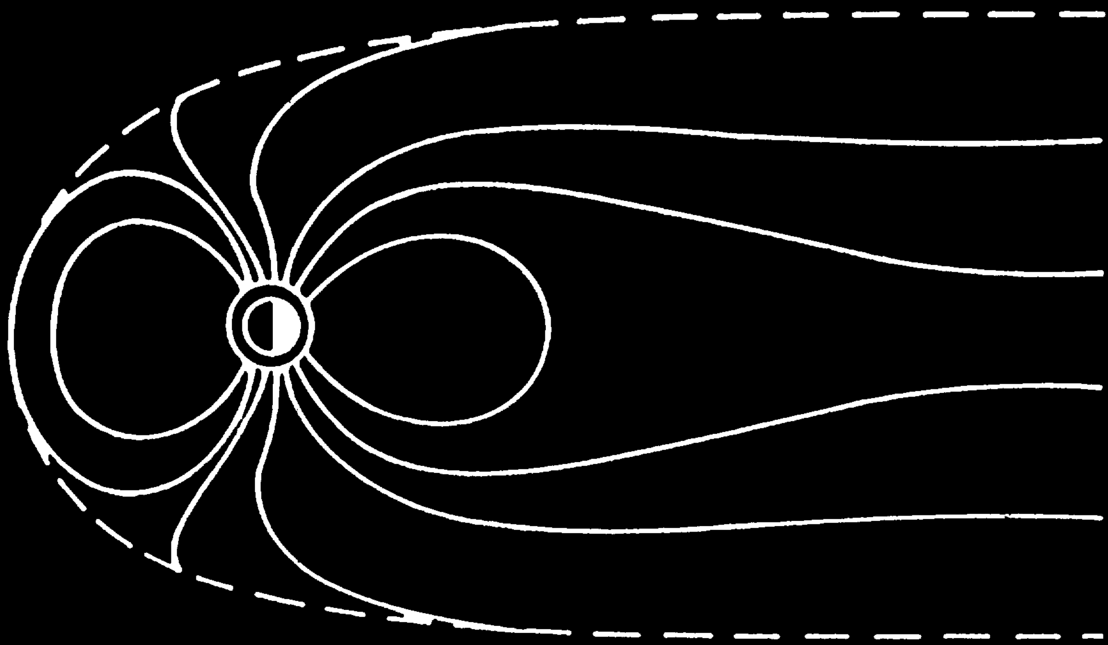
The Scientific Case For Magnetic Field Satellites



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THE SCIENTIFIC CASE FOR MAGNETIC FIELD SATELLITES

1. Introduction

Chemically or thermally regenerated density differences in the earth's electrically conducting fluid outer core drive a motion there which sustains the main geomagnetic field by dynamo action. This dynamo, together with the possibly corrugated core- mantle boundary (CMB) and the electrically conducting lower mantle constitute a physical system, the geodynamo system, of great intrinsic interest and of considerable utility in monitoring and understanding other earth subsystems. For example, it has been suggested that the reversal frequency of the main geomagnetic field (currently about six reversals per million years) may vary with and indicate the thickness of the thermal boundary layer at the base of the lower mantle, thus giving a historical record of the convective phenomena in the lower mantle which produce hot spots and contribute to continental drift. It is also conceivable, though speculative, that the main geomagnetic field affects electrical conditions in the upper atmosphere, where the weather-generating thunderstorm electric circuit closes to the ionosphere. Such an effect could influence climate.

To make full use of modern magnetic data and the paleomagnetic record, we must greatly improve our understanding of how the geodynamo system works. It is clearly nonlinear, probably chaotic, and its dimensionless parameters cannot yet be reproduced on a laboratory scale. It is accessible only to theory and to measurements made at and above the earth's surface. These measurements include essentially all geophysical types. Gravity and seismology give evidence for undulations in the CMB and for temperature variations in the lower mantle which can affect core convection and hence the dynamo. VLBI measurements of the variations in the Chandler wobble and length of day are affected by, among other things, the electromagnetic and mechanical transfer of angular momentum across the CMB. Finally, measurements of the vector magnetic field $\underline{B}$, its intensity B, or its direction $\hat{\underline{B}}=\underline{B}/B$, give the most direct access to the core dynamo and the electrical conductivity of the lower mantle. The 120 gauss coefficients of degrees up to 10 probably come from the core, with only modest interference by mantle conductivity and crustal magnetization. By contrast, only three angular accelerations enter the problem of angular momentum transfer across the CMB.

Satellite measurements of $\underline{B}$ are uniquely able to provide the spatial coverage required for extrapolation to the CMB, and to isolate and measure certain magnetic signals which to the student of the geodynamo represent noise, but which are of great interest elsewhere in geophysics. The body of this report will justify these claims and describe the mission parameters likely to be scientifically most useful for observing the geodynamo system.

