

Papers Presented to the
**CONFERENCE ON
ORIGINS OF
PLANETARY MAGNETISM**

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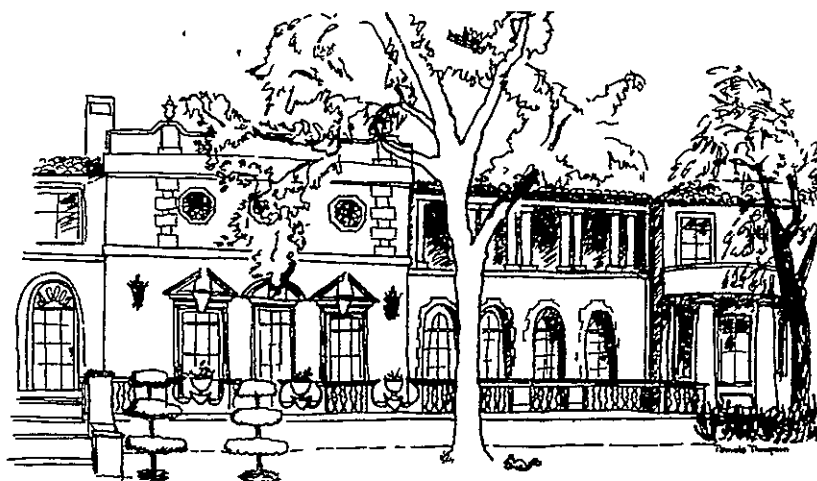
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P R E F A C E

This volume contains papers which have been accepted for publication by the Program Committee of the Conference on the Origins of Planetary Magnetism. Papers were solicited which address one of the following major topics:

- I. Geomagnetism in the context of planetary magnetism
- II. Lunar magnetism
- III. Dynamo theory and non-dynamo processes
- IV. Comparative planetary magnetism (terrestrial and outer planets)
- V. Meteoritic magnetism and the early solar magnetic field.

The Program Committee consists of L. J. Srnka, Convenor (*Lunar and Planetary Institute*), P. R. Brett (*National Science Foundation*), F. H. Busse (*University of California*), R. B. Merrill (*Lunar and Planetary Institute*), N. F. Ness (*NASA Goddard Space Flight Center*), S. K. Runcorn (*The University, Newcastle upon Tyne*), and C. T. Russell (*University of California*).

Logistic and administrative support for this conference has been provided by P. P. Jones (*Administrative Assistant, Lunar and Planetary Institute*). This abstract volume has been prepared under the supervision of P. C. Robertson (*Technical Editor, Lunar and Planetary Institute*).

Papers are arranged alphabetically by the name of the first author. A subject and author index are included. The subject index, based on key words returned by the authors, does not represent the total number of papers in this volume.

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MAGNETIC PROBING OF PLANETARY INTERIORS; E. R. Benton,
Univ. of Colo., Boulder, Colo. 80309

This paper addresses the following general question: what can be learned about a planetary interior from measurements at the surface of the global planetary magnetic field? The specific issues considered in detail are those prompted by pressing problems in the dynamo theory of geomagnetism, but most are clearly of interest for other planets as well. Attention is restricted to the idealized but representative situation of a spherically symmetric planet containing a highly conducting liquid core surrounded by a nearly insulating solid mantle, the rest of space being vacuum. Core conductivity is presumed constant, while mantle conductivity, a function of radius only, is everywhere small in comparison, falling monotonically from its largest value at the core-mantle boundary to zero at the planetary surface.

The planetary magnetism in question then arises entirely from dynamo action in the liquid core, being devoid of external sources. This is probably too primitive a model to permit direct application to such objects as Jupiter or the sun, where external currents are also important, but the model should be relevant for planets such as Mercury, Venus, and perhaps Saturn, as well as Earth. Alternatively, if through spherical harmonic analysis, and/or time averaging, the effects of external sources can be removed from the observations, then the model is relevant to planets which are both internally and externally excited.

The global surface magnetic observations required are spherical harmonic expansions, truncated beyond degree N (that is N^2+2N Gauss coefficients) for the scalar magnetic potential as observed at two epochs separated by a time short compared to the ohmic free-decay time of the planetary core, yet long compared to the time required for secular variations of the magnetic field to penetrate through the mantle.

For such an idealized planet, one can construct a systematic perturbation procedure for analyzing the magnetohydrodynamics of the planetary interior. The procedure, though intricate, is straightforward; it will be reported in detail elsewhere. In essence, three small parameters, proportional to core resistivity, mean mantle conductivity, and variations away from mean mantle conductivity, are introduced to describe perturbations about a ground state in which the core moves like a perfect conductor and the mantle is a uniform insulator. Information about each of the following properties of the planetary interior is then accessible and will be discussed:

- (1) Radius of the liquid core (a result of Hide).
- (2) Depth of substantial vertical motion and intense electric currents in the core.
- (3) Presence or absence of unsteadiness and turbulence in the upper reaches of the core.
- (4) Horizontal fluid motion adjacent to the core-mantle boundary.
- (5) Azimuthal field strength in the upper reaches of the core.

Magnetic Probing of Planetary Interiors
E. R. Benton

- (6) Current system in the mantle and strength of
electromagnetic core-mantle coupling.

This work is supported by the Earth Sciences Section of the
National Science Foundation under NSF Grant DES75-03960.

ON THE GROWTH RATE WITH DEPTH OF HORIZONTAL MAGNETIC FIELDS AT THE SURFACE OF EARTH'S LIQUID CORE; E. R. Benton and L. A. Muth, Univ. of Colo., Boulder, Colo. 80309

A key parameter arising in self-consistent hydromagnetic (as opposed to kinematic) models of the geodynamo is the strength of the toroidal magnetic field developed in the liquid core compared to the poloidal field there. Because it is easy to understand how differential rotation (either with depth or in the horizontal) would shear an initially weak poloidal field into a strong toroidal field, it has been conventional to assume that relatively strong azimuthal fields exist in the core; but their associated electric currents, being precluded from flowing in the nearly insulating mantle, implies that such fields are confined to the core, being undetectable (directly) at earth's surface.

On the other hand, Busse (1975) has made considerable progress with the convectively driven hydromagnetic dynamo (in cylindrical geometry) on the assumption that toroidal fields are no more than comparable with the poloidal fields. In this situation, the Lorentz force acts as a perturbation in the momentum balance, thereby permitting analytic progress to be made.

In this paper, an effort is undertaken to determine how rapidly B_θ and B_ϕ are apt to increase in magnitude with depth at the outer surface of earth's liquid core. The model is based on a nearly insulating mantle, a nearly perfectly conducting inviscid core of uniform density, spherical symmetry, and gravitational driving of unaccelerated flow in the momentum balance. Under these conditions the radial torque balance is magnetostrophic (Coriolis torques counterbalance Lorentz torques). At particular locations on the surface of the core it then becomes possible to evaluate components of the horizontal electric current and subsequently, to infer values for $\partial B_\theta / \partial r$ or $\partial B_\phi / \partial r$. Results obtained from geomagnetic field models at epochs 1965 and 1975 will be presented and discussed. This work is supported by the Earth Sciences Section of the National Science Foundation under NSF Grant DES75-03960.

ANCIENT MAGNETIC FIELDS ESTIMATES FROM METEORITES: A CAVEAT AND HOW-TO GUIDE; Aviva Brecher, Dept. of Physics, Wellesley College, Wellesley, MA 02181 and Dept. of Earth and Planet. Sci., M.I.T., Cambridge, MA 02139

In the two decades elapsed since the exciting discovery of fossil extraterrestrial magnetism in meteorites, a large amount of data has been collected and variously interpreted [e.g., reviews 1,2,3]. The promise and potential which meteorites hold for recovering information on ancient magnetic fields [2-7] are becoming more fully realized, having received renewed impetus from the related riddle of lunar magnetism [7-9]. Since meteorites are extremely well-characterized with respect to equilibration, metamorphic or shock-reheating temperature, their radiometric ages can be determined, and their likely place of origin reasonably well-established, the temporal evolution and spatial configuration of ancient magnetic fields could also be conceivably reconstructed from paleointensity determinations on various classes of meteorites. Their origin in asteroidal, mostly undifferentiated objects is widely accepted, so that any magnetic fields they recorded are probably external, extended solar fields. These objectives can be met provided that stringent sample-selection and data-reliability criteria are imposed and a suitable methodology is followed as outlined below.

The present paper will provide an updated review of and subjective practical guide to paleointensity determinations on meteorites.

(1) Select for magnetic study meteorites for which detailed information on chemical-petrological classification, shock exposure and various radiometric ages already exist. Ideally, data on metal phase structure, abundance, morphology and thermal history should also exist. These will aid in deciding on the appropriate experimental methods and in interpreting magnetic results.

(2) Obtain large samples, typical of the whole (often inhomogeneous) meteorite. It is desirable to examine mutually oriented samples from the same hand specimens, and several from the strewn field, in order to appraise the uniformity of magnetization.

(3) Determine the bulk magnetic properties (e.g., natural remanence NRM, magnetic susceptibility) and compare with other meteorites of the same class, to find out if one deals with a unique or representative meteorite. For example, the ureilite meteorites are extremely coherent magnetically as a group, whereas the achondritic breccias are very dissimilar [2].

(4) Analyze the NRM by standard (cold) methods (e.g., AF cleaning) to isolate a stable and possibly primary component. General criteria for the identification are: hard coercive spectrum of NRM; clustering of NRM directions under AF cleaning, on stereonet projection; and vector component (Van Zijl) analysis, requiring smooth and linear convergence to the origin with progressive AF cleaning [10].

The method of NRM analysis should be appropriate for the magnetic carriers present; thus one could exploit the magnetic memory effect of magnetite in carbonaceous chondrites [2, 10]. Thermal cleaning, as for lunar samples, is counterindicated since heating incurs changes in magnetic texture, phase-structure, and mineralogical-chemical alterations [3, 5-7].

ANCIENT MAGNETIC FIELDS ESTIMATES FROM METEORITES

Brecher, A.

(5) It is desirable to saturate the sample and use the AF coercivity spectrum of IRM_s as an index of the intrinsic capability for stable remanence and as a comparison standard for any alterations due to the process of modelling the NRM in the laboratory.

(6) While realizing that the physical process by which the initial NRM may have been imprinted cannot be duplicated adequately in the laboratory (e.g., sedimentation in low-g, for depositional remanence (DRM); or cooling rates of 0.1-100°C/m.y. with simultaneous phase change, for thermochemical remanence (TCRM)), select the experimental method best suited to the physical history of the meteorite and most likely to reproduce its NRM (e.g., rapid post-shock cooling). This is quite subjective! In any case, one must ascertain that the selected method for NRM modelling does not produce substantial irreversible magnetic alteration, or else the paleointensity estimate is unreliable. For these reasons we have adopted the Van Zijl method for paleointensity determination on meteorites [2, 4, 11], which requires a single heating to a temperature preselected to match information available under (1) above. Other workers prefer the double-Thellier method [5-7]; or other variants involving repeated heating cycles [1,3].

