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SECTION IX.

REPORT OF THE PANEL ON GEOPOTENTIAL FIELDS: MAGNETIC FIELD

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SECTION IX.

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SUMMARY

The objective of the NASA Geodynamics program for magnetic field measurements is to study the physical state, processes and evolution of the Earth and its environment via interpretation of measurements of the near Earth magnetic field in conjunction with other geophysical data. The fields measured derive from sources in the core, the lithosphere, the ionosphere and magnetosphere of the Earth.

Density differences in the Earth's outer core drive a motion there which sustains the main geomagnetic field by dynamo action. This dynamo, together with the core-mantle boundary (CMB) and the electrically conducting lower mantle constitute a physical system of great intrinsic interest and considerable utility in monitoring and understanding other Earth subsystems. For example, it has been suggested that the reversal frequency of the main geomagnetic field 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 are responsible for plate tectonics.

Improving our understanding of how the geodynamo works is a major issue in modern geophysics. This system is clearly non-linear, 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. For example, seismology and gravity give evidence for undulations in the CMB and for temperature variations in the lower mantle which can affect (or be affected by) core convection and hence the dynamo. Measurements of the magnetic field and its temporal variation give the most direct access to the core dynamo and the electrical conductivity of the lower mantle. Only space measurements can provide the global and quasi-instantaneous description of the Earth's field required to fully describe its time and length scales. These measurements should be supplemented by ground based observatory data as well as analysis of the paleomagnetic record.

Permanent and induced magnetization in the Earth's lithosphere reflect the effects of present and past magnetic fields, and chemical composition as well as the thermal and chemical history of rocks in the crust and upper mantle. These have been influenced by tectonic history and retain part of the record of that history. Data are required from surface, as well as by satellite surveys, but only low altitude satellite surveys can be made in a reasonable time and in a sufficiently self-consistent manner to measure the extremely long wavelength magnetization undulations which apparently originate deep in the crust and, possibly, in the upper mantle. Ambiguities in interpretation require that interpretation include geophysical data of all types, with particular emphasis on seismic, tectonics, gravity and heat flow data.

The ionosphere supports a current system almost steady relative to the sun. 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. Satellites are above the ionosphere, so in the gaussian representation of the field 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.

Time and Length Scales

Time variations of the Earth's magnetic field on time scales of seconds to decades allow us to determine mantle conductivity from the surface down the the CMB and to study the kinematics and dynamics of the geodynamo. Thus it is essential to have quasi-continuous measurements of the geomagnetic field **over several decades**, and aliasing prevents us from obtaining such data by taking a snapshot of the field with a short satellite mission every ten years or so.

The spatial scales of the internal magnetic field vary from the very long wavelength features caused by electric currents in the core, to wavelengths of only a few m produced by highly magnetized rocks at the surface. In general terms, wavelengths between 40000 km down to 4000 km represent core sources, while signals shorter than 3000 km represent lithospheric sources. The signals in the middle are from both sources, which cannot be untangled in this domain. In addition, external currents can produce signals at all wavelengths. In order to extrapolate these signals down to the CMB, good spatial coverage is essential, and aliasing from the lithospheric signal must be removed. Similarly, in order to study the lithospheric field, global spatial coverage must be obtained. Magnetic observatories provide neither good spatial coverage nor protection from spatial aliasing. Aeromagnetic and shipborne surveys cannot provide sufficient data in a reasonable time frame. It is difficult to obtain both lithospheric and core fields in one satellite mission. **High resolution in the lithospheric field requires a low satellite altitude and hence a short life time, whereas useful measurements of the core field must extend over decades.**

Progress and Requirements

Satellite vector data are a requirement if the present geomagnetic field is to be studied properly. Additional information could be provided at earlier geologic times if a concerted effort is made to collect magnetic direction information recorded in lakes and in archaeological samples such as kilns and baked hearths. But the main thrust is to improve our collection of data in the future by a long term series of observations of the field from satellite. The Earth Observing System (Eos) is due to launch a mission to measure the magnetic field no earlier than 1998. Effort should be made to fill the gap between this and Magsat data (collected in 1979/1980) by collaboration with CNES in the MFE/Magnolia project, which would monitor the field for a minimum of five years, and with ESA in the Aristoteles project, which would measure the field at a considerably shorter wavelength than Magsat, provided that a vector magnetometer is deployed. Limited coverage of some low latitude areas could be obtained from a tether instrument deployed from the Space Shuttle. A magnetic field gradiometer could be developed for future deployment.

In order to make the most effective use of the satellite information, additional magnetic observatories, especially some in ocean basins, are needed. Accompanying this effort there must be more work done on modeling the external field both to study its sources and to help in isolating lithospheric and core field information. In addition, more rock magnetic measurements need to be done on appropriate samples to help constrain the crustal models produced from the magnetic field and other information.

Panel recommendations

The recommendations of this panel are as follows:

(1) Initiate multi-decade long continuous scalar and vector measurements of the Earth's magnetic field by launching a 5 year satellite mission to measure the field to about 1 nT accuracy. This mission could be MFE/Magnolia.

(2) Improve our resolution of the lithospheric component of the field by developing a low altitude satellite mission. This could be accomplished by including a magnetometer with about 1 nT accuracy on the ESA Aristoteles mission. A vector instrument would give considerably better information than a scalar instrument. If Aristoteles cannot measure the low altitude magnetic field at sufficient accuracy, we recommend that a magnetometer and boom be placed on SGGM to give the lithospheric vector field at about 1 nT accuracy.

(3) If in fact Aristoteles has a 1 nT vector magnetometer and an extended [3 yr] high altitude phase, then it would also contribute to our understanding of the secular changes of the Earth's magnetic field. This is a third priority recommendation, contingent on the above conditions.

(4) Support theoretical studies and continuing analysis of currently available data to better understand the source physics and to improve the modeling capabilities for the different source regions.

(5) Develop a gradiometer device for magnetic field measurements in order to improve the short wavelength information and isolate the fields from **in situ** currents.

(6) Improve the recording of the time varying field by installing new observatories in critical areas (especially the ocean floor) and by upgrading some observatories.

(7) Continue rock magnetic and electrical conductivity studies of deep crustal rocks and analogs of mantle rocks to help constrain magnetic and electrical models for the lithospheric field.

8) Improve our understanding of the long term variations of the field by continuing paleomagnetic studies of archaeological samples and rapidly deposited sediments from lakes and the ocean floor.

9) Look ahead vigorously with plans for future missions, including the GOS experiment on Eos and subsequent missions timed so as to obtain near continuous coverage for several decades. Such plans should also include a multi-satellite mission configuration for thorough study of the local time morphology of the external fields.

DESCRIPTION OF THE GEOMAGNETIC FIELD

Introduction: The evolution and dynamics of our planet affect all of us. Yet the processes and forces involved are in-the-main hidden from view in the deep interior of the Earth. Few tools are available to probe that interior: seismology, gravity, magnetism, Earth rotation, heat flow and geodesy. Three sources contribute to the magnetic field near the Earth: currents in the core, magnetization in the upper lithosphere, and currents outside the Earth and their induced components inside the Earth. The core, or main, field is by far the largest; because it must travel through the mantle, this field yields information both on the region of its generation and on the electrical conductivity of the mantle. In extracting this information the temporal variation of the field is at least as important as its description at a particular epoch. The time variations of the main field occur over periods from months to decades and beyond and, as a result, require continuous and long-term measurements for an adequate characterization.

Electrical currents outside the Earth form a part of the natural plasma laboratory called the magnetosphere and ionosphere. Of particular interest to solid Earth studies are those currents flowing in the ionosphere and along the magnetic field lines connecting the ionosphere and the magnetosphere. The magnetosphere transfers large amounts of energy to the ionosphere and, via ion-neutral coupling, the upper atmosphere through mega-Ampere field aligned electrical currents at auroral latitudes. Meridional currents within the ionosphere have been discovered and associated with the Equatorial Electrojet. These time-changing external fields cause induced fields in the Earth's crust and mantle the study of which yields information regarding the conductivity of these regions.

Lithospheric magnetic fields are called anomaly fields because in practice what is studied is the residual field when estimates of the core and external fields have been subtracted from the measured field. Maps of anomaly fields have been derived from aeromagnetic and shipborne data for many years and used in the formulation of geological/geophysical models of the crust. Investigations with aeromagnetic and shipborne magnetic data have mainly concentrated on the relatively localized anomalies associated with small scale geologic features and localized mineralization. However, in the past few years there has been an increased interest in studies of the broad scale anomalies that appear in regional compilations of aeromagnetic and shipborne data. Satellite anomaly maps are of recent origin and describe only the very broadest scale anomalies. Originally it was thought impossible to detect fields of lithospheric origin in satellite data. However data from the POGO and Magsat satellites showed that low altitude data contain separable fields due to lateral variations of magnetization in the upper lithosphere, thus opening the door to a new class of investigations.

Ground based magnetic field measurements are seriously deficient for main field studies, both in their spatial and temporal distribution. It is fair to say that main field geomagnetism has reached the point where accurate global vector data extending over a significant time span, ideally decades or more, are required for significant advances. It is evident that such data can only be acquired from space. Previous measurements of the geomagnetic field from space onboard Kosmos-49, the POGO satellites, and Magsat, have made significant contributions. However the Kosmos and POGO measurements were of the field magnitude only and the Kosmos and Magsat lifetimes were very short. Accurate global vector measurements over an extended time span are lacking.

Similar deficiencies plague the study of lithospheric fields. Usually, localized surveys conducted by ship and aircraft cover too limited an area to map anomalies of the tens and hundreds of kilometer size and do not provide the three components of the field. Piecing together smaller surveys results in distortion and aliasing, particularly if the surveys were acquired at different epochs and the main field has changed significantly. Satellite data are global in nature, eliminating problems of political boundaries and logistics. Data from the POGO and Magsat satellites have demonstrated that lithospheric fields can be studied from satellite altitudes. However, these analyses have also shown that lower altitude data are needed to bridge the gap between the shorter wavelength features measured by aircraft and ships and the very long wavelengths measured in currently available satellite data.

Gauss's Representation of the Field: In regions where no current flows the magnetic field, **B**, can be represented as the gradient of a scalar potential, Ψ.

$$\mathbf{B} = -\nabla \Psi \tag{1}$$

The usual representation for Ψ is:

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

Here a is the radius of the Earth, r is radial distance from the center of the Earth, θ is colatitude, ϕ is east longitude, $Y_n^m(\theta,\phi)$ is the surface spherical harmonic of degree n and longitudinal order m, and g_n^m and 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 **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 $-\nabla\Psi_g$ is the magnetic field produced in the atmosphere by sources inside the Earth (r<a), and $-\nabla\Psi_q$ is the magnetic field in the atmosphere produced by sources outside the Earth (r>a). If **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, **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 first two methods produce at least one model which fits the data, but do not address the question of how many models will work. This uniqueness problem can be solved by the third method, but the availability of the prior information required for that method is debated. Recently, a fourth method, based on Neyman's confidence sets, has been worked out in principle. It is objective, solves the uniqueness problem, and provides rigorous error bounds for the results of the first two methods. Its software implementation has only just begun. 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; but satellite coverage in the time domain remains inadequate - Magsat having been launched a full decade ago and POGO data ceasing eight years prior to Magsat.