2. Time Scales and Length Scales in the Geomagnetic Field

As detailed in Appendix I, there are interesting signals in $\underline{B}$ at all periods from 10^9 years to a few minutes. Magnetic signals from the core cannot reach the earth's surface through the electrically conducting mantle if their periods are less than one year. However, observations on a time scale of minutes are of interest as probes of the electrical conductivity of the lower mantle because of an interesting and severe aliasing problem. Magnetic storms have onset times of a few minutes and can last several days. They can alter $\underline{B}$ by several percent at the earth's surface. Their time-scale is too short for penetration to the lower mantle. However, these storms result from responses of the magnetosphere to solar events, so their frequency of occurrence and their strength are modulated by the eleven-year sunspot cycle. Good spatial coverage of the storms with a temporal resolution of minutes permits integrating the storm signals and accurately estimating the amplitude and phase of the external eleven-year sinusoid which they apply to the earth. The internal response of the mantle at eleven years probably comes from its deepest part, and furnishes a direct estimate of the conductivity there. This scheme requires both satellite data and data from magnetic observatories, since the former lack the required time resolution and the latter are defective in spatial coverage.

Even the higher estimates of electrical conductivity in the mantle would permit passage of magnetic signals from the core to the surface of the earth if the periods were longer than ten years. Moreover, many workers now believe there is direct evidence that magnetic core events (the impulses or jerks) have been detected which last only a year or two and occur at random intervals. Thus it is essential to have accurate measurements of $\underline{B}$ over several decades, and aliasing prevents us from obtaining such data by taking a snapshot of $\underline{B}$ with a short satellite mission every ten years. The monitoring must be at least quasi-continuous if we are to use rapid random events like jerks and magnetic storms to study mantle conductivity and core fluid motion.

The spatial scales of interest in $\underline{B}$ extend from the forty-thousand km circumference of the earth down to the few meters of a magnetite outcrop. There is good evidence, however, that at the earth's surface the magnetic signals with wavelengths shorter than 3000 km (spherical harmonic degree $n > 14$) come from crustal magnetization, not the geodynamo system, and that only surface signals with wavelengths longer than 4000 km ($n < 10$) do not suffer serious crustal contamination. In order to extrapolate these signals down to the CMB, good spatial coverage is essential, and aliassing from the crustal signal must be removed. Magnetic observatories provide neither good spatial coverage nor protection from spatial aliasing. Satellites in nearly polar orbits with altitudes between 200 and 1000 km can provide good coverage and also will geometrically attenuate the crustal signals to the point where digital filters can isolate them for study or removal.

3. Geomagnetic Field Sources Outside the Geodynamo System

The magnetic signals from the crust, the ionosphere and the magnetosphere are interesting in themselves and must be understood and removed from $\underline{B}$ if the geodynamo signal is to be isolated. Concerning the crust, some workers believe that the MAGSAT data provide reliable internal gauss coefficients up to degree 50. If they are correct, then the internal gauss coefficients of degrees 14 to 50 give the magnetic field produced by the crust at wavelengths between 800 km and 2800 km on the earth's surface. This information is unobtainable except by satellite, and the extent to which it is contaminated by magnetospheric signals will not be known until satellite monitoring has covered at least one eleven-year sunspot cycle.

The ionosphere supports a current system almost steady relative to the sun, although there is a daily wobble produced by the eleven degree deviation between the magnetic dipole and the axis of earth rotation, and an annual term produced by the inclination of that axis to the ecliptic. This current system includes longitudinal jets at the equator and in the auroral zones and a tidal current with a strong longitudinal variation. The earth rotates past these quasisteady systems, and sees them as external signals. Their axisymmetric part looks steady from the earth, and their longitudinally varying part has a 24 hour period when seen from the earth. Satellites are above the ionosphere, so in the gaussian resolution of $\underline{B}$ they see the ionospheric magnetic field as internally generated. Observatories see it as externally generated. If both types of data are available, the ionospheric signal can be isolated unambiguously. Neither data set by itself can pick out the axisymmetric part of the ionospheric signal.

The magnetosphere constitutes a major challenge for satellite observation of the geodynamo system. At the altitudes well above 1000 km, where magnetospheric satellites collect data, those details of the core field which do manage to reach the earth's surface are lost to geometric attenuation. Moreover, observations of the magnetosphere are intermittent in space and time, being designed to understand magnetospheric physics rather than to follow the magnetosphere in detail. If a qualitatively faithful model of the magnetosphere can be developed which permits description by means of a few parameters, then satellite observations below 1000 km will permit evaluation of those parameters and separation of the magnetospheric signal from the geodynamo. Otherwise, we will be forced to parametrize the low altitude magnetospheric signal empirically, a venture whose success is still uncertain; or to discard observations of horizontal $\underline{B}$ at high latitudes, as was done for MAGSAT.