(7) The caveats of modelling experiments are: (a) preconceived ideas regarding the nature of NRM (e.g., thermal) and biasing an open-minded appraisal of results. Thus, every assumption (e.g., NRM is a TRM; a TRM -- and by implication, the NRM -- is proportional to the ambient field at cooling) must be tested experimentally. Such precautions lead to surprising and unexpected results: for example, many lunar rocks and meteorites are not even capable of acquiring a TRM when cooled on laboratory time-scales and fields. Similarly, iron meteorites and many chondrites possess substantial magnetic moments (TRM_o) even when cooled from above the Curie point in field-free space [12]. Hence, even though iron meteorites can well acquire a TRM, the similarity of NRM and TRM to spontaneous magnetization (TRM_o) precludes meaningful paleointensity estimates. (b) preconceptions in the interpretation of magnetic paleofields: is "primary" synonymous with "primordial" magnetization; is a "secondary NRM" always soft, or vice versa? For example, the striking fact is that soft magnetization in carbonaceous chondrites [2, 10] and in some chondrites [2] is often in the same direction as the high-coercivity NRM, so that only a single magnetizing event need be invoked.

(8) Adhere to more stringent experimental tests of the reliability of a paleofield estimate than the often-pleaded "internal consistency"; the case for strong primordial fields is often excellent, until closely examined. Thus, iron meteorites [12] can, and probably do, carry a paleoremanence, and TRM vs. NRM comparisons can be obtained, with impressive "internal consistency," from comparing the NRM with zero-field moments. If similar field strengths are obtained for a group of meteorites by diverse experimental methods (e.g., $\sim 10^6$ oe for carbonaceous chondrites [4-7]), these can be considered trustworthy, divergent interpretations notwithstanding. Zero-field cooling offers a promising new standard for judging the suitability for samples for paleointensity estimates, since it permits one to separate internal vs. external-field contributions to the overall magnetization.

ANCIENT MAGNETIC FIELDS ESTIMATES FROM METEORITES

Brecher, A.

From meteorite data at hand, a plausible -- but not airtight -- case can be made that strong, extended nebular fields were present during the cold condensation-aggregation stages of meteorites (carbonaceous chondrites) and during the collisional-shock modification of ureilites. In contrast, fields lower by one to two orders of magnitude are obtained from (shock)-metamorphosed ordinary chondrites. This may pose a temporal constraint on field decay over the metamorphism interval ($\sim 10^7$ years).

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ON THE OBSERVED RELATION BETWEEN MAGNETIC FIELDS AND ROTATION IN ASTRONOMICAL OBJECTS; A. Brecher, Department of Physics, Wellesley College, Wellesley, Mass., 02181 and K. Brecher, Department of Physics, M.I.T., Cambridge, Mass., 02139.

It is believed that dynamo action resulting from fluid motions in the cores of rotating planets and stars can give rise to magnetic field generation. The quantitative dynamo theories presented to date are mainly kinematical, aiming primarily to show that magnetic fields can exist in rotating fluids, though some dynamical arguments do lead to scaling "laws" involving various physical parameters. Without delving into the dynamo problems directly, it is our purpose here only to re-examine a phenomenological relation which connects the angular momenta and magnetic dipole moments of planets (as well as for stars and galaxies) which we believe should be recoverable from any theory of the origin of magnetism in astronomical bodies which claims to be correct.

That the magnetic field of the Earth is caused by its rotation was suggested at least as early as 1891 by Sir Arthur Schuster and, independently, by Lord Kelvin. Schuster (1) and later P.M.S. Blackett (2) pointed out that a simple relationship exists for astronomical bodies between their magnetic moment $\mu = \alpha BR^3$ and angular momentum $J = \beta MR^2\omega$ of the form

$$\mu = \gamma G^{1/2} c^{-1} J \quad [1]$$

where M , R , B and ω are the mass, radius, average surface magnetic field strength and angular velocity of the object, and α , β and γ are constants of order unity. The presence of Newton's constant G in "Schuster's hypothesis" suggested to Blackett a possible deep new law of physics. The universality of such a new fundamental law was not, however, borne out by subsequent laboratory studies of rigidly rotating conductors or by observations of the radial variation of the Earth's magnetic field (3).

What do measurements of planetary magnetic fields reveal about Schuster's relation? In figure 1 we have plotted the best available data on magnetic moments versus angular momentum for several planets, the Moon and Sun. Many of the values of B are quite uncertain, but are probably correct to within better than an order of magnitude; no error bars have been assigned in the figure since each would be extremely subjective. The solid line has slope $dJ/d\mu = 1$. As it can be seen, (Fig. 1) there is a general trend extending over ten orders of magnitude suggesting a linear increase of μ with J . Is this result meaningful? Most recently, in a review of the same data, Russell (4) finds a similar relation, but concludes that it is not significant. First, he suggests, because there is no physical justification for the result. However, a similar argument could have been made about Kepler's laws of planetary motion before Newton showed how to derive them from an inverse square law force, or about Kirchhoff's observation that $\epsilon_0 \mu_0 c^2 = 1$ before Maxwell's equations were formulated. Since Schuster's relation does not

ON THE OBSERVED RELATION ...

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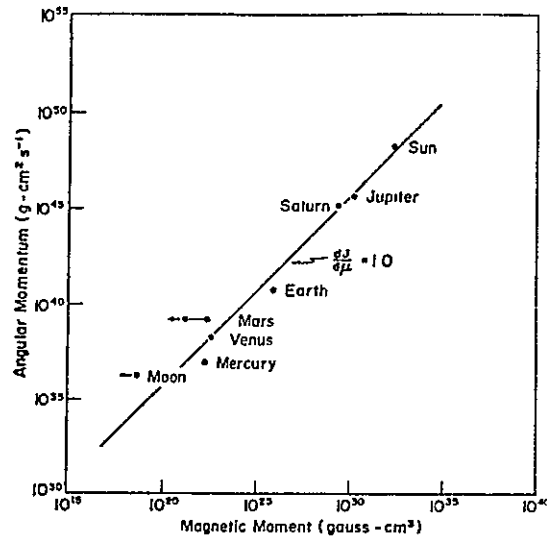
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FIGURE 1

obviously violate any known law of physics, we do not take the absence of a theoretical basis for the phenomenological relation to be an argument against its validity. Second, Russell argues that plotting any two planetary extensive variables against one another would produce an equally appealing result. However, plotting, for example, the magnetic energy E_m versus the rotational energy E_r for each object yields no such simple scaling which is independent of planetary properties. (In fact, for the many astronomical objects which rotate near their breakup speed so that $\omega^2 \simeq G\rho$, one finds using relation (1) that E_m/E_r varies roughly as $(\omega R/c)^2$.) Finally, it is worth noting that if one extends the present result to stars and the Galaxy (as has also been done by Warwick (5)), one finds a similar result (figures 2 and 3) extending over some 40 orders of magnitude in J and μ !

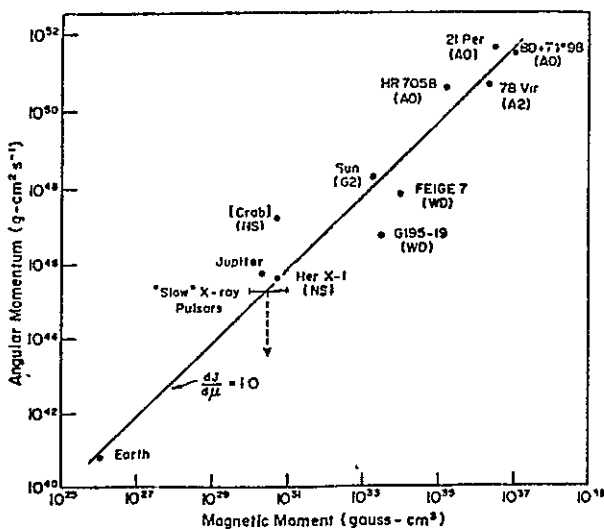


FIGURE 2

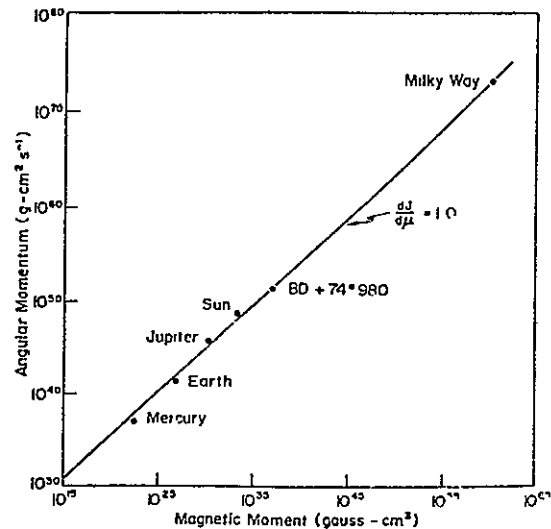


FIGURE 3

ON THE OBSERVED RELATION ...

Brecher, A. and Brecher, K.

What is the significance of Schuster's relation? The Moon, Mercury, Neutron Stars and White Dwarfs do not fit very well; all these are objects in which current dynamo action is probably non-existent. It seems reasonable that the plot is telling us something about objects in which dynamo action is occurring over a significant fraction of their volume, so that the gross macroscopic parameters considered here reflect well some average over their internal "microscopic" fluid motions, heat flows, etc. We think that the Schuster relation bears the same connection to dynamo theory that thermodynamic relations bear to statistical mechanics. We may never know anything certain about the detailed internal features of stars and planets and, therefore, may never fully determine what must go into a detailed dynamo model. However, any dynamo model giving rise to Schuster's relation will probably be a good bet for the "correct" solution to the origin of cosmic magnetic fields.

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CONSTRAINTS ON THE COMPOSITIONS OF THE CORES OF THE TERRESTRIAL PLANETS

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Knowledge of the composition of planetary cores is important: 1) to establish bulk compositions of planets and thus test condensation and accretionary models, 2) to understand thermal regime and history, 3) test model for core formation, 4) understand planetary dynamos. The existence and size of a core can be established by data including bulk density of the planet, coefficient of moment of inertia, existence of a dipole magnetic field, and seismic observations. The composition of a core may be constrained by density considerations from seismic studies, geochemical and cosmochemical abundances, and in the future, by electrical conductivity measurements.

There appears to be no relation between the size of the core and distance from the sun or size of the planet. The earth is the only planet for which we have seismic data that clearly indicate a metallic core. Density considerations indicate that the bulk of the core is metallic iron, and the molten outer core has a density 8-20 percent less than that of Fe. Cosmochemical considerations suggest some Ni in the inner core. Several suggestions have been made for the light element in the outer core; S and O are presently the most popular candidates. No quantitative geochemical arguments exist proving the presence of either element in the outer core, however if the earth accreted with appreciable sulfur, it would be unlikely if the bulk of this did not enter the core. Potassium has also been postulated to enter the core in substantial quantities, thus providing a source of heat deep within the earth, but the hypothesis must remain speculative until the correct critical experiments have been done.

Estimates of the compositions of cores of the terrestrial planets are closely related to our knowledge of condensation, accretion and coreforming processes. If the accretion was a heterogeneous process, with the core forming first, then it is difficult to balance the thermal budget to get a molten core early in planetary history, and it is difficult to understand how appreciable quantities of light element entered the earth's core. Homogeneous accretion with subsequent differentiation to form a core provides a ready heat source since Flasar and Birch have shown the temperature of the earth is raised by 2300 K by such a process. Such a process of core formation readily provides a light element, in fact the presence of a light element alloying with iron considerably lowers the temperature of core formation.