In the relatively thin spherical shell where near Earth satellite measurements are made, electric currents are not negligible, so use of the Gaussian model (1) and (2) for **B** is not entirely accurate. To include average effects from these currents, an alternative is to write $\mathbf{B} = \nabla \times (\mathbf{r}q) + \nabla \times [\nabla \times \mathbf{r}p]$, 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. This, and more detailed, modeling of field-aligned current systems is still at a relatively early stage. Determinations of local current density have been limited by both the quality and quantity of the available field observations, a situation which can be remedied by additional satellite data. Development, testing, and application of quantitative modeling methods for high latitude current systems will further both space plasma physics investigations and our ability to separate magnetic fields from sources internal and external to the Earth.

__Sources of the Field__: Except during magnetic storms, more than 99% of the magnetic field **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 **B** generates e.m.f's which drive the electric currents which maintain **B**. The remainder of **B** is produced by electric currents induced in the mantle and the oceans by time variations in **B**; by permanent and induced magnetization in the crust and, possibly, in the upper part of the mantle where the temperature is below the Curie temperature of magnetic minerals; 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.

__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. The strength of 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 **B** after the dipole field has been subtracted.

A quantitative summary of the importance of the various horizontal wavelengths at the Earth's surface can be obtained by considering

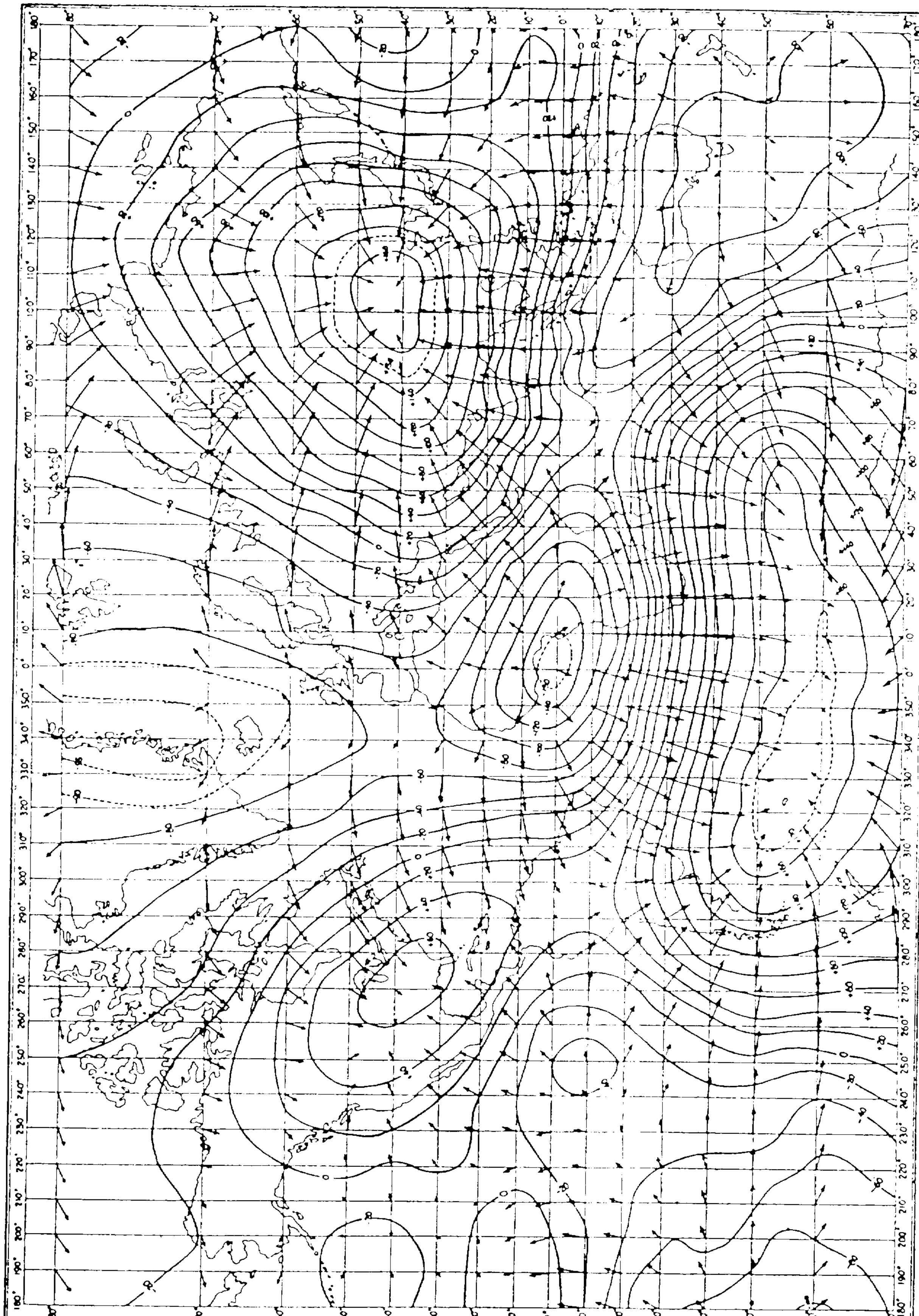


FIGURE 1. Non-dipole field for 1945. The contours give the vertical field at intervals of 0·02 gauss. The arrows give the horizontal component, an arrow 9·3 mm. long represents 0·1 gauss.

separately the fields $\mathbf{B}_n$ for various spherical harmonic degrees n. By definition, $\mathbf{B}_n = -\nabla\Psi_n$ where:

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

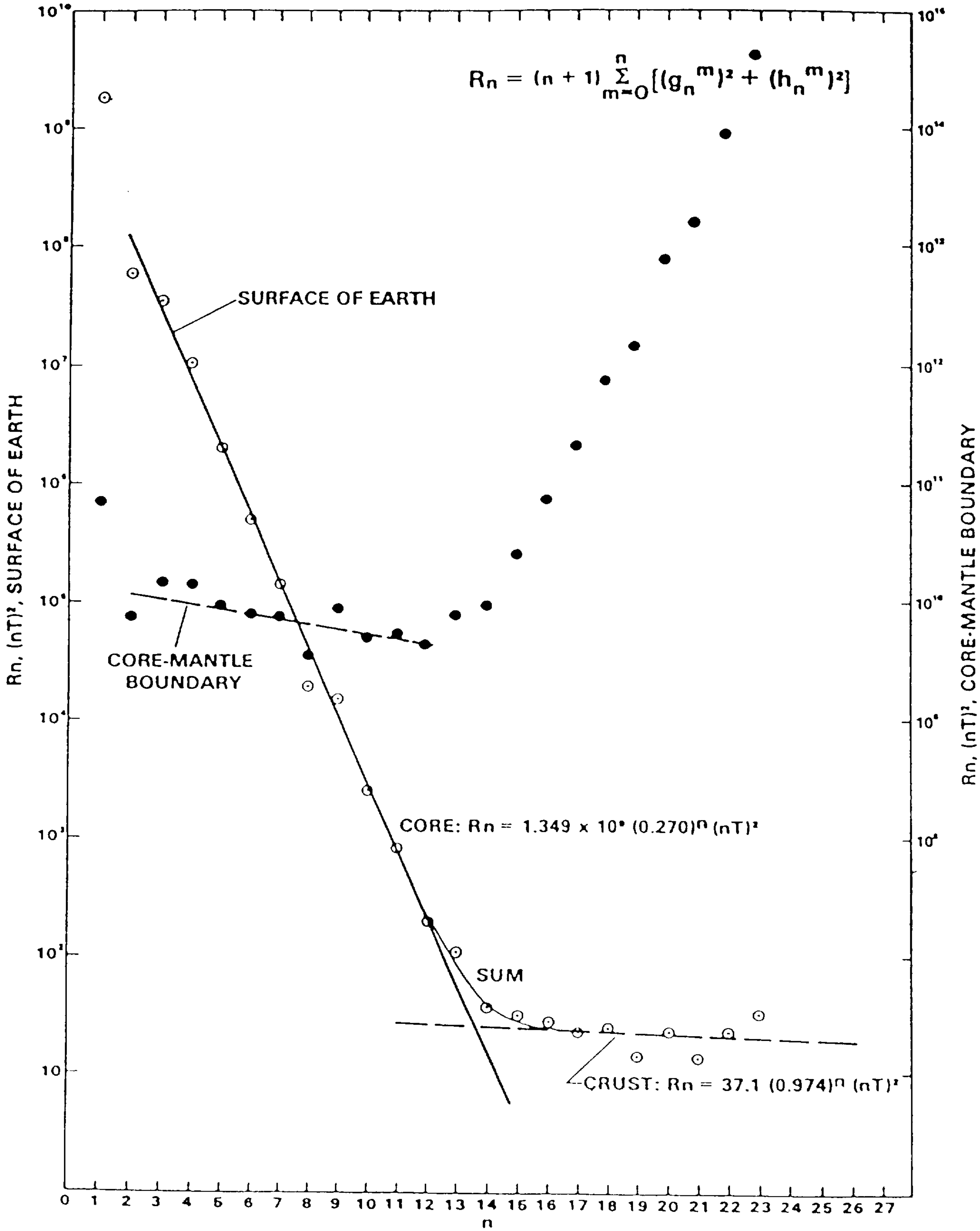
(Fields of external origin are neglected in this treatment). The function Ψ_n has a horizontal wavelength on r=a about equal to $4\pi a/(2n+1)$, so the strength of the "signal" in **B** at this wavelength can be measured by $<|\mathbf{B}_n|^2>$, the average over the Earth's surface of $|\mathbf{B}_n|^2$.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