4. Conclusions

By the first half of the 21st century, permanent monitoring of the geomagnetic field by several satellites will probably be routine, so that randomly occurring core events like the 1969 jerk will be available for scrutiny, and the geomagnetic time series will be long enough both to exploit the sunspot cycle and to remove it as a noise source. If the present generation of geophysicists is to have even a glimpse of these results, it is urgent to begin satellite monitoring of the geomagnetic field as soon as possible. Missions by several different countries and agencies, each living several years, would seem to be the scientifically optimal procedure, given today's financial constraints.

Our recommendations for those satellite mission parameters which directly affect the science are as follows:

A. Accuracy:
- Scalar: 2.0 nanoTesla (nT) rss design spec., 1.0 nT goal.
- Vector: 5.0 nT rss design spec., 3.0 nT goal.
- Discontinuities due to attitude solution to be less than 2 nT, absolute.
- Errors include all sources: instrument, position, spacecraft contamination, attitude determination, etc.

B. Data Rate:

- 1 Sample/second is sufficient for internal fields.
- 16 Samples/second for the vector measurements is desirable for field aligned current measurements.
- Data readouts should be synchronized between the two instruments and, if possible, with the altitude measurements.

C. Geographic Coverage:

- Data spacing with no more than 1000 km between points every 4 days.
- Data spacing with no more than 60 km between points every 3 months.

D. Local Time Coverage:

- All local times sampled at least once every 6 months.

E. Orbit:

- Altitude between 300 and 1000 km.
- Inclination: between 85 degrees and 95 degrees as needed for C.

Accuracy: depends upon error budget to achieve A.
(See Langel, 1976, Effects of Orbit Error on Satellite Magnetic Field Experiments, GSFC Document x-922-76-124).

F. Lifetime:

- Minimum 18 months.

5. Appendix I. Description of the Geomagnetic Field

A. Gauss's Representation of the Field

Neither displacement currents nor electrical conduction currents in the earth's atmosphere have a measurable effect on the geomagnetic field $\underline{B}$. Therefore, Maxwell's equations imply the existence in the atmosphere of a scalar magnetic potential such that

$$\underline{\mathbf{B}} = -\underline{\nabla}\,\psi \tag{1}$$

Maxwell's equations also imply that $\nabla^2\psi = 0$, so that at each instance t, at each position r, θ, ϕ in the atmosphere (r is radial distance from the center of the earth, θ is colatitude, ϕ is longitude) ψ can be written:

$$\psi\ (r, \theta, \phi, t) = a\Sigma_{n=1}^{\infty} \Sigma_{m=-n}^{n}[g_n^m(t)\,(a/r)^{n+1} + q_n^m(t)\,(r/a)^n]\ Y_n^m(\theta, \phi) \tag{2}$$

Here a is the radius of the earth, $Y_n^m(\theta, \phi)$ is a surface spherical harmonic of degree n and longitudinal order m, and g_n^m q_n^m are functions of time alone. As far as Maxwell's equations are concerned, g_n^m and q_n^m can be chosen arbitrarily. They are called the Gauss coefficients of $\underline{B}$ relative to Y_n^m.

If ψ_g and ψ_q are the sums obtained in (2) from the g_n^m alone and the q_n^m alone, then $-\underline{\nabla}\,\psi_g$ is the magnetic field produced in the atmosphere by sources inside the earth (r<a), and $-\underline{\nabla}\,\psi_q$ is the magnetic field produced by sources outside the earth (r>a).

If $\underline{B}$ is known everywhere on the spherical surface r = a at time t, then $g_n^m(t)$ and $q_n^m(t)$ can be determined exactly. Without magnetic satellites, $\underline{B}$ is measured at about 200 magnetic observatories, giving 600 scalar data at time t. Then the series (2) is fitted to the data by one of three methods. The oldest is to truncate the series at a value of n small enough that considerably fewer than 600 Gauss coefficients must be determined. More recently, the whole series has been retained, and uniqueness of the coefficients assured by seeking that field which fits the data to within its estimated error and is smoothest (in one of several precisely defined senses) at the core-mantle boundary. The third method is like the second except that it imposes on the gauss coefficients not a smoothness criterion but a Bayesian a priori subjective probability distribution. The magnetic observatories are mostly in the northern hemisphere, mainly in Europe, so the advent of satellite data has improved the spatial distribution of the data as well as greatly increasing their number.