Models exist (e.g., those of J. L. Lewis) to explain the compositions of planets including their cores in terms of condensation theory. For example, Mercury is proposed to have accreted from high temperature material and thus is low in volatiles and has a Fe-Ni core. No meteorites (apart from some rare Allende inclusions), including the refractory enstatite chondrites are without appreciable S and other volatiles, nor is the earth or refractory moon. This therefore suggests that the accretion of planetary and asteroidal bodies involved both high and low temperature condensates in varying proportions. The cosmochemical models therefore represent a very necessary beginning but may be somewhat unsophisticated. Geophysicists concerned with thermal history calculations for planets should therefore not be constrained at present by the cosmochemists, and all core compositions and sizes consistent with the other constraints should be considered.

GENERATION OF PLANETARY MAGNETISM BY CONVECTION, by F. H. Busse, Institute of Geophysics and Planetary Physics, UCLA, Los Angeles, California, 90024.

Thermal convection is the predominant source of motions in the metallic hydrogen cores of Jupiter and Saturn. Convection driven by thermal and/or chemical buoyancy is likely to occur in the Earth's core and possibly in the core of Mercury. Thus a dynamo driven by convection in rotating spherical shells represents a suitable model for a general theory of planetary magnetism. While a numerical analysis of this problem is in progress, the experience in the non-magnetic case has shown that the physical properties of the problem can be best understood by considering the simpler geometry of a rotating annulus. The annulus model of planetary dynamos (Busse, 1975, 1976) has recently been extended to include in higher degrees of approximation the effects of the Lorentz force. In particular the reinforcement of dynamo action, the nonlinear oscillation of the magnetic energy about its equilibrium value, and the effect on the convection of the Taylor-shear generated by the deviations from the Taylor condition (Taylor, 1963) will be discussed.

Because the toroidal field of the annulus model is roughly of the same order of magnitude as the poloidal field the model can be used for a comparison of the theoretically inferred field strength with the observed values of the planetary magnetic moments. This comparison is based on the hypothetical upper bound on the magnetic field strength for which convection in the form of thermal Rossby waves can exist. In addition an assumption about the maximum scale of motion derived from numerical solutions of kinematic dynamos in a sphere (Gubbins, 1973) enters into the comparison formula. This and other proposed scaling laws for planetary moments will be briefly discussed in relation to the observations.

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GLOBAL GEOMAGNETIC ANOMALY STRUCTURE. Joseph C. Cain, U.S.
Geological Survey, Denver, Colorado 80225.

The results obtained by Russian and U.S. satellites have been reviewed for geomagnetic field components of spherical harmonic degree and order above eight. Resolution is limited to scale sizes of 500 km. The spatial spectra of the average power and energy for each degree harmonic implies that a transition from core to crustal sources occurs between $n = 8$ and $n = 12$. Anomalies that have wavelengths in this range are not uniformly distributed and do not appear to be related to known tectonic elements. Their presence could arise either from small-scale eddies on the core surface or, in some instances, from a spatial modulation of secular change by lateral conductivity contrasts in the mantle. Anomalies corresponding to $n > 12$ (wavelengths below about 3000 km) appear to be more related to crustal structure, although few have yet been investigated adequately to make detailed interpretations.

ON THE POSSIBILITY OF USING LUNAR FINES TO DETERMINE THE INTENSITY OF THE ANCIENT MAGNETIC FIELD OF THE MOON; D. W. Collinson, Institute of Lunar & Planetary Sciences, University of Newcastle upon Tyne, England.

1. Introduction

Reliable determinations of the intensity of the ancient lunar magnetic field are difficult experimentally partly because of restrictions imposed by the lunar sample research programme. Among these are the small sample mass available for investigation, which sometimes leads to low or very low total moment for measurement, lack of samples on which to conduct concurrent tests for mineral alteration during heating, and the availability of samples whose magnetic properties suggest successful palaeointensity investigations.

By imparting a depositional remanence (DRM) to lunar fines it has been demonstrated that they contain particles which possess an NRM of high stability (Runcorn et al 1970), and it is of some interest whether this NRM can be utilised for palaeointensity determinations and whether any field value so obtained can be interpreted in a useful manner. Lunar fines have rather strong magnetic properties and afford the possibility of comparatively strong intensities of magnetization with which to work and the availability of ample material for the testing of thermally-induced changes.

2. Method

The magnetized particles in the fines, normally randomly directed, are aligned by allowing them to settle through a liquid into a shallow quartz dish in an ambient magnetic field. After drying out the fines possess a strong DRM, the magnetic properties of which can be studied in the usual way. A Thellier-Thellier palaeointensity determination can then be carried out by comparing the loss of DRM(NRM) with the gain of PTRM in successively higher temperature intervals.

3. Results

Two independent Thellier determinations were carried out on < 1 mm fines, sample 75081, from Camelot crater at the Apollo 17 site. Mineralogical changes were monitored by measuring low field IRM on a separate sample after each heating. There was evidence of changes taking place above $\sim 500^{\circ}\text{C}$, and measurements were then discontinued. Fig. 1 shows the results obtained from the two samples, giving apparent ancient field intensities of ~ 0.9 Oe and 1.2 Oe.

4. Discussion

The interpretation of the above result hinges on two main points, (a) the origin of the iron particles in the soil and the nature of their NRM, and (b) the validity of the DRM method.

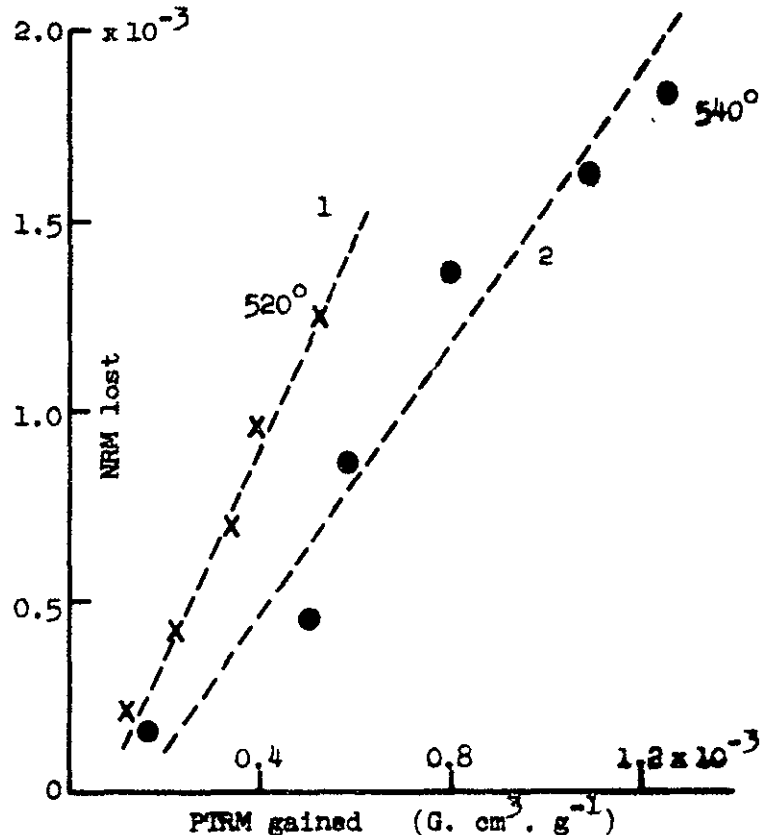
(a) It is well-known that lunar soils are not simply an erosion product of igneous rocks and breccias, and among the important compositional differences is the excess iron metal (Fe^0) in soils relative to that in igneous rocks, by a factor of 5-10 (Housley et al 1970). In addition, therefore, to a basaltic Fe^0 contribution in soils there is other iron metal present the two main sources of which are meteoritic and Fe^0 contained in agglutinates.

Palaeointensities from lunar fines

Collinson, D.W.

Fig. 1

Plots of NRM lost against PTRM gained for two samples of 75081 fines. The applied fields were 0.34 Oe and 0.67 Oe for runs 1 and 2 respectively. The points correspond to approximately 100°C temperature intervals.



Agglutinates are the glass-welded aggregates of rock, mineral and glassy fragments which occur in lunar soils. They are believed to be formed by micro-meteorite or solar wind particle impacts into the regolith (Housley et al 1973, Gardiner et al 1977) and they contain fine-grained Fe^0 in the form of droplets on and inclusions in the glass. The metal is believed to be precipitated out of the silicates, either by the action of reducing gases trapped in vesicles at the time of impact (Housley et al 1973), by subsolidus reduction at elevated temperatures ($\sim 700^\circ - 900^\circ C$) in, for instance, an ejecta blanket (Pearce et al 1972), or by a preferential sputtering process (Gardiner et al 1977).

Thus, the iron metal in lunar soils can occur in particles of basaltic rock, in glass-welded aggregates in which grainsizes spanning the superparamagnetic to multidomain range occur, and as meteoritic iron particles. The likely origin of any NRM possessed by the different types of iron is difficult to assess, and also the contribution of each type of iron to the DRM of the fines. The simplest hypothesis is that the rock fragments and meteoritic iron possess their original NRM, and that the agglutinate iron has also acquired a remanence, either a CRM or TRM, at the time of its formation. However, in view of the meteoritic bombardment history of the regolith, the effect of shock on the NRM of all iron particles is likely to be an important factor, potentially more important than may be the case with rock samples (Cisowski et al 1973).

(b) The assumption that the DRM acquired by a fines sample can be used as the starting point for the Thellier technique appears to be a valid one providing certain conditions are met. The DRM should be at or near saturation, i.e. the particle magnetizations should be approximately parallel, and it is necessary that there should not be a substantial contribution to the DRM from isolated single-domain grains, since they cannot individually record the

Palaeointensities from lunar fines

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ancient field intensity when they become magnetized. The PTRM should be imparted parallel to the DRM direction in the sample.

5. The result from 75081,15

The origin of the iron metal in 75081 fines is not clear. Goldstein et al (1974) report that the composition of the most magnetic particles from this soil is consistent with a mainly basaltic origin for the Fe^0 , with $\sim 20\%$ of the Fe^0 content of meteoritic origin. However, an approximately equal mixture of meteoritic and solar wind reduced Fe^0 is suggested by Chou and Pearce (1976).

The saturated DRM intensity of the two samples was $\sim 2.0 \times 10^{-3}$ emu/g. This strong magnetization suggests that much of the iron metal in the fines is magnetized and contributing to the DRM: with $\sim 0.5\%$ by weight of Fe^0 in the fines, the average intensity of magnetization of the iron is high, about 0.4 emu/g, which, because of the presence of SPM iron, is a minimum value. It also seems likely that some of the agglutinate iron may not be contributing to the DRM, because the combined moment of iron and glass is too weak for good alignment to occur during DRM acquisition.

The results obtained with the 75081 fines suggest that a palaeointensity of ~ 1.0 Oe has been determined, but it must be admitted that without further investigation the result is difficult to interpret and it is not possible to assign the value to any age on the Moon. The thermal and a.f. demagnetization curves of the DRM do not suggest more than one NRM component, and the NRM lost/PTRM gained plot might be expected to show evidence of two magnetic components of different blocking temperature spectra, if they were present. There are reasons for believing that 1.0 Oe may be an underestimate of the field value, for instance because of the possible presence of unmagnetized Fe^0 .