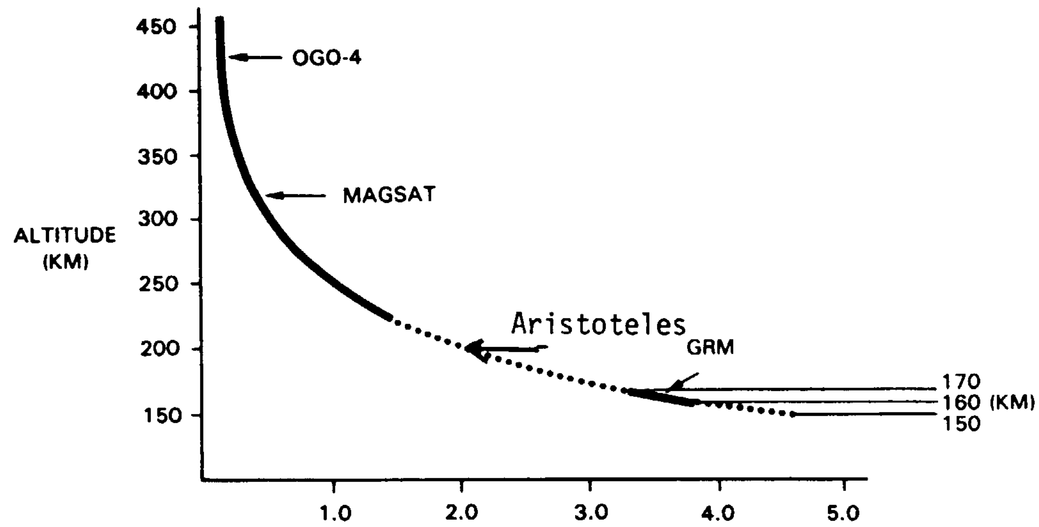
With existing satellite data, the crustal portion of the spectrum can be extended to an n of about 50 which corresponds to a wavelength of about 800 km. It is more convenient for anomaly fields to talk about characteristic size in km, where by characteristic size we mean the distance between the half-amplitude contours of an anomaly. Aeromagnetic and shipborne surveys map anomalies of characteristic size 10's to 100's of km. The upper limit depends upon the area and spacing of the survey, on the altitude of the survey above the sources and on the relative accuracy to which the spacing between data points is known. In principle, all of the crustal field could be mapped by aircraft; in practice the survey size is limited by political boundaries and by logistic considerations. Satellite data can, in principle, be recovered from altitudes as low as 150 km. Characteristic sizes of 150 km and greater should be recoverable from data at that altitude. Figure 3 shows how the relative amplitude and the resolution of magnetic anomalies vary with altitude.

Spatial variations in the anomaly field reflect variations in the near-surface magnetization. Magnetization variations are not only due to variations in the amount of titanomagnetite and titanomaghemite, and the relative amount of titanium in these minerals. Other contributors are variations in topography and in the depth to the Curie isotherm. Magnetization content variations reflect the past history of the rock: the chemistry of the rock source and the processes by which it was formed, the thermal and mechanical history of the rock formation, and alteration due to diagenesis and metamorphism.

Important spatial variations of the magnetic field are produced by external currents. Among these are the field of order 200 nT near the equator, produced by an equatorial electric current in the ionosphere (the equatorial electrojet), the Sq fields of order 10 - 40 nT from the ionospheric dynamo, and the fields of up to 2000 nT which appear in and poleward of the auroral zones. The magnitudes given are at the surface of the Earth. The equivalent spatial resolution of previous magnetic field investigations has been limited to approximately 0.5 km. These experiments have shown that high local field aligned current densities can occur over spatial scales of 1-10 km. Optical observations of auroral forms have shown that fine structure exists down to length scales as small as tens of meters.

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





Relative Magnetic Anomaly Amplitude as Observed at Spacecraft

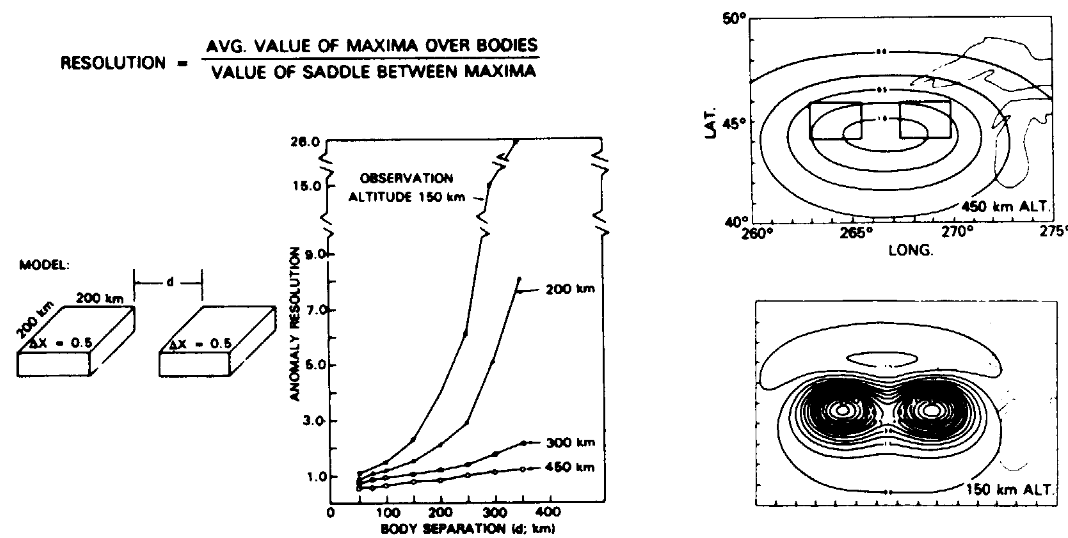


Figure 3 Resolution of Magnetic Anomaly as a Function of Altitude

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 5 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. For example, the typical period between the last few reversals has been of order 170,000 years, but it has been over 700,000 years since the last complete reversal. 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 or less, and the reversal of its direction is typically completed in about 4000 years. Archaeological 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 $\partial_t \mathbf{B}$ is quite easily observable in magnetic observatory records, and is of the order of 60 nT/yr. It varies greatly with position and time. About 25% of the energy in $\partial_t \mathbf{B}$ at the Earth's surface can be removed by subtracting from $\partial_t \mathbf{B}$ the field obtained by rotating the **B** pattern rigidly and steadily westward at about 0.2 degree/year. It is now believed that the secular variation is produced by fluid motion in the upper core, and that this motion may include a general westward rotation but also a large spatially variable component.

Magnetic storms produce large changes in **B**. A very large storm on April 16, 1938 changed $|\mathbf{B}|$ 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 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 field fluctuations with 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 strongly attenuated. 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.

A distinct global change of the rate of secular variation was detected at about 1969-1970. This change was called the geomagnetic impulse or jerk. The change seems to have taken place in less than 1 year. Unfortunately only scalar satellite data are available at this time (and these ceased in 1970!) and these, even together with observatory data, are not able to resolve either the spatial or temporal characteristics of the jerk with sufficient accuracy for definitive studies of either its nature or the conductivity of the mantle through which it passed.

High latitude ionospheric and field aligned current systems exhibit variability on time scales ranging from 11 years, the solar cycle, to periods of <1 sec for electromagnetic plasma waves and instabilities.

To the extent that lithospheric sources are remanent rather than induced, they should not change measurably with time. Induced anomalies have magnetization strength proportional to the main field and so will change with the main field. However those changes are less than 1% of the anomaly magnitude per year which is not detectable with present measurement accuracies. This provides a test of the anomaly source. If it is found that what is thought to be the anomaly field changes significantly with time then that field is most likely not from a lithospheric source but from a contaminating core or external source.

Separation of core, crustal and external fields is one of the major problems in geomagnetism. The spatial spectra of the three sources overlap; e.g. in Figure 2 the lithospheric field is present for degree lower than 13, but it is masked by the core field. Similarly, the core field is present for degree greater than 13, but it is masked by the lithospheric field. It would seem that the fields in these "masked" regions will never be observable. This implies limitations on analyses of both the core and lithospheric fields which must be recognized. The external field covers the entire spectrum, but its temporal variations seem to be distinctly different from those of the core and lithospheric fields for the portion of the spectrum where the core field is observable. Analyses of satellite data have been successful in isolating and modeling the long wavelength external field. There remain some effects of low amplitude which still require modeling, e.g. from persistent field aligned currents and from seasonal variations in the magnetospheric configuration. To date these are apparently only observable in the Magsat data and they have yet to be adequately modeled.

Separation of the lithospheric and external fields has been only partially successful. This is true for both surface (including airborne and shipborne) and satellite data. Periods when data clearly show magnetic disturbance can be used profitably to study the sources of disturbance and the regions of the Earth in which induced fields occur. However, there are apparently no times when the external fields are totally absent from the Earth. Even at the most quiet magnetic times there are some localities where external fields are present. This is particularly true for the auroral and polar regions and also for daylight hours at all locations. In principle, simultaneous global measurements both at the Earth's surface and above the ionosphere would allow a good separation. Such measurements are not available. In practice, separation has been accomplished by decomposition of satellite data into parts which vary with geographic position and/or universal time from those which vary with dip or dipole latitude and/or local time. In some cases surface data beneath the satellite track have also been used. Some of these methods have generated considerable debate regarding the meaning of the results. Further theoretical and analytical analyses are needed.

Correlations of B with Other Geophysical Phenomena: There appears to be a weak correlation between the westward drift rate of **B** and the variations in the length of the day, which is proportional to the reciprocal of the angular momentum of the mantle. A stronger correlation , 0.71, exists between the rate of change of the dipole and the variations in the length of the day (see Figure 4). Such correlations 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 the topography of the CMB.