In the thin spherical shell where the satellite measurements are made, electric currents are not negligible, so use of the Gaussian model (1) and (2) for $\underline{B}$ is not entirely accurate. An alternative is to write $\underline{\mathbf{B}} = \underline{\nabla} \times (\underline{r}q) + \underline{\nabla} \times [\underline{\nabla} \times \underline{r}p]$, q and p being scalar fields. In the atmosphere, q = 0 and $\nabla^2 p = 0$, and this model reduces to that of Gauss. It can be used to find the Gauss coefficients in the atmosphere by combining satellite and observatory data.

B. Sources of the Field

Except during magnetic storms, more than 99% of the magnetic field $\underline{B}$ at the earth's surface is produced by electric currents in the conducting liquid core, driven by a self-sustaining dynamo process: fluid flowing across magnetic lines of force of $\underline{B}$ generates e.m.f's which drive the electric currents which maintain $\underline{B}$. The remainder of $\underline{B}$ is produced by electric currents induced in the mantle by time variations in $\underline{B}$; by permanent and induced magnetization in the crust; by tidal currents excited in the ionospheric dynamo (driven mostly by the thermal solar tide); and by the effects of the solar wind plasma in distorting the magnetopause and in producing the field-aligned currents in the magnetosphere which generate magnetic storms and aurorae.

C. Length Scales

At the earth's surface, all but about 10% of the magnetic field produced by the core is the field of a dipole at the earth's center inclined at 11 degrees to the axis of the earth's rotation (g^{-1}, g^{0}, g^{1} in (2)). The dipole field is about 60,000 nanoTesla (nT) at the poles and 30,000 nT at the equator. The remainder of the field has a complicated spatial pattern; figure 1 shows contours of the radial component of $\underline{B}$ after the dipole field has been subtracted.

A quantitative summary of the importance in $\underline{B}$ of the various horizontal wavelengths at the earth's surface can be obtained by considering separately the fields $\underline{B}_n$ for various spherical harmonic degress n. By definition, $B_n = -\nabla \psi_n$ where:

$$\psi_n = a\sum_{m=-n}^{n} g_n^m (a/r)^{n+1} Y_n^m(\theta, \phi) \qquad (3)$$

(Fields of external origin will be neglected). The function ψ_n has a horizontal wavelength on $r=a$ equal to $4\pi a/(2n+1)$, so the strength of the "signal" in $\underline{B}$ at this wavelength can be measured by $<|\underline{B}_n|^2>$, the average on the earth's surface of $|B_n|^2$. Figure 2 gives a graph of the observed $<|\underline{B}_n|^2>$ as a function of n for low n. For a random (spatial white noise) source in the core, $\ln <|B_n|^2>$ would be expected to decline nearly linearly and at least as rapidly as it does from $n = 1$ to $n = 12$ in Figure 2; and for a random source in the crust $\ln <|B_n|^2>$ would be approximately independent of n (actually $\approx \ln(2n+1)$) as it is for $n > 16$ in Figure 2. The usual interpretation of Figure 2 is that at the surface of the earth $\underline{B}_n$ comes mainly from the core for $1 \le n \le 12$, mainly from the crust for $16 < n$, and that both sources contribute in $13 \le n \le 15$.

Two important spatial variations are the field of order 200 nT near the equator, produced by an equatorial electric current in the ionosphere (the equatorial electrojet), and the fields of up to 2000 nT which can appear locally in the auroral zones during magnetic storms.