If the question of the nature and origin of the ancient lunar magnetic field is to be clarified, it will be necessary to make maximum use of all available lunar material. These preliminary results suggest that the fines may be of use in elucidating the problems of the lunar magnetic field, and they offer the further advantage of economy of sample required (~ 100 mg per determination) and strong, easily measurable magnetizations. At the present stage of lunar palaeointensity research the question of age is not of overriding importance, since a well-defined, reliable palaeointensity at any stage in the Moon's history is of considerable interest.

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ORIGIN OF LUNAR MAGNETISM BY SURFACE PROCESSES

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The first lunar samples returned by Apollo astronauts eight years ago were found to have a thermoremanent type magnetization of 10^{-7} to 10^{-3} emu/g acquired in fields from 10^{-2} to 1 Oe. Subsequent in situ magnetometer measurements reveal that much of the lunar crust is uniformly magnetized over scales of 10 to 100 km. Many hypotheses have been proposed to explain the magnetic remanence but no single theory has been entirely satisfactory. We review here, and schematically depict in the figure following this review, the principal hypotheses which do not involve global effects but which require localized or near surface processes.

(a) Shallow Fe-FeS Dynamo. Pearce et al. (1972) suggested magnetohydrodynamic dynamo action in pockets of molten metal just beneath the lunar surface to produce the fields necessary to magnetize the cooling lunar crust. Murthey and Banerjee (1973) have shown that if these dynamos are Fe-FeS eutectic liquid $\sim 10^{-4}$ lunar mass they may be capable of a regenerative dynamo action. Such dynamos may have a symmetry with the lunar spin axis capable of explaining the observed preferential magnetization alignment perpendicular to the spin axis. Their variety of lifetimes and sizes could also explain the four decade range in palointensity values measured for samples which span nearly 10^9 years in age. However, it is doubtful that these dynamos would produce the fields up to 1 Oe. which have magnetized some lunar samples. In addition, these Fe-FeS concentrations (fescons) are expected to be preferentially located beneath maria. Therefore, the maria should be more highly magnetized than highlands, and this is contrary to results of surface and orbital measurements.

(b) Local Unipolar Dynamo. Enhancement of the solar wind field is possible by induction currents in highly conducting lava pools on the lunar surface. As the solar plasma and magnetic field $\underline{B}$ are convected past the moon with velocity $\underline{V}$, a motional electric field $\underline{E} = -\underline{V} \times \underline{B}$ drives currents through the lava which are closed in the solar wind. Nagata et al. (1971) estimate enhanced fields of 10^{-2} Oe. at the surface for pristine solar wind conditions. This field may account for remanence in some samples, but it should be noted that fields generated by this process vary with the solar wind field and are even absent at night and in the magnetotail when the plasma return path is not present.

(c) Comet Impact Magnetization. Collisions of comets with the moon may magnetize the crust in a two stage process proposed by Gold and Soter (1976). As the comet nucleus approaches, its coma engulfs the moon and the ambient solar wind field is compressed to about 1 Oe. between the partially ionized coma streaming toward the surface and the coma ions reflected from the surface. Coincident with the field enhancement, the comet nucleus impacts, and the second stage is shock remanent magnetization in the high field region. While this mechanism explains many properties of lunar remanence, the observed lack of correlation between craters and magnetic anomalies is difficult to explain.

(d) Thermoelectric Dynamo. Generation of electrical currents and associated magnetic fields by the Seebeck effect in the early cooling crust has been suggested by Dyal et al. (1977) as a mechanism for crustal magnetism. In this model lava basins are electrically connected below the solid crust by magma and above by the solar plasma. Differential cooling results in large temperature differences between the basins which act as junctions in a thermoelectric circuit and for probable electrical parameters of magma and plasma, calculated

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fields are as large as 10^{-1} Oe. in the cooling crust between the basins. In this mechanism the thermoremanent magnetization and field generation processes are both naturally connected to cooling of large regions during crustal formation. However, many of the model parameters such as the relative Seebeck coefficient of plasma and magma are unknown and require empirical determination.

(e) Impact Induced Piezomagnetism. The most studied surface process for explaining crustal remanence is subsurface shock loading during crater formation. Cisowski et al. (1973) have measured shock induced remanence of 10^{-3} emu/g in lunar soil samples shocked to 50 k bar in the earth's field. Assuming linear dependence of induced remanence with field, a magnetization of 10^{-6} emu/g (comparable to that of lunar samples) may be expected for possible lunar fields of 10^{-4} Oe. Hypervelocity ejecta from explosive charges have also been used by Martelli and Newton (1977) to show that shock produced cratering in a basalt target results in a transient magnetic field during crater formation and an induced remanence in the target material. These results, as well as those of other experiments, are encouraging since this mechanism explains the crustal remanence as a natural consequence of an acknowledged lunar process. The mechanism also explains the observed larger remanent fields in heavily cratered highlands even though there is no observed correlation between individual craters and magnetic anomalies.

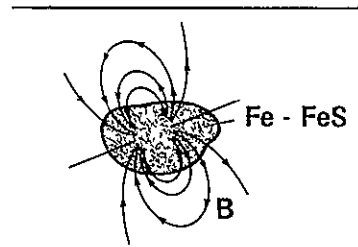
(f) Impact Ejecta Field. Magnetic fields generated in the ionized ejecta of hypervelocity lunar impacts were first postulated by Hide (1972). Srnka (1977) calculated model dependent estimates of magnetic fields in crater ejecta of 10^{-3} to 10^{-1} Oe. lasting for tens of seconds. Laser-target plasma experiments have produced measured fields up to 10^5 Oe. Piezo- or thermoremanence may be acquired by ejecta and nearby rock from these large fields. This mechanism explains the small present global magnetic moment as well as the larger remanent fields measured over heavily cratered highlands and smaller fields over sparsely cratered maria. However, there is little evidence that magnetic sources are correlated with individual craters.

(g) Volcanic Ash Flow Magnetic Fields. Several schemes for generation of magnetic fields by surface processes were investigated by Cap (1972). In the most plausible model fields up to 10^{-2} Oe. were calculated from convection currents in ionized volcanic ash flows. The model generates highly transient fields, however thermoremanent magnetization requires a relatively constant field as the crustal material cools from the iron Curie temperature.

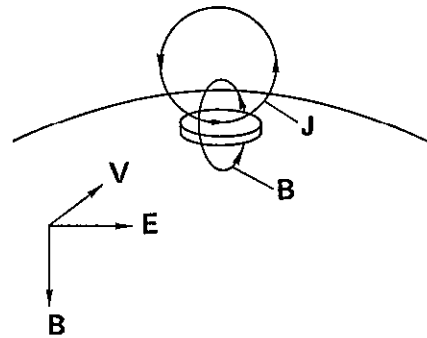
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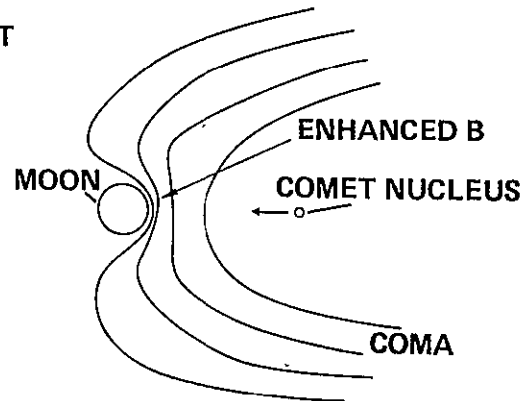
SHALLOW Fe - FeS DYNAMO



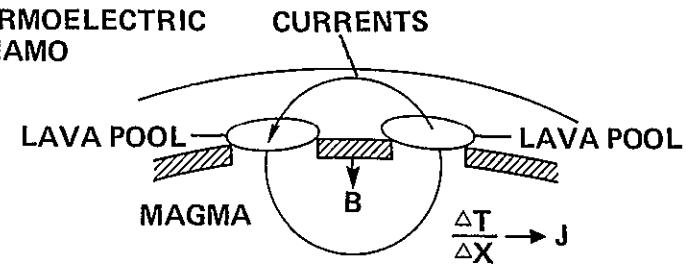
LOCAL UNIPOLAR DYNAMO



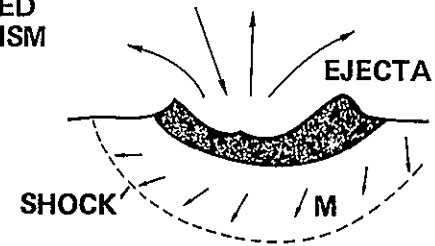
COMET IMPACT



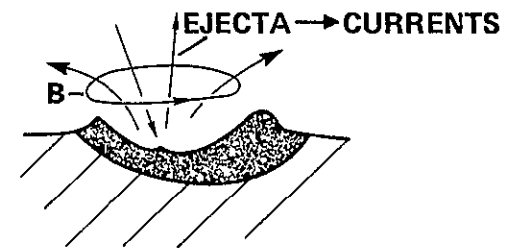
THERMOELECTRIC DYNAMO



IMPACT INDUCED PIEZOMAGNETISM



IMPACT EJECTA



VOLCANIC ASH FLOW

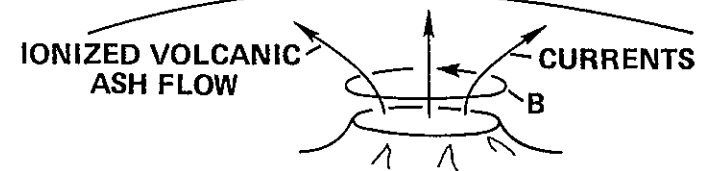


Figure. Schematic representation of surface processes proposed to explain lunar magnetism.

LUNAR MAGNETISM BY SURFACE PROCESSES
Daily, W. D. and Dyal, P.

HYDROMAGNETIC INTERACTIONS IN A FLUID DIELECTRIC

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The subject of magnetohydrodynamics deals with the dynamic behavior of a conducting fluid permeated by a magnetic field – and forms the basis of the planetary dynamo theory [1]. Here we discuss a parallel development dealing with the dynamic behavior of a dielectric fluid having a finite molecular polarizability, when permeated by a magnetic field. Whether or not this development has implications for the possibly fluid dielectric interiors and exteriors of giant planets is not clear at this time.

The basic equations of magnetohydroelectrics (so termed because of the importance of the electrostatic energy which is negligible in magnetohydrodynamics) are first the four Maxwell's equations (in S.I. units):

$$\nabla \cdot \mathbf{B} = 0 \quad (1)$$

$$\nabla \cdot \chi \epsilon_0 \mathbf{E} = - \rho_p \quad (2)$$

$$\nabla \times \mathbf{B} = \mu \mathbf{j} + \mu \epsilon_0 \dot{\mathbf{E}} \quad (3)$$

$$\nabla \times \mathbf{E} = - \dot{\mathbf{B}} \quad (4)$$

where $\mathbf{E}$ is the electric field in the laboratory frame, $\mathbf{B}$ is the net magnetic field, $\mathbf{j}$ is the polarization current, ρ_p is the polarization charge density, and χ , ϵ_0 , and μ are respectively the molecular polarizability*, the permittivity of free space, and the magnetic permeability of the fluid. Note that Eq. (2) – a modified form of Maxwell's equation – is written in terms of the polarization charge since this is the only charge that can arise in the body of the moving fluid, any free charges arising in a bounding conductor. The dielectric permittivity of the fluid is

$$\epsilon = (1 + \chi) \epsilon_0 \quad (5)$$

and is assumed here to be a scalar quantity. To these equations we add the Newton's law

$$\rho \dot{\mathbf{u}} = \mathbf{j} \times \mathbf{B} \quad (6)$$

(ρ = fluid mass density, $\mathbf{u}$ = fluid bulk velocity), and the equation relating the polarization current to the electric field [2]

$$\mathbf{j} = \chi \epsilon_0 \left[\dot{\mathbf{E}} + \frac{\partial}{\partial t} (\mathbf{u} \times \mathbf{B}) \right] \quad (7)$$

*Referred to the bulk fluid

HYDROMAGNETICS OF FLUID DIELECTRICS

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The set of equations (1) - (7) completely describe the ideal magnetohydroelectric behavior. Note that there is no energy dissipation involved in these equations.