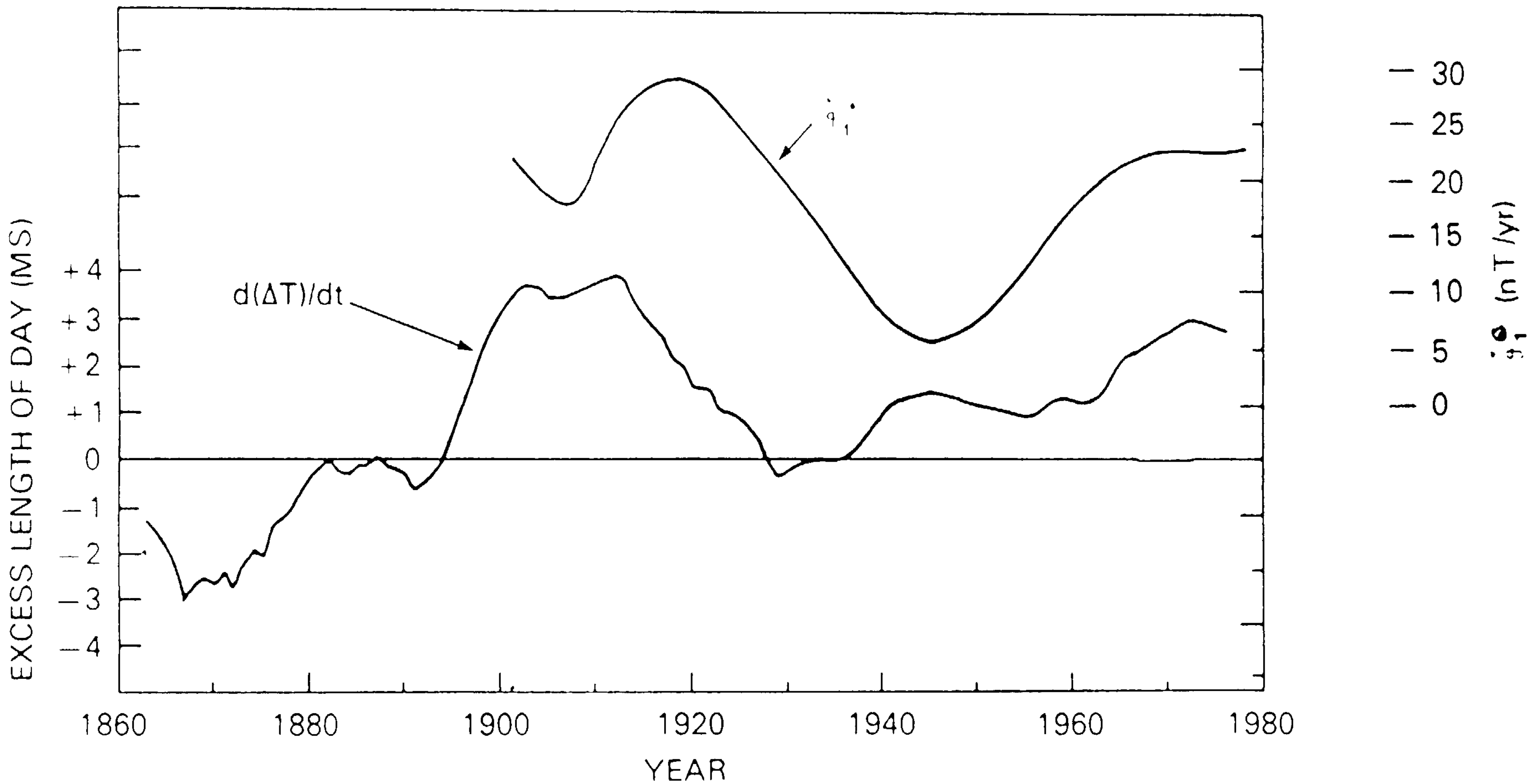

EXCESS LENGTH OF DAY (MS)
+4
+3
+2
+1
0
−1
−2
−3
−4
d(ΔT)/dt
$\dot{g}_1^0$
30
25
20
15
10
5
0
$\dot{g}_1^0$ (nT/yr)
1860
1880
1900
1920
1940
1960
1980
YEAR

Figure 4.

A correlation of about 0.8 has been observed between the non axisymmetric parts of Φ and the geoid if both are truncated at $n = 4$. Whether this correlation is important 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 in some way, since presumably both have the same cause.

Of all physical quantities that can be measured in the plasma environment of the Earth, the magnetic field plays the defining and unifying role, physically tying points of different regions of geospace together, guiding charged particles, plasma waves and electric currents, trapping thermal plasma and energetic particles, and transmitting stresses from one region to another. The magnetic field provides the fundamental link between the solar wind and the magnetosphere (regulating the efficiency of the energy transfer and the solar wind dynamo), and between the magnetosphere and the ionosphere (with the field-aligned currents playing the role of communicator between the corresponding MHD and resistive plasma regions). Also, it is the magnetic field which "projects" the effects of magnetospheric processes onto the high latitude ionosphere, converting the latter into a giant viewing screen of geospace behavior. Measurement of the magnetic field thus provides basic information on the magnetosphere without which any attempt at quantitatively piecing together and understanding the entire solar-terrestrial system would be severely impaired.

As will be discussed later, the geologic interpretation of magnetic anomaly data is non-unique. Observed anomaly values can be reproduced by an infinite number of distributions of magnetization within the upper lithosphere. A similar situation is true for interpretation of gravity anomalies. Reduction of this ambiguity is achieved by constraining models through the use of other geologic and geophysical data. One method of reducing this ambiguity is by the joint analysis of gravity and magnetic anomalies. This is complicated by the fact that the gravity data do not have two distinct source regions like the magnetic data (core and crust, not mantle); gravity sources are contained continuously throughout the interior of the Earth. Models can also be constrained by noting the boundaries of known geologic and tectonic regions, by noting seismic boundaries, and by taking into account measurements of topography, heat flow and other quantities.

<u>Representation of anomaly fields.</u> In practice, equation (2), truncated at suitable n, is used to represent the field from the core. Other methods are generally used to represent the lithospheric field. In particular, a formalism has been developed for synthesizing Bouguer gravity measurements or magnetic anomaly data on an irregular three dimensional grid. The synthesis consists of a mathematical representation of the data in terms of discrete point masses for gravity or of dipoles for magnetics at the Earth's surface or at some appropriate fixed depth below that surface. In this method the satellite magnetic anomaly data are represented by an array of dipoles at the Earth's surface. The dipoles are assumed to be aligned along the direction of the Earth's main field, as determined by a spherical harmonic model, and their magnitudes are determined so as to best reproduce the anomaly data in a least-squares sense. The resulting synthesized fields are correct whether the underlying

magnetization distribution is induced, remanent or some combination. But the resulting magnetization distribution has meaning only if the magnetization is induced, an assumption that is often made but is still subject to debate. This kind of model is called an equivalent source model. Figure 5 shows a satellite anomaly map of the U.S. reduced to 400 km altitude by this method. The resulting dipole moments can be converted to depth-integrated magnetization, provided the appropriate depth is known. Such magnetization values (defined as the magnetic moment per unit volume) are determined under the assumption that the magnetic crust is of constant thickness with constant magnetic moment throughout the layer. Because the satellite altitude is large compared to typical layer thicknesses, the anomalies computed from the equivalent source solution depend directly upon the product of magnetization and layer thickness, i.e. if the magnetization is doubled and the thickness halved the computed anomaly will be unchanged.

Nonuniqueness of anomaly data interpretation. There are at least two fundamental sources of non-uniqueness to be considered in analysis of magnetic anomaly data. The first is that the presence of the large field from the core effectively masks all surface anomalies below spherical harmonic degree 14 and adds unknown amounts to the degree 14 and 15 coefficients. Some of the consequences for interpretation of not knowing the low degree field have been investigated, but more effort is needed. A related problem is that in some instances it was found that an along track trend removal applied to the data for the average map has distorted the zero level and seriously affected interpretation of the anomaly pattern.

The second source of non-uniqueness is inherent in the nature of the inverse problem. When any equivalent source solution is produced, it is possible to add or subtract certain magnetization distributions which have no effect on the external field. These magnetization distributions are known as annihilators, and they vary according to the geometry of the source region. This means that the original solution is not unique.

For the case of a spherical shell, the annihilator is any magnetization whose direction and intensity are directly related to any field whose source is within the shell. Various schemes have been suggested to deal with this problem of non-uniqueness. One approach has been to compute magnetizations and then add the magnetization due to some annihilator so as to obtain solutions which are considered more realistic than the equivalent source solution alone. The justification for this procedure is that magnetizations computed from the equivalent source technique are both positive and negative but that only positive magnetizations should result from induction. The suggested procedure is to add the magnetization due to some annihilator until all resulting magnetizations are greater than or equal to zero. This is a reasonable procedure provided it is realized that this solution is still not unique.

The existence of one or more annihilators means that a basic ambiguity exists in magnetization solutions which can only be removed by using other geophysical experience or data to restrict the solution.

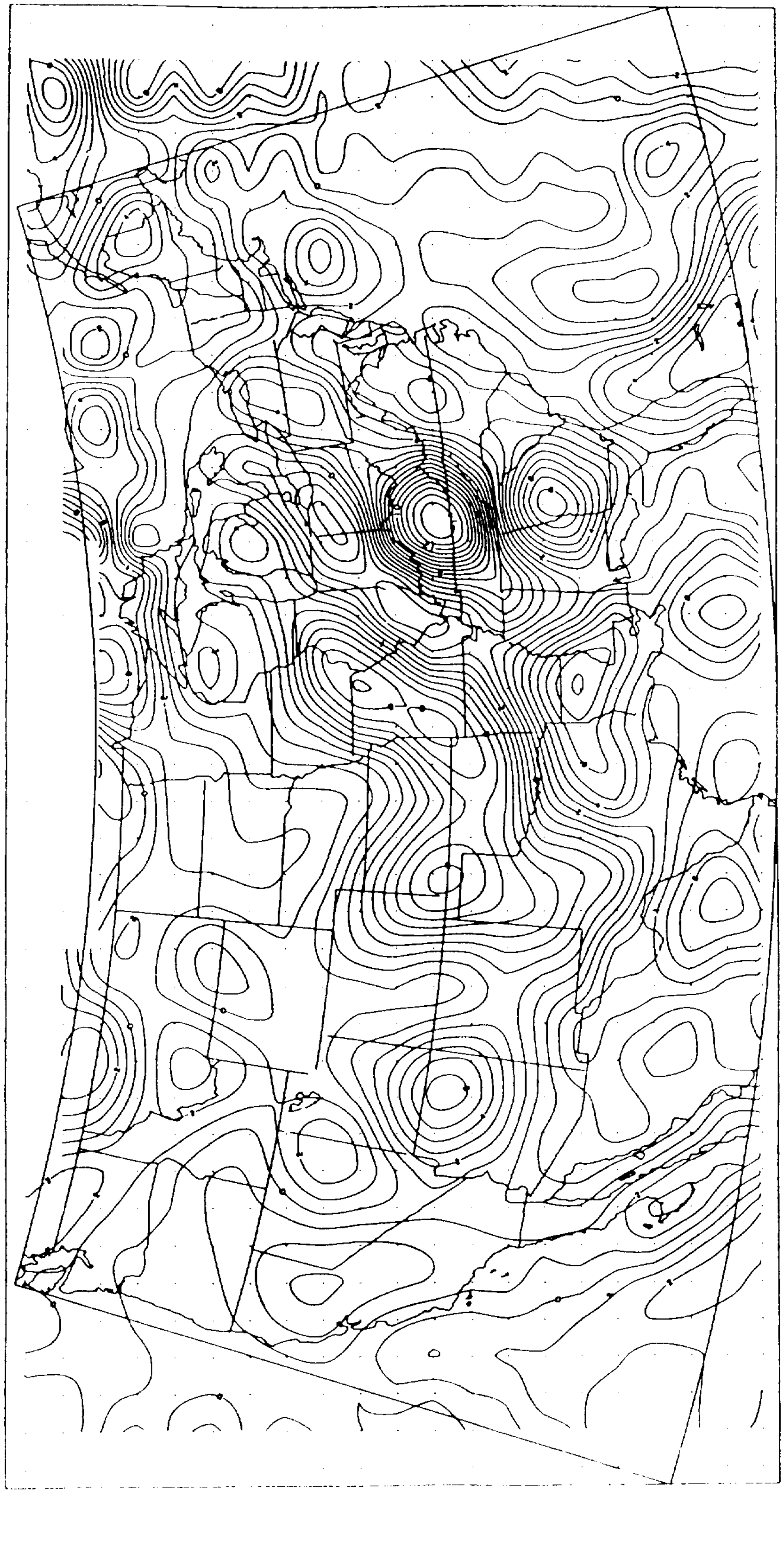


Figure 5

Physical processes which operate on the Earth may be duplicated or may have related counterparts on other planetary bodies. For example, it appears that not only the gas giant planets but also Mercury, and possibly some of the Jovian satellites, have functioning dynamos. While the exploration of our solar system functionally lies in a separate division within NASA, it is not only proper but essential that this separation not hinder the integrated studies that are so necessary for understanding the underlying physical phenomena.

