event. This result implies that a reversal neither inhibits nor encourages future reversals, and this has been the situation right through from the end of the Cretaceous superchron (when the reversal process was in abeyance) to the present. Thus the long-term change in reversal rate must be caused by change in the inherent rate of the process, not by any change in inhibition, and this provides an important constraint on the reversal process (*22*).

Muller and Morris (*23*) have argued that external sources could produce sudden changes in mass distribution at the earth's surface, which in turn could produce change in shear at the core-mantle boundary and thereby lead to reversals of the magnetic field. Such a sudden change in mass distribution could occur, say, if the impact of an extraterrestrial object (a bolide) led to a sudden climatic change and a redistribution of water (for example, from ice caps) at the earth's surface. They cite a range of evidence supporting such a model, the most convincing of which is based on data from marine sediments. However, there are numerous pitfalls (such as gaps in the record) in interpreting marine stratigraphy, and thus the evidence should be viewed cautiously. For example, some of the data come from Glass *et al.* (*24*), who pointed out that there are four well-documented tektite fields (indicating bolide impacts) in the Cenozoic, three of which appear to coincide with reversals. However, only two of the four are reasonably well placed in the magnetostratigraphic record, and, although both of these had previously been thought to coincide with reversals, more recent detailed analysis indicates that one does not (*25*). Important geological evidence also seems to be inconsistent with the model. For example, at the Cretaceous-Tertiary boundary (66 Ma), where a major catastrophic change in climate has been attributed to a bolide impact or voluminous volcanic eruptions (*26*), there is no evidence of a reversal, aborted reversal (excursion), or any change in reversal rate (*27*). Another example involves the ice ages. The past few million years is only one of the documented major periods of ice ages in the geological record and coincides with a time of rapid reversal rate, consistent with the Muller and Morris model. In contrast, the next best documented ice age occurs in the Paleozoic and significantly overlaps the Kiaman interval in which there were no, or at least only a few, reversals (*27*). A further problem is that the qualitative mechanism for reversal suggested by Muller and Morris does not appear to work when quantified (*28*). Considering such arguments, together with the observation that other dynamos (for example, the solar dynamo) can reverse in the absence of large surface mass redistributions, we conclude that reversals are most likely of internal origin.

Roberts and Stix showed that dynamo fields are made up of two families and that for certain symmetry conditions in the core these families are noninteractive (*29*); in a somewhat oversimplified sense, it is as if there were two independent dynamos operating in the core. Furthermore, they showed that when these symmetry conditions are violated, the two dynamo families will interact. We call these the primary dynamo family (previously referred to as the dipole family and characterized by spherical harmonics whose order and degree sum to an odd integer) and the secondary dynamo family (previously referred to as the quadrupole family and characterized by spherical harmonics whose order and degree sum to an even integer). An example of the integration of theory with analyses of paleomagnetic data is the recent suggestion that there is an increased chance for reversals when the ratio of the magnitude of the secondary dynamo family to that of the primary dynamo family is high (*27*). Paleomagnetic secular variation data can be used to estimate the relative magnitudes of these two families (*30*). On the basis of analyses of such data and other paleomagnetic data, Merrill and McFadden suggested that reversals are more frequent when the magnitude of the secondary family relative to that of the primary family increases; this increase in reversal rate presumably reflects the increased interactions between the two dynamo families (*27*). This reversal model predicts that the secondary family should have been relatively low during the Cretaceous superchron; preliminary analyses support this prediction (*31*).

Even if the above model is sustained by future studies, it is of course not a complete model for reversals, because it does not explain what causes changes in the interactions between the two dynamo families. Such changes could, say, occur through (deterministic) chaotic fluid motions that are essentially confined to the interior of the outer core, or, at the other extreme, the instabilities leading to increased interactions between the two families could originate essentially from instabilities associated with the solid boundaries of the fluid outer core.