Upon eliminating $\underline{J}$ and $\underline{E}$ from Eq.(7) by means of Eqs.(3) and (4), and using Eq.(1), we have

$$\ddot{\underline{B}} = \frac{\chi\epsilon_0}{\epsilon} \frac{\partial}{\partial t} \nabla \times (\underline{u} \times \underline{B}) + \frac{1}{\mu\epsilon} \nabla^2 \underline{B} \quad (8)$$

In this equation the two terms on the right hand side have magnitudes:

$$\frac{\chi\epsilon_0}{\epsilon} \frac{\partial}{\partial t} \nabla \times (\underline{u} \times \underline{B}) \sim \frac{\chi\epsilon_0}{\epsilon} \frac{u B}{\tau L}$$

$$\frac{1}{\mu\epsilon} \nabla^2 \underline{B} \sim \frac{B}{\mu\epsilon L^2}$$

where L and τ are the characteristic length and time scales of the problem. The ratio of the first of these terms to the second is

$$\chi = u L \mu \chi\epsilon_0 / \tau \quad (9)$$

If this dimensionless parameter χ is much greater than unity, then Eq. (8) reduces to

$$\ddot{\underline{B}} = \frac{\chi}{1+\chi} \frac{\partial}{\partial t} \nabla \times (\underline{u} \times \underline{B}) \quad (10)$$

Integrating once with respect to time, we have

$$\dot{\underline{B}} = \frac{\chi}{1+\chi} \nabla \times (\underline{u} \times \underline{B}) + \underline{C} \quad (11)$$

where the integration constant $\underline{C}$ must be set equal to zero, since otherwise it indicates a monotonic change in $\underline{B}$ with time even when $\underline{u} = 0$ everywhere. Under the condition $\chi \gg 1$, Eq.(11) reduces to

$$\dot{\underline{B}} = \nabla \times (\underline{u} \times \underline{B}) \quad (12)$$

which is the familiar condition for a state of "frozen flow" to exist in a magnetized fluid: the magnetic flux through any contour attached to the fluid stays constant with time.

If on the other hand $\chi \ll 1$, then Eq.(8) reduces to

$$\ddot{\underline{B}} = c_d^2 \nabla^2 \underline{B} \quad (13)$$

where $c_d = (\mu\epsilon)^{-1/2}$ is the velocity of electromagnetic waves in the dielectric. This is a three dimensional wave equation - indicating that the disturbance is now purely electromagnetic in nature.

We thus find that the magnetohydroelectric interactions are in principle very similar to the magnetohydrodynamic interactions, except that the former interactions are energy-conservative so that a situation with a diffusion of the magnetic field lines does not arise (as does in the latter case).

HYDROMAGNETICS OF FLUID DIELECTRICS

B. R. DE

It has been shown elsewhere[2] that analogous to the Alfvén waves of magnetohydrodynamics, magnetohydroelectric waves – consisting of oscillations of magnetic, electrostatic and kinetic energies – are theoretically possible. Such waves propagate with a velocity

$$V = c \frac{[X^{-1} + v_a^2/c^2]^{1/2}}{[1 + X^{-1} + v_a^2/c^2]^{1/2}} \quad (14)$$

where c is the velocity of light, and $v_a = B/(\mu\rho)^{1/2}$. Note that v_a would be the Alfvén velocity in the medium had the medium been conducting. The effect of introducing a finite conductivity in the fluid has also been discussed in the above reference.

It can be shown that the induced magnetic energy, the electrostatic energy, and the kinetic energy can reach approximate equipartition when $V \sim v_a \sim c_d \ll c$. However, even when this condition does not prevail, magnetohydroelectric interaction may still be of interest. In the possibly superdense core fluid of giant planets, X may be quite large (since $X \propto$ number density of the molecules). On the other hand, the velocity u of the fluid may be quite small and τ large, making X smaller than unity. Thus the importance of magnetohydroelectric effects in the planetary dynamo theory is questionable, but may possibly be worth bearing in mind.

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MAGNETIC FIELDS IN THE VICINITY OF VENUS
 ACCORDING TO "VENERA" AND "MARINER" DATA

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The nightside Venus cavity is the magnetosphere in which the magnetic energy exceeds that of the solar wind, and the field regularity distinguishes the magnetosphere field from the magnetic field within the magnetosheath.

At distances of $2.5 \pm 6 R_V$ from the planet center the nightside magnetosphere topology is characterized by stretching out of lines of force in radial direction; it has two lobes of magnetic lines of force: "from the planet" one (northward of equator), and "toward the planet" one (southward of equator), which are divided by a layer with the magnetic energy of deep minimum, which is quite analogous to the earth magnetosphere neutral layer [1].

During four of measurement seances from "Venera-10" there were complete magnetosphere crossings at distances of $\sim 10000 \pm 25000$ km. During other seances a boundary region was revealed only with $Y_{SE} < 0$. During 10 seances of measurements from "Venera-9,10" a neutral layer was observed when the ecliptic plane was being crossed at distances of $X_{SE} = 8000 \pm 23000$ km.

At closer distances from the planet the magnetotail is becoming narrow. At distances of $\sim 6 R_V$ the magnetosphere diameter exceeds that of the geometrical umbra, and close to the planet the magnetotail is inside the region of the planet's geometrical umbra. The magnetosphere sizes depend on conditions both in the magnetosheath zone and solar wind.

Assuming a circular shape of the magnetosphere cross-section it has been estimated that the magnetosphere radius at a distance of $X_{SE} = -34000$ km is ~ 7000 km, and at a distance of $X_{SE} = -10000$ km it is ~ 4500 km (Fig.1).

The intensity of field B_x radial component in the tail is of about $17 \pm 15 \mu$ without occurrence of its diminution moving away from the planet which should have been expected due to a growth of magnetosphere sizes with the growth of distance from the planet. This is probably connected with a non-conservation of flux constancy in the venerian magnetosphere due to a permanent course of "reconnection" processes of the magnetosheath field with the fields of either north or south lobe of the magnetosphere field.

This "reconnection" is manifested by a diminution of the field B_x radial component intensity in the tail over the ecliptic with a solar-pointed field B_x in the transition zone, while observed under the ecliptic - with a antisolar field B_x . In these cases, the field B_x is decreasing to zero rather quickly, and then a field B_x occurs directed either from the planet

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(over the ecliptic) or "toward the planet" (under the ecliptic).

The best way to explain the totality of data is to assume that this field is the planet intrinsic field blown off to the night side and amplified by the fields of surface currents generated by solar wind [2], as it takes place in the earth magnetosphere.

The above-mentioned magnetosphere boundaries at close distances have been determined by the field B_x abrupt diminution commencements.

The magnetosphere field characteristic feature is a small value of "southward" field B_z , although sometimes a weak field component of inverse sign has been revealed which is apparently due to the external field activity.

A complicated topology of the field at distances of $\sim 1500+2500$ km can be most clearly explained by the model of the planet intrinsic field interaction with the external field of solar origin. The intrinsic magnetosphere field is seen when the orbiter penetrates inside the geometric umbra up to $Z_{SE} < 3000$ km with $\theta = (Y_{SE}^2 + Z_{SE}^2)^{1/2} < 4500$ km.

With greater values of Z_{SE} some solar-origin fields are observed, B_x component of which coincides by sign with the field B_x in the transition zone. Sometimes in such cases there is also observed a considerable component B_z contrary to the magnetosphere field in which B_z is small.

At greater distances ($3+5 R_V$) the tail topology also depends on the field intensity in the magnetosheath zone and retroactive. The existence of an inner magnetotail-like wake structure was supposedly suggested on the base of the spatial variations observed by "Mariner-10" [6] in the distant magnetosheaths up to $750 R_V$ behind Venus. According to [13] these variations were probably due to temporal changes.

Nearly in all measurements performed at distances of $1500+2500$ km the component B_y is directed toward $(-Y_{SE})$. There has been discussed a probable cause of this by a deviation error occurring during the satellite work on the "pericentre" programme as well as a probable effect of currents flowing along the lines of force in the tail.

Contrary to [3] it has been concluded that a probable effect in the day-time ionosphere of a unipolar induction mechanism from B_1 component is not clearly manifested in the night-side tail topology. At the same time, during some seances meridian crossings small effects have been observed.

$$\text{Equation } H_T = 4H_0 \frac{R_V^2}{R_T^2} \sin^2 \theta$$

which application is limited in case of venusian magnetosphere, permits to infer that the tail might be formed with $H_0 = 10 \mu$ and $\theta = 31^\circ$ (that is, with $\theta_V = \theta_m$ of Mercury) as well as with $H_0 = 5 \mu$ and $\theta = 46^\circ$ which corresponds to values $M_V = (1, 2+2, 5) 10^{22} \text{ Gs} \cdot \text{cm}^3$.

In any case, this field is not an obstacle to solar wind in according with the conclusions on the obstacle being of ionospheric origin made on previously performed experiments [4,5,6].

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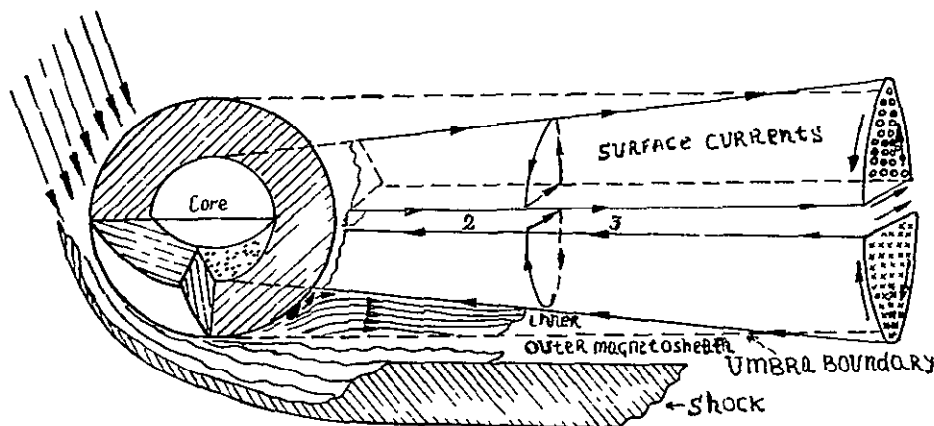
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and conclusions [7,8].

Initial assumptions and model, from which it had been concluded [9] that Venus had a field with the magnetic moment $M = 6,5 \cdot 10^{22} \text{ Gs} \cdot \text{cm}^3$, contradict the "Venera-9,10" experimental data, although there is some agreement between the suggested and observed field signs, that is, the venusian field is opposite to the earth field by sign.