The study of planetary geopotential fields has already begun with flyby missions of several planets and satellites and orbiting missions at Venus and Mars. Magnetic field measurements conducted by Pioneer Venus Orbiter found no measurable planetary magnetic field; it is concluded that Venus has no internal dynamo. The interpretation of measurements at Mars from the Soviet Mars 2 and 3 missions have been disputed. Some researchers claim evidence for a weak planetary field, with magnetic moment $\lesssim 10^{12}$ Tm^3; others think that the measured fields were magnetosheath field lines compressed by the solar wind and draped across the planet. Hopefully, measurements on the upcoming Mars Orbiter will prove definitive. Two useful passes of magnetic field data from Mariner 10 showed that Mercury had a magnetic field that could only be explained as due to an internal dynamo field. Its moment is about 10^{13} Tm^3. It is debated whether the dynamo is self-sustaining or driven by thermoelectric currents. Dynamo fields have been found at Jupiter (moment = $1.5 * 10^{20}$ Tm^3), Saturn (moment = $4.6 * 10^{18}$ Tm^3), Uranus (moment = $3.9 * 10^{17}$ Tm^3), and Neptune, but all of these measurements are limited to flyby passes and do not provide enough information to characterize more than the dipole, quadrupole, and some of the octupole fields. Though limited, these data have provided grist for the mill of theoreticians since each measurement has had an element of surprise. Jupiter's magnetosphere is extremely large and corotating at a very fast angular velocity; Saturn's field is considerably weaker than was expected on theoretical grounds and it's tilt is smaller than can be measured ($< 1^o$); the field at Uranus shows strong quadrupole and octupole terms and is tilted at a 60^o angle from the rotation axis; the field of Neptune is apparently similar to that of Uranus.

There is no evidence of dynamo fields from any of the Jovian satellites or from Titan, although the possibility of a dynamo field at Io has not been ruled out.

In each case, the magnetic field measurement has provided key data used in formulating theories regarding deep internal planetary structure. However, except for Venus, the available measurements are inadequate for thorough studies. In principle, measurement of the temporal change of a planetary dynamo is sufficient to determine the radius of its conducting core. This, however, requires higher accuracy data over a longer time period than now available.

There is a good possibility that Mars once had a dynamo field. If so, there may be measurable crustal anomaly fields that, properly mapped, would give clues about the structure and evolution of the Martian crust.

In depth investigations will require planetary orbiters rather than flyby missions. Present plans call for such measurements to be acquired by the Mars Observer, by the Cassini spacecraft at Saturn and by Gallileo at

Jupiter. Unfortunately, Galileo will be in an equatorial orbit and so will not survey the global field distribution of Jupiter. Further, a Phase A study for a Mercury orbiter is scheduled to begin in FY 91. Such missions are to be encouraged. Further, investigations comparing physical processes on the different planets, including the Earth, should be called for by NASA and "Earth scientists" encouraged to participate in both missions and studies.

The data relevant to lunar magnetism are controversial. Some people believe that the paleomagnetic data from lunar rocks, and the presence of magnetic anomalies from impact structures, argue for an ancient lunar dynamo and therefore a liquid nickel-iron core. Others suggest that plasmas induced by the impacts could have produced sufficiently strong fields to have caused the observed magnetizations without the need to call for a lunar dynamo.

SCIENTIFIC OBJECTIVES FOR GEOMAGNETISM

The following objectives have been formulated on the basis of the background presented in the previous sections. Although broad in nature, the objectives will be met through very specific analyses, the outline of which is set forth in the section on Investigation Procedures.

Objective 1: Main Field Modeling: Accurate determination of the Gauss coefficients is a prerequisite for all scientific studies involving the main field. Secondarily it is important for such applications as derivation of magnetic charts, computing charged particle trajectories in the magnetosphere, small craft navigation and background removal for measurements of lithospheric fields. The first objective of the Geomagnetic Studies is to derive an accurate description of the main magnetic field and its secular variation for all epochs possible.

Objective 2: Secular Variation and LOD Variations: Accurate determinations of length of day (LOD) variations are obtained today with very long baseline interferometry (VLBI) and Satellite and Lunar laser ranging (SLR, LLR). It has been shown that on time scales of weeks to a few years, LOD variations are remarkably well correlated with the variations of the angular momentum of the atmosphere. This is interpreted as an exchange of angular momentum between the atmosphere and the Earth's mantle. On a larger time scale, from years to decades, there appears to be a correlation between the secular variation of the magnetic field, in particular the westward drift and the change of the dipole, and the variations of the length of the day. Such a correlation could be accounted for by electromagnetic coupling between the core and the conducting lower mantle. For example, it has been proposed that LOD variations follow secular variation changes with a time delay of the order of 10-15 years. If this correlation, which is debated, could be substantiated and the time delay more precisely determined, or if a different (reversed) correlation, proposed by some, could be established, it would give information both on the lower mantle conductivity and on the strength of the toroidal field at the core-mantle boundary. This latter field has no surface magnetic expression, and hence is not accessible to direct magnetic measurement. Further, if there is such coupling, then there is also a possibility of a connection between the secular variation of the geomagnetic field and excitations of the Chandler wobble. The second objective of the Geomagnetic studies is to investigate the correlation between the geomagnetic field, variations in the LOD, and the Chandler wobble.

Objective 3: Outer Core Fluid Motions and Earth Rotation: As previously mentioned, the secular variation is believed to be produced by the motion of the fluid in the outer core. Therefore, secular variation data provide information about the kinematics of this motion and a starting point for understanding its dynamics. Eventually, we may learn its energy source and whether the fluid in the upper core is stably stratified. This influences the net rate at which heat is transferred out of the core into the deep mantle. Knowledge about the core motion will also make a valuable experimental contribution to dynamo theory. Therefore, the third objective of the Geomagnetic Studies is to study properties of the fluid core.

Objective 4: The Electrical Conductivity of the Mantle: The electrical conductivity of the mantle is an important physical parameter for geodynamic studies. Indeed, it may provide information regarding the thermal state of the mantle and thus the convection pattern therein. Also, in order to estimate the motion $\underline{u}$ at the top of the core the fields $\underline{B}$ and $\partial_t\underline{B}$ have to be continued downward to the Core-Mantle boundary (CMB). Uncertainty in this computation arises from the extremely poor knowledge of the conductivity of the mantle; presently available models for the conductivity of the lowest mantle differ by several orders of magnitude.

Conceptual advances in this subject are needed; however it is already clear that the problem cannot be solved without good values of $g_n^m(t)$ over a period of decades. The fourth objective of the Geomagnetic Studies is to study the conductivity of the mantle.

Objective 5: Geologic and Geophysical Models of the Crust and Upper Lithosphere: Magnetic anomalies reflect important geologic features such as composition, temperature of rock formation, depth to Curie isotherm, remanent magnetism, and geologic structure (faulting, subsidence, etc.) on scales from local to global. Satellite measurements are limited to regional and global scales. Magnetic field data can help delineate the fundamental structure of the very old crystalline basement underlying most continental areas. Magnetic fields can also elucidate facts about ocean lithospheric structure. Long wavelength anomalies due to time periods of constant geomagnetic polarity have been detected in satellite data. Shorter wavelength observations will allow finer scale features to be seen. Also, some areas of the oceanic crust seem to have become demagnetized due to large thicknesses of sediment, driving the crustal temperature above the Curie point. Thus magnetic anomalies, when correlated with other geophysical and geologic information, furnish information important to understanding the evolution and state of the lithosphere. The fifth objective of the Geomagnetic Studies is to model the state and evolution of the crust and upper lithosphere.

Objective 6: Ionospheric, Field-aligned, and Magnetospheric Current Systems: Field-aligned currents (FAC's) play an important role in the coupling of energy between the distant magnetosphere and the lower ionosphere and atmosphere and form the basis of the three-dimensional magnetospheric current system for a wide range of conditions. These currents are also the critical ingredient in a wide variety of auroral processes including particle acceleration and wave generation. The basic characteristics of FAC's have been determined with previous near-Earth satellites, but the relationship of these currents to interplanetary phenomena and the primary generation mechanisms for these currents are not well known.

The characteristics of large-scale currents that flow in the ionosphere have been studied for nearly a century using ground-based magnetic field observations, but few satellites have obtained magnetic field observations with the accuracy and knowledge of attitude and baselines needed to measure these currents. Observations from Magsat have demonstrated that these ionospheric currents can produce significant magnetic perturbations at low satellite altitudes and that, at high latitudes, these ionospheric currents are intimately related to the FAC system. It was also shown from Magsat data that a current of a few million amperes flows in an antisunward direction below the satellite altitude during storms but is absent during magnetically quiet times. This is consistent with a partial ring current model with closure through the ionosphere. An understanding of these currents is critical not only to an accurate representation of the core and lithospheric magnetic field, but also to our understanding of the coupling of energy from interplanetary space into the magnetosphere and into the Earth's atmosphere.

Low-latitude ionospheric currents, the Sq and equatorial electrojet systems, are prominent at sunlit local times. Low-latitude meridional currents have been postulated to be a result of the upwelling of plasma at the magnetic equator. These meridional currents were first measured by Magsat at dawn and dusk local times. The sixth objective of the Geomagnetic Studies is to measure and characterize field aligned and ionospheric currents and to understand their generation mechanisms and their role in energy coupling in the interplanetary-magnetospheric-ionospheric system.

Objective 7: Measurements of Planetary Fields: Physical processes vary from planet to planet and both the differences and similarities are important to our understanding of those processes. Just the existence or nonexistence of a planetary dynamo field places requirements on any theory regarding the internal structure of the planet. Measurement of the fields from different dynamos reveals something of the range of characteristics which the theory must take into account. Lithospheric field characteristics give information regarding lithospheric structure and evolution and about the past history of the planet's interior. The seventh objective of the Geomagnetic Studies is to cooperate with and support programs to measure and model the magnetic fields of planetary bodies.

ELEMENTS OF A GEOMAGNETIC PROGRAM

Based on the preceding paragraphs, this section outlines the elements needed in a balanced Geomagnetic Program. It is to be emphasized that such a program must be part of a broader program in modeling the Earth and its environment as a whole. The proper interpretation of geomagnetic data requires simultaneous understanding of related geophysical data, including Earth rotation, polar motion, seismic data (including tomography), heat flow, gravity, and topography, and other relevant data types including laboratory measurements.

Measurements: Measurements of the geomagnetic field are routinely made at the standard observatories. Most data are sent to and archived at the National and World data centers.

Periodic field surveys are conducted by various organizations around the world, on the ground, by aircraft and by ship. Again most of the data are archived at the data centers mentioned above.