Polarity Transitions

The modern era of the study of polarity transitions was probably ushered in by Hoffman and Fuller about a decade ago when they collected, categorized, and analyzed polarity transition data (*32*). The data come from a variety of sources, including marine and terrestrial sediment and extrusive and (rarely) intrusive igneous rocks.

A fundamental problem in characterizing polarity transitions is that even a perfect, continuous record would only represent information from a single point on the globe. Because of the existence of the complex nondipole field, there is, in principle, no a priori reason why the strength of the field should not actually be increasing at some points on the globe at the onset of a reversal, even though the field strength would, on average, be decreasing. Furthermore, extremely simple models that exhibit distressing properties can be constructed. For example, in a model with just a zonal dipole (that gradually reduces its strength and then builds up in the opposite direction), a zonal quadrupole, and a small nonzonal field, both the time at which the directional transition appears to occur and the time required for the transition depend on the geographic location of the observer. Thus, without a good global distribution of records for a single transition, together with accurate and precise temporal correlation of those records, it is difficult to elucidate the details of a transition. Furthermore, different transitions are probably quite different in detail but have certain gross properties in common. Time-varying spherical harmonic coefficients have been used to model individual records of polarity transitions. We believe a simpler approach is needed. For example, if one of the two dynamo families discussed in the previous section dominates during the middle of transitions, then, by investigating whether the transitional field is predominantly symmetric or antisymmetric with respect to the equator, one should (in time) be able to determine which one it is. This approach would then provide a powerful constraint on the geodynamo problem.

As usual, because of the problems associated with paleointensity data and their interpretation, attention is usually focused on directional data. However, because of the lowered intensities, the relative magnitude of secondary magnetizations (acquired later) will be statistically greater for transition records than for other times, so that even the directional information for transition times is less reliable than usual. Thus the major goal at this stage should be to identify gross properties and systematic changes in reversal transitions rather than to attempt to interpret detail.

Certain such gross properties are already apparent and do provide powerful information about the geodynamo. Polarity transitions do not appear to be characterized by a simple rotation of the dipole field through 180°, and, conversely, the intensity of the field does not appear to vanish (the minimum value reached is generally believed to be about 10 to 20% of the mean field intensity during nontransition times). Instead, in the middle of the polarity transition, the magnetic field appears to be dominated by nondipole components. This observation, free-decay time estimates, and related phenomena suggest that the polarity transition is dynamic throughout (*5*); that is, the decrease in intensity during the first part of the transition is caused not primarily by free decay but by a reorganization of the fluid motions such that the magnetic field is actually broken down dynamically (through the last term in the magnetic induction equation). That the field intensity becomes small and then rebuilds during each reversal transition (and there have been many of them) is a clear indication that the second term on the right side of Eq. 1 must be able to dominate the first term. The average duration of changes in direction associated with a "typical" reversal is ~4000 years, and, although far less well documented, the time associated with intensity changes during a transition seems to be $\sim 10^4$ years (*33*).

Polarity Asymmetries

By averaging paleomagnetic observations distributed through time, one can estimate the time-averaged field for a given region. Such averages distributed in space then give an estimate of the overall time-averaged field. When this analysis is performed for the past 5 million years (*34*), the time-averaged field appears to be essentially axially symmetric; Cowling's theorem about axisymmetric fields does not rule out an axisymmetric time average, so this observation is not evidence against a dynamo as the field source. Of particular interest is the observation that the time average is not simply the field of a geocentric axial dipole (*34*); one requires at least a persistent geocentric axial quadrupole and octupole to explain the data (*35*).

Perhaps the most interesting aspect of the time-averaged field for the past 5 million years is that there appears to be asymmetry between the normal and reversed polarity states (*32, 35, 36*). This difference is best evidenced in the geocentric axial quadrupole field. The governing equations for the geodynamo are symmetric with respect to the field direction, and thus the asymmetry must be a consequence of boundary or initial conditions (*35*). The asymmetry is more likely a consequence of boundary conditions than initial conditions, and therefore the polarity asymmetry is most likely associated with lateral variations in material properties at the base of the mantle (*35*). Analyses of changes in polarity asymmetry farther back in time could provide valuable information on "lower mantle tectonics."