The model of Venus magnetosphere shown on Fig.1 explains the observed [10,11] peculiarities of plasma distribution.

The observed field being very small, its origin can be most naturally explained by the operation of a dynamo-mechanism in the venusian liquid core. The Venus field appears to be similar to the earth field if to take into consideration the peculiarities of its rotation and precessions [12].



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MAGNETIC FIELDS OF PLANETS AND PECULIARITIES OF THEIR ROTATION

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It has not justified the supposition of the planetary magnetic fields proportionality to the velocities of their angular rotation which appeared to be natural [1]. At the same time it has been revealed the close correlation between the changes of the Earth's rotation velocity (the accelerations) and the changes of the geomagnetic field parameters [2,3].

The permanent acceleration of the Earth that is associated to the planetary precession

$$W = [\omega \times \Omega] = 2,25 \cdot 10^{-10} \text{ rad/sec}^2 \text{ (the Poincare acceleration [4])}$$

is $10^3 \cdot 10^6$ times as much as the others known accelerations of the Earth [5,6]. Naturally, that the possible role of the Poincaré acceleration in the planetary magnetic fields generation

4 should be studied regardless of the problem of the precession efficiency as the source of the dynamo-process energy [7]. This work as well as previously published ones [8,9] is related to that problem.

The following experimental facts and physical principles were initial to the establishment of the analytical dependence $H = H([\omega \times \Omega])$

1. The scaling of some parameters of the planetary magnetic fields [8,9].

2. The principle of scaling law that is read: if two physical phenomena are similar the characteristics one of them may be obtained according to the others by means of the simple recalculation that is similar of the measurement units transformation [10].

3. The induction equation

$$\frac{\partial H}{\partial t} = \eta \nabla^2 H + \text{rot} [V \times H] \quad (1)$$

where the first term is proportional to $\tilde{\gamma} = R_c^2 / \eta$ - (R_c and η are size and magnetic viscosity in the fluid core) and the second term is supposed to be proportional to $k \cdot F_p$ ($F_p = \rho [\omega \times \Omega] \times \bar{r}$ is the Poincaré force [4]), "k" is the factor that depends on the relation of the Poincaré force and the other ones acting in core, ρ - the core density, $\bar{r}$ - radius vector).

According to the principles mentioned above the planetary field is

$$H_i \sim \tilde{\gamma} k F_p \sim k \rho R_c^3 \omega \Omega \sin \alpha / \eta \quad (2)$$

where α is angle between $\bar{\Omega}$ and $\bar{\omega}$. In comparison to the geomagnetic field H_{oz} the planetary field will be "i"

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$$H_i = K H_{OE} \frac{Q_E^3}{Q_i^3} \frac{R_{ci}^3}{R_{CE}^3} \frac{\omega_i \Omega_i}{\omega_E \Omega_E} \frac{\sin d_i}{\sin d_E} \frac{\rho_i}{\rho_E} \frac{\eta_E}{\eta_i} \quad (3)$$

where $Q = R_n/R_c$ allows to compare the dipole planetary fields on the surface at equator. Under the perfect similarity and accuracy of the initial parameters $k = 1$.

There are the values of factor "k" in the table 1 that are calculated according formula (3) for the investigated planets and the Moon, the initial data accepted for the calculation and references.

Table 1

P.	H_0		Polarity	$f = 3/2 J_2$		α°	R_c/R_p	$\rho_g R_c m^{-3}$	$6 \cdot 10^5 m_o/m$	"k"
♀	170	[11]	S	$(1,2 \pm 0,9) 10^{-4}$	[15]	1±2	0,74	8,0	9	1,2
♀	5±10	[12]	N	$(6,0 \pm 2,0) 10^{-6}$	[16]	1±2	0,53	10,0	7	1,3
⊕	30950		S	$3,27 10^{-3}$	[17]	23,5	0,53	11,0	5	1,0
♂	60	[13]	N	$5,24 10^{-3}$	[17]	25,2	0,37	7,5	4	1,2
♂	419200	[14]	N	$6,66 10^{-2}$	[17]	3,0	0,70	1,9	7	0,4
☾	< 1		-	$3,07 10^{-4}$	[18]	1,5	0,20	7,3	5	0,8

The dispersion of the factor "k" is insignificant in spite of the immense difference of parameters and is in the limits of the accuracy of modern knowledge of H_0 , f , α , R_c and ρ . It indicates that at least the precession effects on the mechanism of the planetary magnetic field generation and in last one the external gravitation forces the important part play.

Note that the values of "f" for Mercury and Venus determined in [15] and [16] were predicted in [21,9] on base of formula (3).

The fact of a small dispersion of the factor "k" leads with the necessity to the equality

$$H_0 = K' \rho Q^3 R_c^3 \omega \Omega \sin d / \eta \quad (4)$$

where "k" - factor is the same for all the planets.

We are capable to determine the dimension of "k"

$$[K'] = C_m^{3/2} C_m^{-1/2} = \frac{1}{|\partial|} = \frac{1}{|\epsilon E|},$$

where $|\partial|$ - the dimension of the electric induction in CGSM units, ϵ - the permittivity.

$$\epsilon = c^2 v_{ho}^{-2} = v_{ho}^{-2} \text{ (in CGSM)}$$

where C - the velocity of light, v_{ho} - the velocity of the hydromagnetic oscillations.

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Then (4) can be expressed as

$$[HE] = \frac{\rho N Q^3 R_c^3 \omega \Omega \sin \delta V_{ho}^2}{\eta} \text{ erg/cm}^2 \text{ sec} \quad (5)$$

where N - a dimensionless coefficient.

Thus the physical sense of the formula (3) is that the planetary magnetic fields are proportional to the density of energy flux (a flux of the Poynting's vector) carrying by the hydromagnetic waves across the fluid core boundary.

It is necessary the new experimental data to precise the data of table 1. From formula (3) and modern models of the internal structure of Jupiter and Saturn 19 the intensity of Saturn's magnetic field should be $H_0 \approx 14000 \gamma$.

If it occurred that the polarity of Saturn's dipolar field is opposite of Jupiter's one then it should not consider as a chance phenomenon the observed alternation of the dipolar field signs of planets (see table 1) that is disturbed where the Bode's Law of the planetary distances is disturbed.

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DOES MARS HAVE AN ACTIVE DYNAMO?

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A negative answer to the question, mentioned in the title, is given by Russell [1] who argue it in the following way:

1. Mars spins at a rate very similar to the Earth and much faster than Mercury spins. It's size is intermediate between Mercury and the Earth. Thus, one might expect Mars to have a magnetic moment intermediate between that of the Earth and Mercury. However, the reported moment is much less than that of the Earth and that of Mercury.

2. The magnetic field measured near Mars [2] is induced in its ionosphere by the perpendicular to the solar wind flow B component of the interplanetary magnetic field. The version of Mars magnetic field origin suggested by Russell, allows him, in his opinion, to eliminate the established fact that observation magnetotail field sign is independent on the interplanetary magnetic field sign, which contradicts another version of Mars magnetic field origin, suggested earlier by Russell [3,4].

Reference the reader to [3-6] in which the reality of the martian intrinsic magnetic field is discussed, let us consider the validity of the new Russell's arguments, in favour of the induced origin of the martian magnetic field.

1. In accord with the unipolar induction theory [7] the electrical field of the solar wind $E = B_\varphi V_r = B_r V_\varphi$. Consequently if this mechanism is effective, the component B_r must be also effective. Further, it is known that the 3-hour average B_y and B_r components of the IMF are in 90-95% coincide in their sign change. The daily averaged B_z -component of IMF is small [8].

It was shown in [11] that a probable effect in the day-time venusian ionosphere of a unipolar induction mechanism from B_z -component is not clearly manifested in the nightside tail topology at the distances 1500+2500 km above the surface.

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The same may be observed even to a greater extent in the martian nightside field.

2. The magnetic fields of planets are not proportional to their constant spin rate but they seem to depend on any accelerations in the spin rate. As the Poincaré acceleration $W = (\bar{\omega} \times \Omega)$ connected with the precession is the largest one of the accelerations positively known on the Earth, it defines (besides the other parameters R_c, ρ, δ) the proportionality between the planetary magnetic field intensity and the value $W = \omega \Omega \sin \alpha$ but not the value ω [9].

The discovery of the magnetic field of the planet Mars which has the values ω and α similar to that of the Earth, but the values Ω different, allowed to reveal the mentioned dependence [10].

The magnetic field of Mars-5, being completely similar to the terrestrial magnetic field, can not be of large intensity.

. This probably indicates the magnetic field of Mars to be the field of an active dynamo.

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SURFACE MEASUREMENTS OF LUNAR MAGNETIC FIELDS

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Magnetic fields originating from two sources have been measured on the lunar surface: (1) remanent fields associated with magnetized crustal material, and (2) induced lunar fields associated with the solar wind and terrestrial magnetic fields.

Six magnetic field experiments have been deployed on the lunar surface by the Apollo and Lunokhod programs; one each on the Apollo 12, 14, 15, and Lunokhod 2 landings and two on the Apollo 16 lunar landing. The magnitudes of the remanent fields vary from ~ 3 gamma at the Apollo 15 site to ~ 300 gamma at the Apollo 16 site and the time dependent induction fields vary from 0 to ~ 100 gamma. Measurements of these fields have been used to determine the vector properties and scale sizes of the remanent fields, the electrical conductivity and temperature profile of the lunar interior, and the global permeability of the moon.

The three stationary lunar surface magnetometers were deployed by astronauts on the Apollo 12, 15, and 16 missions and measured the vector field 3.3 times per second continuously over a period of years. A lunar portable magnetometer was developed for the Apollo 14 and 16 astronauts to obtain measurements at preselected locations along their lunar traverses. Dolginov et al. (1977) developed a magnetometer for the automated Lunokhod 2 lunar vehicle which obtained vector field measurements every 5 seconds during several traverses in the Le Monnier Bay region.

A local remanent magnetic field of 38 ± 3 gamma was measured by the first magnetometer (Apollo 12) deployed on the eastern edge of Oceanus Procellarum. We attributed this field to local subsurface magnetized material since analyses of Apollo 11 lunar samples first demonstrated that crustal material had a natural remanent magnetism of 10^{-7} to 10^{-3} G cm³ g⁻¹ (Fuller, 1974). Simultaneous Apollo 12 magnetometer and solar wind spectrometer measurements have been analyzed and show that the lunar crust at this site is uniformly magnetized on a scale of ~ 18 km (Dyal et al., 1978). Astronauts Shepard and Mitchell used the Apollo 14 portable magnetometer in the Fra Mauro region to measure fields of 103 ± 5 and 43 ± 6 gammas at two sites located 1.1 km apart. Later in the Apollo program a steady field of 3.4 ± 2.9 gammas was measured near Hadley Rille by the Apollo 15 stationary magnetometer. At the Apollo 16 landing site in the Descartes highland region the stationary magnetometer measured 235 ± 4 gammas and astronauts Young and Duke obtained four additional measurements of 112 ± 5 , 113 ± 4 , 189 ± 5 , and 327 ± 7 gammas at locations along a 7 km traverse. These traverse measurements and subsequent solar wind plasma interaction measurements both show the local magnetic field scale size to be 5.7 km. The Lunokhod 2 magnetometer measured fields from 2 to 30 gammas in the Le Monnier Bay region (Dolginov et al., 1977). This automated lunar rover measured fields associated with craters and plains which varied in scale size from 0.5 to 7 km.