NASA can play a positive role with respect to the data collecting and archiving organizations. Both the observatories and the data centers are undersupported. As a user, NASA should assume some of the responsibility for encouragement and support of these facilities. Organizations, both domestic and foreign, who acquire data should be urged, by NASA, to provide data rapidly to the appropriate data center. The development of methodologies for data reduction and quality control by data users should be coordinated with and made available to those engaged in data reduction and archiving.

However, ground based magnetic field measurements cannot do the whole job. The distribution of magnetic observatories is limited to land areas. Within the land areas they are concentrated in the more developed nations, e.g. Europe, North America, Japan and Australia. This situation is partly addressed by aeromagnetic and shipborne surveys. However, such surveys provide only scalar measurements, with limited additional geographic coverage and then only at particular epochs and are therefore of limited use in detailed investigation of temporal change.

In order to determine the spatial and temporal spectra of the magnetic field of the core, accurate global vector data are required, extending over a significant time span, ideally decades or more. Logistic, economic and political considerations dictate that such data can only be acquired from space. Previous satellite measurements of the geomagnetic field have made significant contributions. However, many of these measurements were of the field magnitude only, often from satellites whose lifetimes were very short. Accurate global vector measurements over an extended time span are lacking and NASA is in the best position to acquire and make such data available.

This deficiency will be partly addressed by the Geomagnetic Observing System (GOS) investigation on the Earth Observing System (EOS) second NASA platform. That mission is not scheduled for launch until 1998 and is scheduled for termination in 2013. If it comes about, GOS will make a significant contribution to main field geomagnetism. However, the last previous satellite survey was conducted by Magsat in 1980, 18 years prior to the scheduled launch of GOS. The time gap is large and is a limiting factor on the usefulness of the combined data. It is highly recommended that the gap be filled as much as possible by a free-flying mission launched before Eos. One candidate is the MFE/Magnolia mission under study jointly by NASA and CNES, the French space agency. With possible launch in 1994, MFE/Magnolia would significantly lessen the gap between Magsat and GOS and, with GOS, would provide 19 years of continuous monitoring of the geomagnetic field. That is more than 1.5 solar cycles, which is desirable. The other candidate is the Aristoteles gravity gradient mission under planning by the European Space Agency (ESA). This is planned for launch in 1994 or 1995 into a near polar orbit. The mission profile includes several months at a high altitude followed by at least six months at 200 km after which the altitude would be raised to 600-700 km and data acquired for about another three years. Studies are underway to evaluate the feasibility of including a magnetometer experiment as part of the mission.

Aristoteles would acquire data useful not only for studies of the main field, but would also provide the best available data for lithospheric studies. At 200 km the anomaly data would have amplitudes of several times and a resolution at least twice as fine as that of Magsat. This mission, or one like it, is clearly the next step in mapping and understanding large scale magnetic anomalies on a global scale.

Another possibility for obtaining short wavelength data to study lithospheric magnetization is to put a magnetometer on the Superconducting Gravity Gradiometer Mission (SGGM). This is due to fly in 1999, in a near 200 km, high inclination orbit with a lifetime of six months.

As pointed out in the Background, the geomagnetic field has time scales from years to centuries. If we are ever to really understand the geodynamo measurements will have to be made over the longest as well as the shortest time scales. Such measurements need not be continuous. For time scales of centuries, samples at 10-15 year intervals would be adequate. However, thorough study of periods up to 100 years dictates that at some time a nearly continuous span of data spanning a century is needed. It is highly recommended that NASA immediately follow up the GOS experiment with follow on missions to extend the time interval of continuous coverage to at least two solar cycles, preferably more. Plans should then include periodic missions at 10-15 year intervals until beyond the end of the 21st century.

There has been some success in determining the Earth's field during the past few centuries by reanalyzing old data, which in the early days was almost entirely shipborne inclination and declination data. The farther back in time this is attempted, the longer the minimum wavelength recorded in the data and the greater the time smoothing.

It is possible to get a very rough idea of the temporal change of the magnetic field by analyzing paleomagnetic samples, in particular those from lake sediments and from archeological remains such as baked hearths and kilns. Although these data are of much lower quality than directly measured components of the field, it is possible that with enough results the time varying dipole and quadrupole components of the field could be established.

At some time during the period of continuous data acquisition, an attempt should be made to separate the external and internal fields for an extended period of time. This will require multiple satellites in orbits of differing local time together with a massive campaign for simultaneous acquisition of global surface measurements.

Acquisition of satellite data in no way negates the value of surface data. The observatory data, in particular, are as important to satellite observations as vice versa. The variations of several hundred nT in less than an hour which occur during magnetic storms would be very difficult to interpret from satellite data alone. Such time scales can result in serious aliasing of the satellite data. Further, definitive separation of fields from ionospheric sources both from fields of magnetospheric origin and from fields of origin internal to the Earth requires measurements both below and above the ionosphere. A goal of the Geomagnetic Studies program should be to enlarge and enhance the array of sites at which absolute magnetic measurements are acquired on a continuous basis. There are two concerns to be addressed to accomplish this. First, technology for making such measurements on the ocean bottom needs to be developed and implemented. The most difficult part of this task is the problem of measuring the attitude of the instrument relative to three known axes so that the field components can be measured. The second concern is the reliable acquisition of data in third-world countries, especially those prone to governmental unrest. This problem is at least partially addressed by the recent and continuing development of relatively low cost, automated, vector instruments. Such are in use and/or under test at several of the national observatory networks. However their development does not guarantee their deployment and operation. That involves political as well as scientific issues which we would encourage NASA to address.

Data deficiencies plague the study of lithospheric fields. Aeromagnetic surveys usually (though not always) are of too limited an area to study anomalies over regions tens of kilometers in size and almost never cover enough area to study regions hundreds of kilometers in size. Piecing together smaller surveys results in distortion and aliasing, particularly if the surveys were acquired at different epochs and altitudes, and the main field has changed significantly. Satellite data are global in nature, eliminating problems of political boundaries and logistics. Data from the POGO and Magsat satellites have demonstrated that lithospheric fields can be studied from satellite. However, these analyses have also shown that lower altitude data are needed to bridge the gap between the shorter wavelength features measured by aircraft and ships and the very long wavelengths measured in currently available data. At present, no missions at lower altitude than Magsat are approved. Studies are underway to investigate the feasibility of including a magnetometer on the Aristoteles mission of ESA, to fly at about 200 km altitude in about 1996. Such a mission is highly desirable for lithospheric anomaly mapping.

For core studies, satellites must collect vector measurements of the geomagnetic field. Even perfectly accurate scalar data can fail to determine an external harmonic potential field. Numerical experiments suggest that scalar data can determine such a field if it is nearly dipolar and some vector data are available on the surface at the magnetic equator. The USGS is installing twelve such surface stations. Unfortunately, the results from these numerical experiments must be used with caution since most surface stations lie on local magnetic anomalies which can be several percent of the main dipole field. These anomalies can be measured by means of a campaign of vector satellite measurements, and once measured they can contribute to a subsequent scalar campaign. MAGSAT estimated the anomalies under the observatories extant in 1980, but such information will not be available for the new USGS observatories. A further difficulty is that the numerical experiments assume no external sources. At the magnetic equator, the external sources associated with Sq and the equatorial electrojet are particularly strong. Moreover, at satellite altitudes, electric currents can produce one percent deviations of the geomagnetic field. At the accuracy required for core studies, vector measurements of the field are essential. The coverage must be global and of high accuracy. Each component of the field should be measured to at least 3 nT, including instrument, position, and attitude errors. The scalar field should be measured to 1 nT to serve as a check on drift in the vector instruments.

Another possibility for studying shorter wavelength features is to use a magnetic gradiometer. The gradient of a periodic magnetic field of a certain wavelength is obtained by dividing by the appropriate wavelength. The result of this is that if a structure gives a peak in its field power spectrum at a certain wavelength, the peak in the power spectrum of the gradient will be shifted to a shorter wavelength. Of more importance is that the core field is of very long wavelength, some external fields have very long wavelength signals, and local fields from field aligned currents contribute a curl in the gradient measurement. Thus, measurement of the total gradient tensor should, in principle, permit more accurate separation of fields from these sources from fields due to lithospheric sources. All nine components of the tensor must be measured because currents do flow at the satellite position, so that the field is not curl free. Gradient measurements should be taken in addition to, not in place of, the usual three component field measurements.

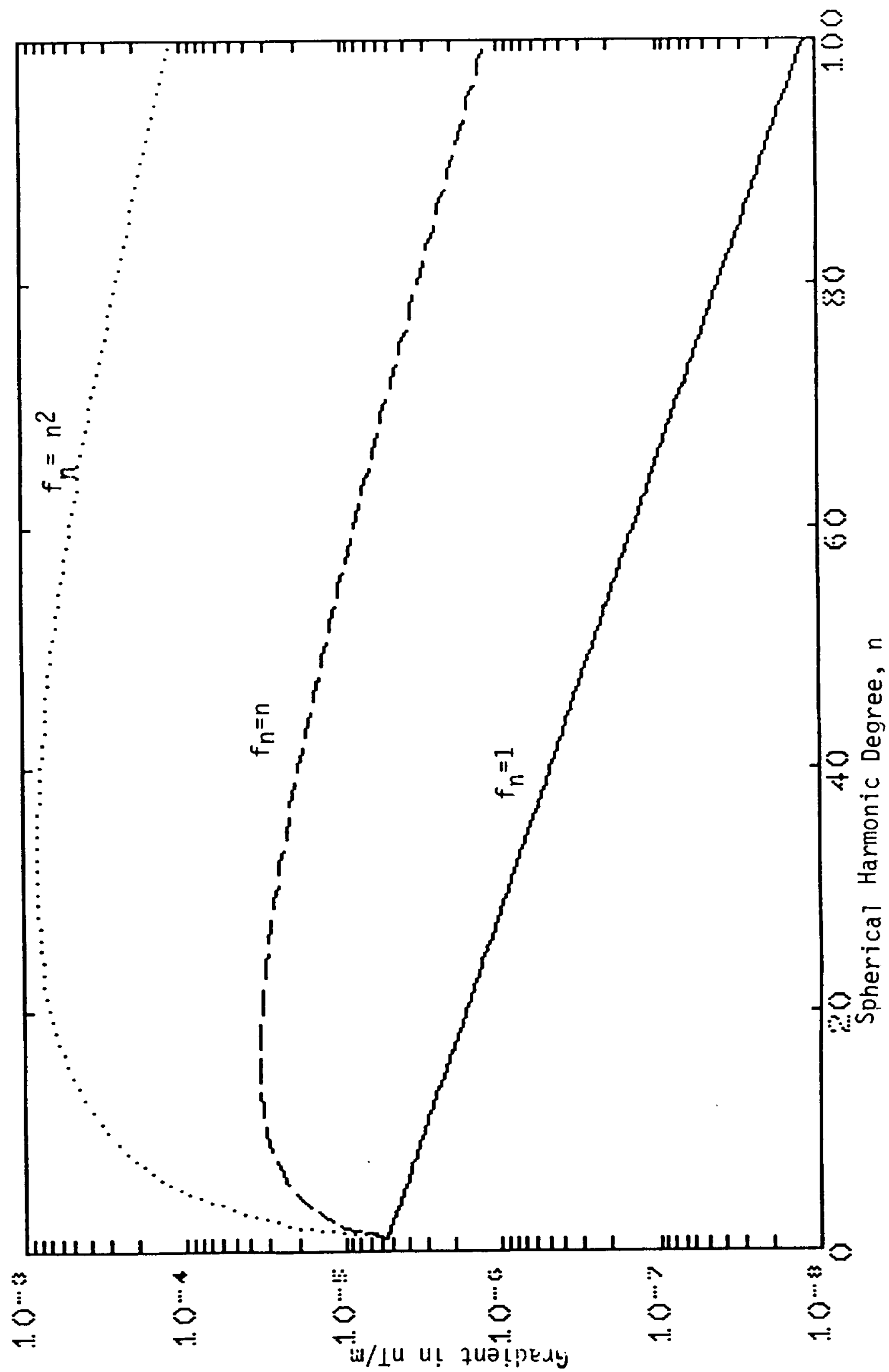
The sensitivity requirements for measuring gradients due to lithospheric fields can be roughly estimated by consideration of a spherical harmonic representation of the field potential, e.g. equation 2. In this case, for each degree and order, the elements of the gradient tensor have the form,