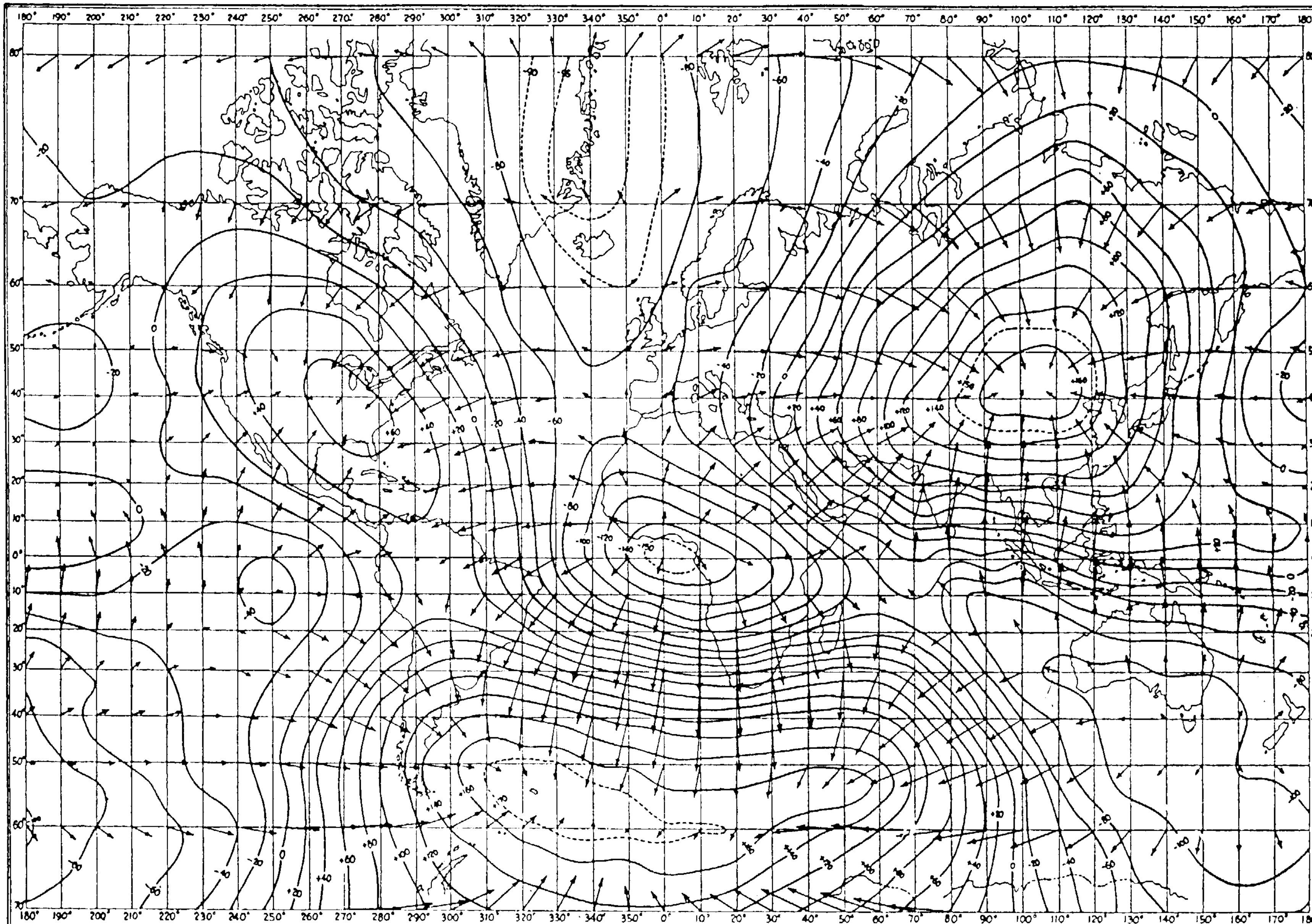
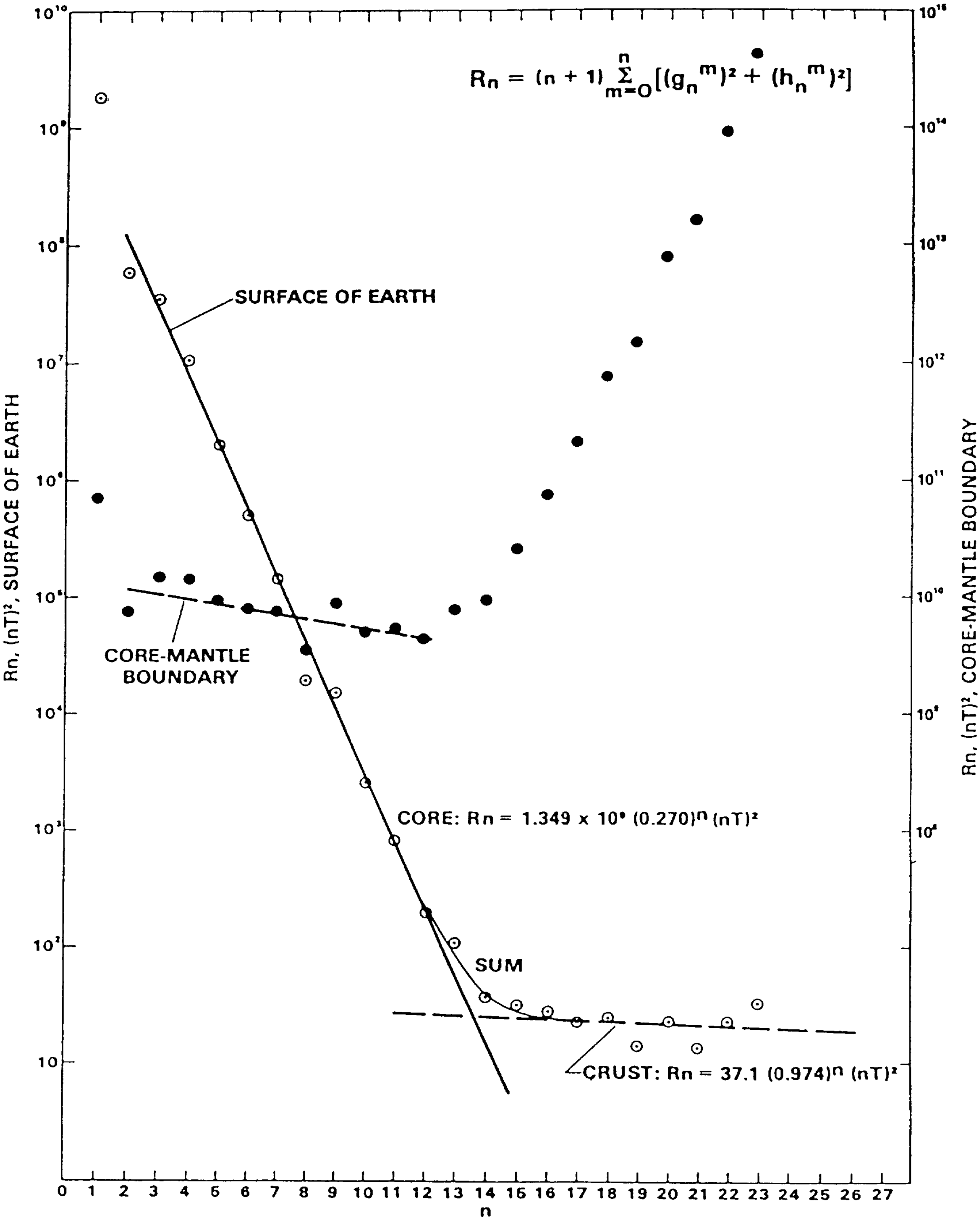