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The Apollo 12, 15, and 16 stationary and Lunokhod 2 rover magnetometers measured the induction of global lunar fields by time varying extralunar (solar or terrestrial) magnetic fields. These data were used to investigate the electrical conductivity of the lunar interior. Toroidal ($\vec{V} \times \vec{B}$) induction measurements by simultaneous Apollo 12 solar wind spectrometer and magnetometer experiments indicate an electrical conductivity of $\sim 10^{-8}$ mhos/m for an average global crust thickness of ~ 80 km. Poloidal (eddy current) induction measurements from different combinations of simultaneous orbital Explorer 35, Apollo subsatellite, and Apollo surface experiments have been used to determine the interior electrical conductivity profile. Both the time-dependent transient response technique (Dyal et al., 1976) and the frequency-dependent Fourier harmonic technique (Wiskerchen and Sonett, 1977) yield a conductivity profile rising from $\sim 10^{-4}$ mhos/m at 200 km depth to $\sim 10^{-2}$ mhos/m at 1000 km depth. This conductivity profile corresponds to a temperature profile which rises rapidly with depth from the surface to 1100 °K at 200 km and 1800 °K at 1000 km depth for a moon composed principally of olivine (Duba et al., 1974).

Simultaneous Apollo surface and Explorer 35 orbital field measurements were also used to construct a whole moon hysteresis curve, from which the global lunar relative magnetic permeability was determined to be 1.012 ± 0.006 (Parkin et al., 1974). Assuming the lunar composition to be olivine with a ferromagnetic component of free metallic iron the total iron abundance was calculated to be 5.5 ± 1.2 wt %.

In summary, the origin of the magnetized crustal regions on the moon remains unknown and a central problem in lunar magnetism. A high resolution (≤ 3 km) mapping of lunar remanent fields to correlate field topology with surface features, crustal thickness, and mineralogy would resolve the problem. The possible existence of a chemically or physically differentiated lunar core is one of the most important unknowns in studies of the thermal history and evolution of the Moon. Induction experiments involving two simultaneous lunar orbiting magnetometers and three widely dispersed surface instruments with 0.1 gamma resolution would be able to resolve this very important question.

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LUNAR PALEOINTENSITY DETERMINATIONS - A REVIEW, M. Fuller, Dept. of Geological Sciences, University of California, Santa Barbara, California 93106

The determination of the intensity of the ancient lunar fields, in which the natural remanent magnetization (NRM) of the Apollo samples was acquired, has proved to be an extremely difficult task. Moreover, this task is not yet satisfactorily completed. In addition to such obvious difficulties as paucity of sample, lack of sample history, no record of initial orientation of the sample, inability to heat samples without degradation, there now appear to be more subtle difficulties which further restrict the use of classical intensity determination techniques. Intensity determinations have by now been attempted on both breccias and crystalline rocks. The NRM of the breccias is, however, so poorly understood, and their history potentially so complicated, that it is probably wise to restrict our discussion initially to crystalline rocks. Even the intensity determinations for mare basalts and impact melt rocks, which might be expected to have the simplest histories, give inconsistent results for sub-samples from the same rock and for rocks of the same age. These results suggest that the NRM is not simply a primary NRM of thermal origin acquired homogeneously as the rocks initially cooled on the lunar surface. Nevertheless, most workers would probably agree that fields of as much as one oersted are required on the lunar surface at some time to account for the observed NRM. Whether the NRM records the ambient field at the time of the formation of the rock, or that associated with some later event in its history is not clear. In this paper the methods of intensity determination are reviewed and the results discussed, with special emphasis on the mare basalts.

The principal methods of intensity determination were developed for terrestrial samples, whose NRM was known to be a thermoremanent magnetization (TRM). The intensity estimate is obtained from the ratio of NRM to TRM acquired in a known field, assuming the process of TRM acquisition to be linear in field. The method was refined by the Thelliers (1) to take advantage of the additivity property of pTRM, so that many independent determinations could be obtained for each sample by utilizing comparisons between NRM and TRM blocked in different temperature increments. This permits one to compare the blocking temperature distribution of NRM and TRM and can reveal changes in the sample brought about by the heating. Because lunar samples are severely degraded by the necessary heating to generate TRM, a microwave equivalent of the Thelliers method was developed. In this technique exposure time plays an analogous role to temperature in the original experiment (2).

Single heating methods have been developed with independent checks for sample degradation by Shaw (3) and by Rigotti (4). In the former method the demagnetization of anhysteretic remanent magnetization is used to monitor the effects of heating, and in the latter ARM acquisition. Both of these methods can in principle be modified to correct for sample degradation. Like the Thelliers technique they use a comparison between NRM and TRM as the basis of the intensity estimate.

Other methods utilize normalization of the NRM with a remanent magnetization induced at room temperature, so that difficulties associated with heating the samples can be avoided. In the simplest technique saturation remanent magnetization (IRM_s) is used (5). This method has the advantage that it is applicable to many samples for which other data are not available. However, it is a rather crude method, relying only on the alternating field (AF) demagnetization characteristics of the NRM for interpretation of its nature. A more sophisticated technique is to normalize with ARM, using com-

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parisons in discrete increments throughout the microscopic coercivity range of the sample. However, to obtain an absolute intensity estimate, from either this technique, or the IRM_s normalization method requires a calibration of ARM or IRM_s in terms of TRM. This is of course precisely the experiment, which we have all been unable to carry out, without irreversibly changing the sample, so it is not surprising that the various calibration constants are in dispute. Probably it is best to regard both of these methods as giving rough indications of the ancient field intensity, good to perhaps a factor of 5, with the ARM method the more reliable of the two.

A successful Thellier-Thellier determination was completed using 62235, a highland impact rock and yielded a field value of 1.2 oe (6). No successful Thellier-Thellier intensity determinations have been completed on mare basalts. It was indeed attempts to carry out such determinations which led to the observation of the aberrant partial TRM acquisition (7,8). Similar behaviour has also been seen now in 12022 (9). A microwave Thellier-Thellier experiment was also attempted with this rock but a reliable intensity determination was not achieved.

A convincing Shaw (3) analysis has been reported by Hoffman et al. (10) for the fine grain Apollo 11 basalt 10017 which gave a field value of 0.71 oe. Shaw (3) and Rigotti (4) analyses with correction estimates have been carried out for 10049, a fine grain Apollo 11 basalt and gave mutually consistent results for a single sub-sample of 0.09 and 0.1 oe (11).

IRM_s normalization has been carried out for several mare basalts with calibration by TRM experiments with 12053 and 10048 (12). Estimates of approximately one oersted were obtained from mare basalts 10017, 10020, 12022 and the highland crystalline rock 62235, but 10049 gave an estimate of an order smaller.

ARM normalization has been used extensively by Stephenson et al. (6) and led them to postulate a major decrease in the field during the time span of the formation of the various rocks studied. It remains to be seen whether this analysis will stand in the light of new intensity determinations.

Despite the apparent similarity of the intensity estimates quoted above, taken as a whole, the mare basalt intensity estimates present a picture of remarkable variability. For example, for almost all of the results quoted above, other determinations from different sub-samples of the same sample have given different results. In fact, individual sub-samples generally give widely divergent results, as do individual rocks of comparable age, from similar localities on the moon. It therefore seems unwise to assume that the NRM is simply a primary TRM acquired at the time of formation of the rock. Indeed this concern is reinforced by the variability of the AF demagnetization characteristics of sub-samples from a single rock. The mare basalts, which give the highest intensity determinations tend to have a fine grain size. However, some vitrophyres give low estimates, so that the rapid initial cooling implied by fine grain size does not always correlate with high intensity estimates. Anomalous pTRM acquisition behaviour is exhibited by several of the mare basalts, which give the high intensity estimates. These samples exhibit a remarkable increase in their capability to carry remanence and in their magnetic hardness upon heating. This makes one wonder if some of the variability in the NRM could be due to inhomogeneous heating of the samples on the moon. One event in the sample's history which might give such inhomogeneous heating is impact related shock. Until the variability of the NRM from individual samples is better understood, one can only conclude that the fields required to account for NRM appear to be variable and to include fields of the order of magnitude of an oersted. When the magneti-

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zation was acquired is not yet clear. Neither is it clear what the nature of the fields may have been.

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VENUS: a) MAGNETIC FIELDS, and b) IONOPAUSE ALTITUDE
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Does Venus have a permanent magnetic field? Russell (1976) interpreted the increase in magnetic field measured by Venera-4 with decreasing altitude as due to a planetary field; Russell and Goldstein (1976) interpreted the reversal in the B_x component of the Mariner V field upon entering and leaving the magnetotail as also indicating a planetary field. However, the reversals can also be explained by assuming a prior change in the solar wind transverse magnetic field and that ionospheric field had not adjusted to the more recent solar wind draping direction. More recently, Dolginov et al. (1977) have reported on Venera-9 and -10 data in the wake of Venus. The magnetic field observations as a whole indicate a B_x component directed away from the sun in the Northern hemisphere and a solar directed component in the Southern hemisphere, in agreement with the previous estimate of direction. However, Eroshenko (in Dolginov et al.) shows that in the near wake of Venus the B_x component is directed as would be expected from draping of the x transverse component of the solar wind field.

These observations pose several interesting problems. On theoretical grounds, the induced magnetic fields are due to currents that close through the dayside ionosphere (Daniell and Cloutier, 1977). Field lines passing into the nightside tail must pass above or through the ionosphere at the terminators and close through the dayside ionosphere. However, the observed flux in the tail seems to be considerably more than can pass through the dayside ionosphere; which has a much smaller cross-sectional area than the tail. If this is in fact a real problem, we must conclude that currents are also flowing through the nightside ionosphere and allowing flux to pass through the equator beneath the ionosphere. The conductivity of the nightside ionosphere is at least a factor of ten less than the dayside due to the lower ion densities. Additionally, in the model of Daniell and Cloutier the solar wind makes effective electrical contact with the ionosphere only at low altitudes near the subsolar point; electrical contact near the terminators is inhibited by poor conductivity across field lines at high ionopause altitudes. Alternatively, we suggest that field lines crossing the equatorial plane beneath the ionopause may be connected electrically to the solar wind at higher latitudes (an open ionospheric model rather than a closed bag-like field model). Such a model would allow a significant portion of the current to close through the nightside ionosphere. However, even in this case it is clear that most of the flux will pass through or above the terminator ionosphere.

In the opposite extreme, we note that thermospheric winds will induce large currents if a permanent magnetic field is present; a dipole moment aligned parallel to the solar wind direction would be completely confined beneath the ionosphere by a toroidal current system. A perpendicular dipole can not be

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completely confined; instead, currents will flow over the poles due to $V \times B$ drift and return at lower latitudes due to the consequent electrostatic field. The magnetic flux will be pulled to the nightside and emerge from the ionosphere in a large "spot" where convection balances diffusion.

From these considerations, we believe that directly behind Venus and near the planet induced magnetic fields will dominate near the terminator and a permanent magnetic field should be better observable closer to the anti-solar point. However, Eroshenko notes that the best evidence for dominance of induced magnetic fields is in the region where we believe that changes for detecting a permanent field should be favorable. The observational and theoretical uncertainties in interpreting the Venus magnetic field data make it difficult to reach a conclusion as to the presence or absence of a permanent magnetic field at the present time.