$$T_{ij} = f_{nm} c(m\phi) L_n{}^m(\theta) [a/r]^{n+3}/a, \tag{4}$$

where a is the mean Earth radius, $c(m\phi)$ is the sine or cosine function, n and m are the degree and order, and L is an associated Legendre function or its first or second derivative. The quantity f_{nm} is a multiplying factor which depends upon the degree and order; for large n it goes as n^2, n, m, nm or 1. For estimation purposes, neglect c and $L_n{}^m$ and assume that the gradient goes as

$$G = f_n [a/r]^{n+3} [R_n]^{1/2}/a \tag{5}$$

where f_n is taken to be 1, n or n^2 and where R_n is the lithospheric component of the spectrum shown in Figure 2. Using a conservative estimate of R_n, the gradients computed from equation (5) are shown in Figure 6. The lower curve is for $f_n = 1$ and gives an estimate of the sensitivity of gradient measurement necessary if the smallest meaningful gradients are to be detected. The top curve is for $f_n = n^2$, which also gives information about terms in n*m, and applies particularly to radial gradients of the radial and eastward field and to longitudinal gradients of the radial and eastward field. The most difficult gradient to detect will be the latitudinal gradient of the north component. From this figure it is apparent that useful gradient information should be obtainable with sensitivities of 10^{-4} nT/m and better but that sensitivities of the order of 10^{-7} will be necessary to detect the smaller gradients.



1.0⁻³
1.0⁻⁴
1.0⁻⁵
1.0⁻⁶
1.0⁻⁷
1.0⁻⁸
Gradient in nT/m
0
20
40
60
80
100
Spherical Harmonic Degree, n
$f_n = n^2$
$f_n=n$
$f_n=1$

Figure 6

The gradients from field aligned currents will be larger, often reaching 10^{-2} nT/m or higher. Detection of the main features of in situ currents in the auroral belt should be possible with gradiometers of minimal sensitivity. The gradients expected of meridional currents at lower latitudes are unknown.

Magnetic gradiometers have been developed using superconducting quantum interference device (SQUID) magnetometers. A squid magnetometer designed as a gradient device measures the difference in flux linkage through two parallel coils separated by a known distance. Sensitivities as low as $5*10^{-7}$ nT/m have been claimed in the laboratory. In order to improve sensitivity it is necessary to increase the area of the coils and/or increase their separation. The sensitivity is directly related to the product of these quantities, i.e. to the volume of the instrument.

Further studies of the usefulness, accuracy and sensitivity requirements of gradient measurements are called for. These should include meaningful simulations of realistic lithospheric features. It should also be kept in mind that the presence of field aligned currents at satellite altitude is both an opportunity and a complicating factor. In situ measurements of the curl of the field will give the first direct measurement of the total current density, a key quantity in determining the energy input into the auroral belt and in studying ionosphere-magnetosphere coupling. However in the regions of such currents, their gradients will likely dominate the measurements and mask gradients from lithospheric sources. Also, such currents are likely to be highly filamentary, at least in some locations. In that case, the distance between the parallel coils will, in part, determine if one is measuring a very local phenomenon, i.e. filamentary current, or a broad average current. Finally, gradient measurements in a plasma are particularly sensitive to local effects of the probe and vehicle, which may deflect the current carriers. This question needs careful study.

A third way of collecting better information on the lithospheric field is by using a tether from another spacecraft, such as the Space Shuttle. Tethers can be 100 km long; if then the shuttle is orbiting at an altitude of 200 km, the magnetometer would be only 100 km above the Earth's surface. This altitude would give tremendously higher resolution than Magsat. However, the global coverage would be lost, for two reasons. Firstly, in order to make use of the lower elevation, a mission would have to continue for twice as long as Magsat to obtain adequate spacing of flight lines to make full use of the lower elevation. Secondly, the present inclination capability of the shuttle is limited to below about 57°, excluding data acquisition at the higher latitudes regardless of the flight length. Nevertheless, certain low latitude targets could be very well surveyed in a few days if the mission coverage profile were well planned.

An additional complicating factor is that the measurements would be acquired near to or in the actual location of E region ionospheric currents. This, of course, presents an unprecedented opportunity to study those currents. Study of the data separation and analysis problems should be carried out as part of the feasibility studies for such an experiment. Preliminary flights of tethered magnetometers are expected in 1991 or 1992, but these are engineering flights and of little scientific value.

While it is clear that the scientific objectives set forth issue a clarion call for additional data, particularly from satellites, this is not to negate the usefulness of existing data. It may be true that the most obvious studies and applications of the existing data have been carried out, but there are still many questions that can be addressed with existing data. If anything, such analyses require increased support just because the easy work has been done. Examples of such ongoing studies are refinements of the decomposition of satellite data into core, lithospheric and external parts; improvements in analysis and interpretation of the lithospheric signal; optimization of field models by refinement of the external field representation; and application of new error analysis methods so that the true extent of our knowledge of inner Earth parameters becomes clear.

Main field and large scale external field modeling is the backbone of much of the data analysis. This will be accomplished as an extension of currently ongoing research. Models will include internal, induced, and external fields and will describe those fields as a function of time. Where possible, earlier data from satellites and/or surface measurements will be combined with future data to form the longest possible time series. The ability of any inverse problem to describe the real world depends both upon an accurate parameterization of the problem and on the observability of the phenomena in the data. Satellite data result in good geographic data coverage in a period of time short compared to significant variations in the core and lithospheric fields. This results in good observability. But the external fields are relatively fixed in local time with substantial variability in universal time. A single satellite cannot isolate these fields. Even with proper parameterization, no model is able to represent the combined internal and external fields to the accuracy of the data; from a single satellite the parameters are simply not observable. Multiple satellites are required, e.g. three in orbit simultaneously at evenly spaced local times.

Studies of the core, mantle, and Earth rotation include theory and numerical experiments. The following paragraphs describe how surface magnetic data are used to study the fluid motion just below the core-mantle boundary (CMB). Estimates of the electrical conductivity in the lower mantle range from 10^{-2} to 10^{5} Ω^{-1} m^{-1}; clearly it is not well known. Electrical conductivity in the core is about 10^{6} Ω^{-1} m^{-1}. In discussing magnetic fields whose length scales are of the order of the Earth's radius and whose time scales exceed a few years, the mantle can be treated as an insulator and the core as a perfect conductor. If $\underline{B}_n$ and $\partial_t\underline{B}_n$ from the core can be measured for $1 \leq n \leq N$ on the Earth's surface, $\underline{B}$ and $\partial_t\underline{B}$ can be extrapolated down through the mantle and computed at the CMB with a circle of confusion whose radius is about 180/N degrees. Since B_r and $\partial_t B_r$ are continuous across the CMB, their large scale structure is known in the fluid just below that boundary. Treating the core as a perfect conductor with velocity $\underline{u}$ implies that $\underline{B}$ is advected with the fluid. In consequence, just below the CMB, where $u_r = 0$,

$$\partial_t B_r + \underline{\nabla}_H \bullet (B_r \underline{u}) = 0 \tag{7}$$

here $\underline{\nabla}_H$ is the horizontal gradient. Equation (7) is one scalar equation connecting the two horizontal components of $\underline{u}$, u_θ and u_ϕ. Clearly (7) cannot determine $\underline{u}$ uniquely unless other assumptions are made about $\underline{u}$ or other information is available. One assumption now being tested, which would determine $\underline{u}$ from (7), is the possibility that $\underline{u}$ is steady during intervals of a few decades or less, (to be precise, $b|\partial_t \underline{u}| << |\underline{u}|^2$, where b = core radius). One source of additional information about $\underline{u}$ is the momentum equation. It has been argued, for example, that $\underline{u}$ is geostrophic except near the equator. Although this does not suffice to determine $\underline{u}$ from (7) everywhere at the CMB, it does greatly reduce the non-uniqueness. Another assumption commonly adopted to minimize the non-uniqueness is to make the spatial variation in u_ϕ/sin(theta) as small as possible and to minimize u_θ. This is the westward drift hypothesis.

Obviously, good values of $\underline{B}_n$ and $\partial_t \underline{B}_n$ near the Earth's surface for $1 \lesssim n \lesssim 12$ (the degrees for which lithospheric fields can be neglected) are essential to these investigations. Furthermore, near-surface magnetic data can be used to test the assumptions underlying the foregoing discussion: is the conductivity of the lower mantle negligible in such calculations (present data suggest it could be)? Is flux conserved (mean field dynamo theory suggests not, but present data suggest it might be)? Is the motion approximately steady (present data suggest it could be) or geostrophic (present data suggest that it could be except in a band several degrees wide at the magnetic equator)? Such questions probably cannot be settled without several decades of satellite observations.

Experimental studies will utilize the results of the main field and secular change modeling. Models of the fluid flow near the CMB can be derived directly from the data or from the analyzed data by numerical application of geophysical inverse theory. For example, the non-linear inverse problem posed by equation (7) and the hypothesis of piecewise steady flow is solved iteratively for $\underline{u}$ using a weighted least squares algorithm with options for requiring a geostrophic radial vorticity balance and spatially smooth flow. Hypotheses such as steady flow are tested by determining if the resulting flow predicts an adequate fit to the evolving Gauss coefficients. Corrections for mantle conductivity or core ellipticity, or some allowance for core resistivity, can in principle be made to make such tests more definitive. Success or failure of a hypothesis demands improvement of the means to test it or the geophysics supporting it. This systematic approach provides a basis for improved scientific understanding of the core geodynamo; it neither requires nor excludes major breakthroughs in theory, but does require global, long term, quality data.