FIGURE 1. Non-dipole field for 1945. The contours give the vertical field at intervals of 0·02 gauss. The arrows give the horizontal component, an arrow 9·3 mm. long represents 0·1 gauss.

Figure 2. The observed $<|\underline{B}_n|^2>$ as a function of n for low n



D. Time Scales

Remanent magnetization of ancient rocks suggests that the geomagnetic field strength has not changed by more than a factor of three or four over the last 3 billion years, except that it drops by a factor of 10 or more during reversals. These reversals have occurred with increasing frequency during the Cenozoic, and the current average rate is about six per million years, but the pattern is random rather than periodic. There was a 30 million year period in the Cretaceous and another long period in the Permian which were free of reversals. During an individual reversal, the intensity of B may be anomalously low for about 10,000 years, and the reversal of its direction is completed in about 4000 years. Archeological evidence (magnetization of baked hearths and pots) indicates that the field intensity 2000 years ago at some sites was 50 percent higher than at present. In the last century the dipole has been decreasing at an average rate of about 0.05 percent per year. Between 1600 and 1800 the direction of the field at Paris and London changed by 34 degrees.

The secular variation $\partial_t \underline{B}$ is quite easily observable in magnetic observatory records, and is of the order of 20 nT/yr. About 25% of the energy in $\partial_t \underline{B}$ at the earth's surface can be removed by subtracting from $\partial_t \underline{B}$ the field obtained by rotating $\underline{B}$ rigidly and steadily westward at about 0.2 degree/year. This "westward drift" was observed qualitatively by Halley in the seventeenth century and led him to conjecture that the source of $\underline{B}$ rotated westward relative to the solid earth. It is now believed that the secular variation is produced by motion in the upper fluid core, and that this motion includes a general westward rotation and also a large random component.

Magnetic storms produce large changes in $\underline{B}$. A very large storm on April 16, 1938 changed $|\underline{B}|$ at Potsdam by 2000 nT in 15 minutes. Most storms produce changes of a few hundred nT, and die out over several days. The changes are so rapid that a satellite cannot distinguish them from spatial variations, and it would appear to be essential to use observatory records to supplement the satellite data during magnetic storms.

It remains a hotly debated question what are the minimum time periods detectable in the internal magnetic signal from the core. The conductivity of the lower mantle is poorly known, but if it is at least as high as the conductivity at a depth of 1000 km, then core periods shorter than one year will be screened out and will not be seen at the surface. The data at present do not exclude the interpretation that periods shorter than 10 years are screened out. What is needed to resolve these questions is an accurate measurement of the frequency content of the Gauss coefficients. Continuous satellite measurements might resolve the issue.

E. Correlations of $\underline{B}$ with Other Geophysical Phenomena

There appears to be a weak correlation between the westward drift rate of $\underline{B}$ and the variations in the length of the day, i.e., the component of angular momentum of the mantle about its axis of figure. Such a correlation would be expected from the electromagnetic coupling between the core and a conducting lower mantle, or from the effects of non-hydrostatic core pressure on CMB bumps.

A correlation of about 0.8 has been observed between the non axisymmetric parts of ψ and the geoid if both are truncated at $n = 4$. Whether this correlation is real and what it means is an open question.