Excursions

There is an increasing amount of evidence for the occurrence of magnetic field excursions and very short reversal events (*37*). Excursions occur when the magnetic field deviates significantly (that is, 45° or more) from that of a geocentric axial dipole field, and evidently they can give rise to local apparent reversals of the magnetic field. Substantial work is still required to determine how common excursions are and whether they are aborted reversals, as often assumed, or some other core phenomenon.

Secular Variation

Secular variation refers to change in the magnetic field over time intervals of a few years to a few thousand years. Thus it spans the domains of direct, archeomagnetic, and paleomagnetic observation. As such, it appears to be an intimate part of the geodynamo process, and substantial theoretical effort has been devoted to understanding the phenomenon [see (*3*) and (*4*) for further information]. Whether the changes present in the short historical record are typical is still uncertain, even though changes with similar "periods" are clearly manifested in paleomagnetic data. Considerable effort has been expended (particularly through analysis of lake sediments) to characterizing the longer term secular variation, and the data base is rapidly reaching the stage where inferences can be drawn about the geodynamo. For example, Olson and Hagee (*38*), on the basis of theoretical considerations, argued that magnetic changes observed in lake sediments are indicative of poleward propagating waves in the geodynamo.

Intensity of the Field

Considerable controversy remains over the reliability of absolute paleointensity measurements (*39*), and therefore the full vector information nominally contained in the paleomagnetic record is still unavailable. Paleointensity information is particularly needed during certain critical times in the geological record. For example, given the nonstationarity of the reversal process during the past 100 million years, there should be some difference in mean intensity during the Cretaceous superchron relative to the mean for recent times, but the data appear too few and are of too poor quality to determine if this is the case, let alone what that difference might be (*40*). Reliable evidence regarding the field intensity during polarity transitions is also needed.

Notwithstanding the uncertainties, it has been possible to arrive at some general conclusions. The intensity of the dipole field is far from constant, and it has varied by about ±40% during the past 10,000 years (*3*). The field strength during a polarity transition is thought to drop to about 10 to 20% of the mean field intensity during nontransition times. Also, the overall distribution of paleointensities has been used to argue that nonlinear processes are being observed in the field generation (*5*); such processes suggest that the toroidal field is strong enough that it affects the fluid velocity.

Conclusions

With the improvement of both dynamo theory and the paleomagnetic data base, results from these two areas are now being combined to provide realistic constraints for the geodynamo problem. Interpretations of paleomagnetic data range from those that are now beyond reasonable doubt to those that appear to fly in the face of conventional theory. For example, reversals of the geomagnetic field and the nonstationarity of the reversal chronology (but not the precise character of the nonstationarity) are now well-established results. In contrast, hypotheses that reversals are associated with increased interactions between the primary and secondary dynamo families, or that there are no significant times of convection stasis during a polarity transition, or that there are discernible asymmetries between the polarities, should be regarded as stimulating, but far from proven, models that appear compatible with available paleomagnetic data and dynamo theory. Paleomagnetic data from lavas at one location in the western United States may imply that large changes in direction and intensity of the geomagnetic field have occurred in less than a year's time during a polarity transition, an interpretation that appears incompatible with conventional wisdom with respect to the properties of the earth's mantle and processes acting in the outer core.

The examples presented here provide only a restricted view of the contributions paleomagnetism is making to our understanding of the history and origin of the earth's magnetic field. Undoubtedly some of the hypotheses will prove to be inaccurate or in error, because of factors such as improper rock magnetic or statistical analyses. If the past provides any indication on how to interpret the future, it seems relatively safe to predict that at least some of the apparently more outlandish paleomagnetic interpretations of today will be taught as scientific fact to students of the 21st century.

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