Geomagnetic field behavior (e.g., westward drift) apparently correlated with changes in the LOD are calculated from the evolving Gauss coefficients. These coefficients and improved mantle conductivity estimates lead to improved estimates of electromagnetic core-mantle coupling. Topographic core-mantle coupling can be calculated from seismo-tomographic estimates of the CMB topography and any geostrophic core surface flow. The latter implies a specific non-hydrostatic pressure field at the top of the core which exerts a mechanical torque on the CMB topography. This torque predicts changes in the angular velocity of the Earth which can be compared with Earth rotation data to test the agreement of fluid flow and topography models.

All of the analyses described above contribute to our knowledge of the state of the core and the functioning of the geomagnetic dynamo. Similar analyses should be applied to data from other planetary bodies, as the data quality and amount permit. Comparative studies will then provide additional constraints for the theory of planetary interiors in general and for the possible modes of operation of planetary dynamos in particular.

Conductivity models will be derived for the mantle and crust. There are two ways of determining the conductivity of the mantle. It can be probed "from below" using signals originating in the upper core and observed at the surface. This method requires a precise determination of the Gauss coefficients during rapid and isolable events such as the 1969 jerk. It can also be probed "from the top" by the analysis of the external and internal (induced) parts of $\underline{B}$ at various frequencies. This last method also requires a good knowledge of the time dependence of the Gauss coefficients. The penetration depth, p, of a signal at a given period T in a layer of conductivity σ is given by $p = (T/\pi\mu_0\sigma)^{1/2}$. If data were available for one or more 11 year sunspot cycles, it might be possible to obtain the conductivity of the lower mantle without making prior assumptions about the kinematics of fluid motion at the top of the core, as is required in the method "from below".

If an abrupt impulse, or jerk, is measured in an otherwise slow change, then mantle filter theory can be applied to constrain mantle conductivity. If not, then the frequency dependent amplitude ratios and phase shifts of separated, long period external and externally induced internal fields will still constrain laterally homogeneous estimates of the mid and deep mantle conductivity. Furthermore, the cross correlation between external and internal Gauss coefficients with different degrees and different longitudinal orders can test the importance of lateral variations in mantle conductivity. More detailed interpretation of the analyzed data in terms of laterally heterogeneous mantle conductivity and other mantle properties requires more detailed theoretical studies. Acceptable separation of externally induced internal fields from core induced fields at periods exceeding 1 year may require iteration between core flow and mantle conductivity models. If the mantle conductivity is low enough, and if the core current boundary layer is weak enough, then it may be possible to probe motions below the CMB using horizontal field components. If not, then there may be a strong toroidal magnetic field in the deep mantle.

For determining conductivity models of the crust and upper mantle, data from individual satellite passes will be closely coordinated with simultaneous ground-based observations at standard magnetic observatories and temporary variometer sites. Surface maps of magnetic transients will be constructed in space and time at ground level which can be compared against profiles of data sampled by overhead satellites. Algorithms exist for interpolating ground-based data along the flight path of the satellite at its so-called "foot-print". A number of efficient local interpolation forms have been developed. Some of these methods have been recently extended to allow for a variety of **a priori** constraints on the attributes of 2-D smoothing polynomials (e.g. smoothness, flatness, etc.).

Typical satellite data samples signal periods extending from a few minutes to many months. This encompasses Sq and broad-band Dst at latitudes of $\pm 45^{\circ}$, auroral substorms at higher latitudes, and pulsations at all latitudes. Because ionospheric current systems will generally be found between the satellite and the ground sites, one has to account for these intervening sources in the analysis. This problem will be explicitly addressed as a joint generalized inversion, where both satellite and ground based data are used to solve simultaneously for the distribution of primary current sources in space and for the induced secondary sources in the solid

Earth. The latter, of course, leads to the conductivity as a function of depth and, with refined modeling, to lateral variations in that conductivity.

The mathematical formulation involves the inversion of the observed vector field components in space and time directly, rather than, as is conventionally done, going through formal spherical harmonic and Fourier decompositions of the observed data. The direct approach is appropriate since it is well known that one can effectively estimate vertical conductivity profiles without full knowledge of the source field geometry; one can do so knowing only the actual field components and their horizontal first and second derivatives over a restricted region of the Earth's surface. These data parameters will be used in a maximum-likelihood inversion scheme that minimizes a weighted L_2 norm of the prediction error and one of several minimization conditions on the solution simplicity (e.g. L_2 norm of the solution length, model flatness, or model roughness).

Continued laboratory measurements of the electrical conductivity of analogs of lower mantle materials are required to support both theoretical studies and geomagnetic deep sounding results.

Models of the lithosphere are the objective of the study of lithospheric anomaly fields. There are many kinds of models; their common purpose is to generalize observations and make predictions. In order to arrive at a most comprehensive model, and to minimize model non-uniqueness, other data must be used. Figure 7 shows the sorts of data which can be used to constrain lithospheric models, and their relationship to satellite magnetic field or gradient measurements. A first attempt to use magnetic data might be to construct an equivalent source model. Then application of various amounts of annihilator might be done to arrive at a magnetization model which makes most sense in light of the other data available for the area under consideration.

Equivalent source models require **a priori** selection of the direction of magnetization. It is possible that information might be gained about magnetization direction by producing equivalent source models for a variety of directional constraints on the magnetization and comparing these to what is known about the structure from gravity or seismic information. Alternatively, if the shape of a magnetized body can be determined **a priori** from other information, then an inversion may be done on lithospheric magnetic anomaly data to obtain magnetization vector information. This method has been successfully applied to seamount magnetization, but it could, in principle, be applied to satellite information over well defined magnetized bodies.

Application of model constraints from seismic, gravity and heat flow measurements to magnetic models is perhaps most easily applied in forward rather than inverse modeling. Prior inverse modeling can furnish guidance to the magnetization distribution to be chosen in a forward model. Forward models have the advantage of greater ease of representing sharp boundaries delineated by seismic and aeromagnetic data and by known tectonic features discovered in surface exploration.

Models must be constrained by knowledge of the magnetic properties of the rocks responsible for the magnetic anomalies. This class of investigations is of scientific importance and wide applicability in its own right, but its specific importance to NASA programmatic goals is clear: without such "ground truth", modeling of magnetic anomaly data is lacking an important physical constraint.

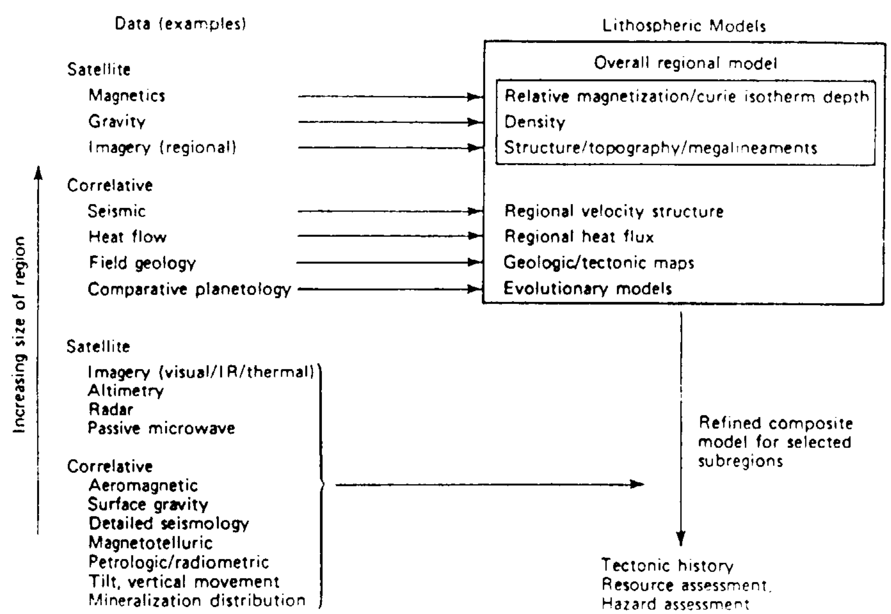


Fig. 5—Synthesis of geologic/geophysical models (idealized outline).

Fiqure 7

Stimulated by the Magsat mission, laboratory studies of deep crustal rocks began at Goddard Space Flight Center. A new picture of deep crustal magnetization has emerged from these studies, and has significantly aided geologic interpretation of Magsat data. But at the same time a number of important scientific controversies have developed. Resolution of these controversies will have significance not only to satellite data interpretation specifically, but to several fundamental questions of the petrology of the lithosphere and its evolution. Examples of these questions are: Given the evidence for enhanced magnetization (several A/m) at deep continental crustal levels, is this due to viscous enhancement of magnetization, and if so what petrologic evolution is implied by the required distributions of mineralogies and magnetic domains? Can lower crust formed by prograde metamorphism of rocks of sedimentary parentage be distinguished from lower crust formed by magmatic underplating through contrasting magnetic petrologies and associated long wavelength magnetic anomaly fields? Can regions of the upper mantle have anomalous magnetic properties because of large-scale variations in chemistry or oxidation state? If so, under what conditions should the continental magnetic layer be associated with the entire lithosphere instead of just the crust? Is it generally true, as has been suggested, that processes in the deep crust related to chemistry and oxidation state tend to produce magnetite Curie points near 550^{o} C, or are there significant regions characterized by low Curie points? The question is relevant not only to problems of petrology and chemical flux in the deep crust, but to models of crustal temperature. These questions can be answered only through considerable further sample collection and laboratory measurement. For the continental areas, results are still relatively few, and ongoing support is essential.

There are considerably more data on the rock magnetic properties of the oceanic lithosphere both from direct measurements on samples collected from the ocean basins by drilling and dredging, and by sampling ophiolite suites on land. Further work is necessary to determine how the magnetic properties of ophiolites are changed during the process of emplacement of these oceanic lithospheric sources into continental blocks. Controversy exists concerning the location of the magnetization resulting in the long wavelength field measured in satellite data. Some would maintain that crustal magnetization may be sufficient, while others insist that a large part of that magnetic field must originate in the lithosphere below the crust. Also, controversy still exists as to the source of the short wavelength sea floor spreading anomalies, some calling for magnetization through the total oceanic crust, while others require only a 0.5 km thick layer composed of the pillow lava sequence of seismic layer 2A. Long wavelength magnetic anomaly studies, more rock magnetic measurements, and emplacement models are necessary to unscramble this important problem. The problem of oceanic lithospheric magnetization is also made more complex in that both remanent and induced magnetization may be important, especially for the deeper seismic layer 3. Serpentinisation of ultramafic rocks or of olivine crystals with the gabbroic rocks of layer 3 can also cause complexities because this process creates secondary magnetite which can produce very high secondary magnetization in the serpentinites, whose time of acquisition may be uncertain.