It has been suggested recently that the Chandler wobble may be excited by fluctuations in the fluid motion in the core. If so, the Chandler wobble and magnetic time series should be correlated, since both have the same cause.

The external part of $\underline{B}$ is modulated by the sunspot cycle. Therefore, magnetic measurements may also be a useful supplementary measure of solar activity.

F. The Field of the Crust

The magnetic field at wavelengths less than about 3000 km is generally believed to be caused by sources within the crust and upper mantle. However, there are difficulties in accounting for all of the fields shorter than this wavelength using crustal sources because of the required strength of these sources, which exceeds the magnetic strength of appropriate materials studied in the laboratory. Possible resolutions of this difficulty are to call for significant upper mantle sources, or to suppose that some of the signal of wavelength shorter than 3000 km is produced in the core of the Earth.

Failed rifts (aulacogens), and their associated hot spot swells, are of particular interest as sources of magnetic anomalies since they are the sites of intense volcanism, and of mineral accumulations and petroleum deposits. They are commonly reactivated by alien tectonic forces leading to concentrations of earthquakes (e.g., Mississippi Embayment). The Rio Grande Rift shows as a prominent negative gravity anomaly from northwest New Mexico up along the Colorado-Utah border. This rift shows up less well on a short wavelength magnetic anomaly map than on an unfiltered map. Other rifts in the United States have been discovered using surface gravimetry and aerial magnetometry, and have been inferred on other continents from MAGSAT data. A combination of gravity field observations and magnetic field measurements by GRM and MFE would allow many more continental rifts to be discovered, modelled, and their geologic evolution inferred. Hot spots can also be inferred from magnetic field observations. The Yellowstone Region is an example of a hot spot swell which has a magnetic field signature.

Magnetic anomalies over the oceanic crust were the key which unlocked the plate tectonic pattern. Sea floor spreading anomalies are of too short a wavelength to be sensed by satellite, but there are longer wavelength anomalies whose sources are still unknown. For example, MAGSAT data show a significant magnetic anomaly associated with the bend in the Hawaiian-Emperor seamount chain, as well as other large anomalies in the western Pacific. But much of the oceanic crust appears to give only small signals at MAGSAT altitudes.

Although the resolution of MFE will not be any greater than MAGSAT, a second observation of total field at an epoch significantly different from that of MAGSAT will help in determining which fields come from the core, and which from crustal or upper mantle sources. Fields from the core are expected to change significantly over a 10-15 year period, especially those fields of wavelength close to the traditional core-crustal crossover wavelength of degree 13 or so. Degree 14 spherical harmonics are probably changing by about 6% per year if they come from the core. Higher degrees of harmonic will change by a relatively greater amount. Therefore a second observation will significantly aid in separating core and crustal fields, a major unsolved problem in magnetic anomaly interpretation.

6. Appendix II. Why Satellite Observations of the Geomagnetic Field?

A. Observatories

Magnetic observatories provide records of high accuracy and time resolution at about 200 places around the globe. They are very unevenly spaced, with blank "holes" of diameter 10,000 km in the South Pacific, 5000 km in the South Atlantic and 4000 km in the Arctic. A satellite can cover these holes, both to see what they contain, and to give better Gauss coefficients. The Gauss coefficients obtained from observatory data alone are contaminated by the uneven observatory distribution and by the rather large crustal magnetic anomalies on which many observatories lie, and which were ascertained only with the help of satellite observations. (The power density in $\underline{B}$ at high n is so large that the observatories' wide spacing permits serious spatial aliasing).

It should be emphasized that the observatory data are as important to satellite observations as vice versa. The variations of several hundred nT in less than an hour which occur in magnetic storms would be very difficult to interpret from satellite data alone. Such time scales permit serious temporal aliasing of the satellite data because a satellite will return within 5 degrees of the same spot on earth only once every 72 orbits, or about every 5 days. It will be very useful in dealing with such temporal aliasing in any program of near earth satellite observations of B if the existing array of ground observatories can be supplemented by magnetometers at the IDA and GEOSCOPE seismic stations.

B. Long-term Monitoring of $\underline{B}$

There are two justifications for long-term monitoring of $\underline{B}$ by satellite. First, although the crustal field is constant to the accuracy at which it is presently known, its accurate measurement is obscured by time-dependent external fields and, to a lesser extent, core fields. These fields must be studied over a time interval long enough that they can be understood and removed from the crustal field.

Second, the fluid motion in the core is probably chaotic on decade time scales, since the secular variation appears to be so. If this is the case, then many of the interesting properties of the core dynamo will be statistical, and cannot be measured except by examining a long time series. It is clear from the secular variation that the temporal spectrum of $\underline{B}$ has appreciable power at all periods between several months and several millenia, so there is really no upper limit to the lengths of interest for an accurate time series of $\underline{B}$. Moreover, there is convincing evidence that some of the interesting behavior of the dynamo consists of isolated events, of short duration compared with the time between them. One example is field reversal. Another may have been a worldwide jump in $\partial^2_t \underline{B}$ around 1969, which may have occurred in only one or two years. Similar events may have occurred in 1913 and 1978.

There is considerable optimism in the geomagnetic community that a knowledge of the detailed time and space dependence of $\underline{B}$ at the earth's surface can be used to study the properties of the fluid motion in the core. The actual values of the fluid velocity in the upper core may be recoverable from the magnetic data, and a sufficiently long time series may give insight into the forces which drive the dynamo. For example, is the upper core stably stratified? Is flux conservation a good approximation over the decade time scale, even though flux diffusion (probably by turbulent advection) is essential to the operation of any dynamo? Does Coriolis convection twist the field in the deep core, occasionally releasing energy and angular momentum in a catastrophic kink instability? The kink instability has been suggested as one explanation for the jerk of 1969; and that instability may be one way of achieving dynamo action to generate $\underline{B}$. These events, if they exist, are isolated, random, and at present unpredictable. To study them requires continual monitoring of the field. They are to geomagnetism what earthquakes are to seismology, and the argument for permanent monitoring is the same in both disciplines. Such monitoring is justified at least until we have seen enough events to understand the phenomena.

No one has yet succeeded in building a fluid dynamo in the laboratory, so our best chance to study this fundamental physical phenomenon experimentally is to examine the closest available dynamo, the one in the core. The sun is another source of dynamo data, but the balance of forces there seems to be quite different from that in the earth, so the earth and sun probe different parts of the parameter space of fluid dynamos. As an example, consider the magnetic Reynolds number av/ρ, the ratio of the production rate of $\underline{B}$ by fluid motion at velocity v to the ohmic decay rate of $\underline{B}$ if its length scale is a and the magnetic diffusivity is ρ ($\rho = (\mu_0 \sigma)^{-1}$ where σ is the electrical conductivity and in MKS units $\mu_0 = 4\pi \times 10^{-7}$). It is known that dynamo action is impossible if $av/\rho < 10$. In the earth's core, av/ρ is probably about 300. In the sun, it is perhaps 10^6.

If long term monitoring of $\underline{B}$ is to capture isolated core events, the monitoring must resolve the spatial details of these events contained in the spherical harmonics of high degree. On the basis of figure 2, the crustal contribution to the field $\underline{B}_n$ of harmonic degree n can be neglected at least up to degree $n = 12$. Core details on this scale are in principle available at the earth's surface, but to study them requires measuring 168 Gauss coefficients in the face of the difficulty that $\underline{B}_n$ at the earth's surface is attenuated relative to the core-mantle boundary by the factor $(0.545)^{n+2}$. If $|\underline{B}_{12}|/|\underline{B}_1|$ is 1 at the core- mantle boundary, it is 4×10^{-3} at the earth's surface, so $\underline{B}_{12}$ there is of the order of 200 nT. Energy equipartition (which has at present no physical justification, and is an argument from ignorance) suggests that each of the 25 internal Gauss coefficients at $n = 12$ will contribute about 15 nT to this signal at a satellite altitude of 600 km, or 40 nT at the earth's surface. To study these phenomena requires high accuracy and good spatial resolution.

The 200 existing magnetic observatories do provide 600 magnetic components of high accuracy, but their spatial resolution is poor for two reasons. They are very inhomogeneously distributed on the earth's surface, and there is enough crustal energy at high n to produce serious spatial aliassing. Another way to put the problem is that each observatory sits on a local crustal magnetic anomaly. The POGO and MAGSAT data have made it possible to study these local anomalies, and they appear to be as large as 500 nT. Supplementing the observatory data with magnetometers at the IRIS and GEOSCOPE seismic stations will do much to homogenize the station distribution, but it will not solve the problems of spatial aliassing and station correction. Only satellite data can provide the necessary density of spatial coverage at a realistic cost.

C. An International Program

There is much to be learned about the core by satellite monitoring of $\underline{B}$ with high accuracy and high spatial resolution indefinitely into the future. It is proposed that the NASA EOS program will furnish a long-term platform from which continuous measurements can be made. To fill the intervening gap, it is hoped to persuade a number of countries to launch successive satellites with 3-5 year lifetimes. Overlap is to be encouraged, both for cross-correlation of instruments and because the electric current distribution at satellite altitudes is of interest in itself, and can be studied in some detail with multi- satellite observations. The MFE was proposed as the first of such satellites, to be followed by the French satellite Magnolia.