Additionally, we have analyzed Radio Frequency Occultation (RFO) estimates of ionopause altitudes (Fjelbo et al., 1970; Ivanov-Cholodney et al., 1978) by taking into account upstream solar wind dynamic pressure, the dependence of solar wind pressure upon the zenith angle, and the variation of ionospheric pressure with the solar ultraviolet flux. The atmosphere pressure as a function of altitude obtained in this fashion is somewhat uncertain. However, a reasonable estimate of the pressure scale height between altitudes of 260 km to 320 km is about 23 ± 5 km; this figure is in good agreement with density scale heights obtained from direct RFO measurements of density. In addition, the scale height between 300 km to 450 km appears to be substantially larger, with a value of 100 ± 40 km. The increased value of the scale height depends upon two ionopause observations at zenith angles of 74° . The observation suggests a higher temperature in the upper portions of the Venus ionosphere; regions at high altitudes in some cases may be on magnetic field lines that connect to the solar wind so that cooling by electron interchange with the lower ionosphere would be inhibited. Alternatively, other interpretations such as a decreasing molecular weight or an effect of a solar wind boundary layer flow are possible.

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ORIGIN OF MAGNETIZATION IN LUNAR BRECCIAS, Wulf A. Gose, Geophysics Lab., Univ. of Texas Marine Sci. Inst., Galveston, TX 77550

The origin of magnetization in lunar breccias has been the subject of controversy. Part of the difficulty in interpreting magnetic data on breccias stems from the fact that there are no suitable terrestrial analogs with which to compare them. Secondly, none of the returned samples is an oriented bedrock sample and thus no information can be obtained from the direction of magnetization. Thirdly, the breccias contain at least two, mostly more generations of metallic iron, the dominant magnetic mineral.

The most suitable samples for magnetic studies are the samples from the boulders at the Apollo 17 landing site. These samples are unique in that a series of handspecimens were collected from one (or a few related) boulder and for some of these handspecimens it was possible to obtain relative orientations. We have reported the results on 26 samples from seven hand specimens from the station 6 boulder (Gose et al., 1978). Based on these data we propose that the natural remanent magnetization in these breccias is the vector sum of two magnetizations, a pre-impact magnetization and partial thermoremanence acquired during breccia formation. The relative contribution of the two components is controlled by the thermal history of the ejecta, which in turn is determined by its clast population (abundance and size). Depending on the clast population, the NRM can be a total thermoremanence, a partial thermoremanence plus a pre-impact magnetization, or a pre-impact magnetization.

This model of thermal overprinting seems to be applicable to all lunar breccias of medium and higher metamorphic grade. In the low metamorphic grade breccias the magnetization might well be controlled by shock effects as argued by Cisowski et al. (1974). A proper understanding of the origin of magnetization in lunar breccias is important in that the breccias provide the possibility of obtaining information outside the time range of mare basalt crystallization ages. In addition, they are the prime candidates as a source rock for the magnetic anomalies measured with subsatellites. If the above model is correct then it should be possible to map ejecta blankets or rays from orbit thereby significantly contributing to our understanding of lunar geology.

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ENERGY SOURCES FOR THE EARTH'S MAGNETIC FIELD AND CONSEQUENCES FOR THE OTHER PLANETS

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We know more about the Earth's magnetic field than we will ever know about that of the other planets, and geomagnetism is a showpiece for studies of magnetism of the terrestrial planets. The poor electrical conductivity of liquid iron means that a constant energy supply is needed to maintain the field, the rate of supply being quite close to the observed heat flux from the interior. There are several different possible sources of energy: radioactive heating, tidal dissipation and the energy release of earthquakes are all just about adequate. Numerical estimates of the energy required, that allowed, and that available are all close to one another. If we lived on another planet, it would be quite easy to deduce that Earth had no dynamo driven magnetic field at all. The terrestrial planets, plus some satellites like the Moon and Ganymede, are not that dissimilar and the calculations for the Earth serve as a warning against firm conclusions being drawn for other planets. Jupiter and the outer planets are quite different: they have no energy shortage and in many respects resemble the sun more than the Earth. The central problem here is one of explaining the temporal and spatial variations of the magnetic field rather than its very existence. This discussion will be restricted to the Earth, and comparisons with the terrestrial planets.

Dynamo theory assumes magnetic fields to be generated by fluid motions. In steady state the work done by the fluid against magnetic forces is equal to the Ohmic heating, which is one estimate of the energy requirement. It can be computed from the electrical conductivity ($5 \times 10^5 \text{ mho.m}^{-1}$) and the magnetic field. The latter is only known at the surface of the core, and the values inside must be guessed. Comparing numerical values based on lower bounds and idealised dynamo models, the Ohmic heating for the Earth's core must be between 10^9 and 10^{12} watt, depending on the spatial length scale of the total field and on how large the toroidal field is.

A power source is required to drive the fluid motions, not all of which goes to Ohmic heating. Viscosity and diffusion of different components of the liquid mixture also dissipate heat, but these effects are probably small. A lot of power is lost without ever being converted to kinetic energy of the fluid motions. Two adequate energy sources are the slow loss of the Earth's rotational energy, and seismic energy. The mechanisms available can only drive oscillatory motions directly which are ineffective for magnetic field generation, except for turbulent effects which would probably involve much larger Ohmic heating anyway. Thermal convection can be driven by radioactive heat sources or by gradual cooling of the core. Heat is converted to electrical energy with an ideal thermo-

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David Gubbins

-dynamic efficiency of only 20%. More heat is conducted down the adiabat leaving only a 5% conversion. Heat might generate a modest magnetic field without exceeding the observed heat flux, but the numerical values are uncomfortably close.

Gravitational power has received a lot of attention recently. Cooling will shrink the core which involves a loss of gravitational energy. This is particularly large where core liquid freezes because of the density contrast of about $.05 \text{ Mg.m}^{-3}$ between solid and liquid iron at core pressures. Some work is done in deformation and some appears as heat, helping to fuel a thermal dynamo. The large density jump at the inner core boundary indicates that there is a compositional difference between the inner and outer cores. In this case freezing of the liquid will lead to freezing out of a heavy, iron-rich component, leaving a light fraction in the outer core. The light material rises up, driving convection directly. Some light material may leak away by molecular diffusion, but most of the available energy is dissipated in magnetic fields. Hundreds of Gauss can be sustained while keeping within the Earth's thermal budget, and this makes gravitational power an attractive contender for the Earth's dynamo.

A similar idea is that the core originally formed near the melting point and solid particles have been settling out ever since, stirring the liquid. This problem is bound up with that of core formation. Further complications like that of phase changes have been studied and lead to interesting dynamical consequences. But the controlling factors are the heat loss through the mantle and the initial thermal state of the core, which will determine the cooling and differentiation rates. Comparisons with other planets must incorporate these two features accurately. The other terrestrial planets all have small magnetic fields that could be due to magnetised rocks. Only the demonstration of a liquid core or the discovery of secular variations would prove a dynamo driven field. The cooling of the planetary core is determined by the complete energy budget discussed above, and the uncertainties are too great to make sound deductions about whether particular planetary cores are molten or not.

A NEW METHOD FOR DETERMINING THE RADIUS OF THE ELECTRICALLY-
CONDUCTING FLUID CORE OF A PLANET FROM EXTERNAL MAGNETIC OBSERVATIONS:
RECENT DEVELOPMENTS.

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The new method (Hide, R., 1978, *Nature* 271, 640-641) exploits the applicability of Alfvén's "frozen field" concept when dealing with changes in the planetary magnetic field over intervals that are much less than the time-scale of ohmic decay. The outer radius r_c of the electrically-conducting fluid core is determined from magnetic observations near the surface of the planet $r=r_s$ by finding from those observations that surface $r=r_c$ within the planet at which the time variation of the total number of lines of magnetic force $N(r)$ (equal to $4\pi r^2$ times the average magnitude of the vertical component of the magnetic field) vanishes. When applied to the Earth (Hide, R., loc. cit.; Hide, R., Lowes, F.J. and Malin, S.R.C., in preparation), for which $r_s = 6371$ km, the method gives for r_c a value which differs insignificantly, by less than 2%, from the seismological value of 3486 ± 5 km. This determination is based on one of the best available models of the geomagnetic secular variation, where determinations of the main geomagnetic field (namely that field obtained when contributions due to ionospheric and magnetospheric currents have been eliminated by taking annual mean values of the observations) for the period 1965 to 1975 were fitted by a spherical harmonic series up to and including terms of degree 8. The apparent success of the method when thus tested on terrestrial data indicates that it might prove valuable in the investigation of the internal structure of magnetic planets with unknown values of r_c , such as Jupiter, Saturn and Mercury, provided that those responsible for planning future missions with space probes ensure that sufficiently detailed measurements are made of the spatial configurations of the magnetic fields of these planets and of secular changes.

HOW STRONG IS THE MAGNETIC FIELD IN THE EARTH'S LIQUID CORE?

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The geomagnetic field in its region of origin, the Earth's liquid metallic core, consists of a poloidal part $\vec{B}_p$, with lines of force that emerge from the core and penetrate to the Earth's surface and beyond, as well as a toroidal part $\vec{B}_T$, of unknown strength, with lines of force that have no radial component and are thus confined to the core. Toroidal fields are readily produced by inductive interactions between poloidal fields and horizontal fluid motions and it is likely but by no means certain that $\vec{B}_T$ greatly exceeds $\vec{B}_p$ in magnitude: since the volume integral over the whole core of $(\vec{\nabla} \times (\vec{B}_T + \vec{B}_p))^2 / \mu^2 \sigma$ is a lower limit to the power requirement for the geomagnetic dynamo (where σ is the electrical conductivity and μ the magnetic permeability) any discussion of the energetics of the dynamo process must involve inter alia the estimation of $\langle B_T \rangle$, the average magnitude of $\vec{B}_T$.

We review here various methods that have been used to estimate $\langle B_T \rangle$ and we propose a new method, which exploits certain dynamical considerations of hydromagnetic flow in a rotating fluid. The determination of the electric field at the surface of the Earth that would be associated with $\vec{B}_T$ in the core is in principle the only direct method, but the electric field is so weak as to be undetectable. The earliest "eddy" models of the geomagnetic secular variation require $\langle B_T \rangle \approx 300\Gamma$, whilst the interpretation of the geomagnetic secular variation, including the westward drift, in terms of hydromagnetic oscillations (strongly modified by rotation) of the core requires $\langle B_T \rangle \approx 100\Gamma$. The various kinematic dynamo models that have been applied to the Earth give a range of values of $\langle B_T \rangle$, from the 5Γ associated with " α^2 -dynamos" to 1750Γ associated with " $\alpha\omega$ -dynamos". The new method explored here, exploits determinations of features of the flow field in the upper reaches of the core on the basis of the "frozen field" hypothesis. When combined with dynamical considerations of the radial component of vorticity an expression can be obtained for the vertical gradient of $\vec{B}_T$ in terms of the flow field and $\vec{B}_p$.

BEHAVIOR OF THE GEODYNAMO DURING THE MATUYAMA-BRUNHES REVERSAL,
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Paleomagnetic records of polarity transitions are seen to display a dependence on the site of observation when paths of the virtual geomagnetic pole (VGP) are compared (1,2). For example, nearly all available reverse-to-normal (R→N) transition paths from sites in the northern hemisphere are found to reside on the hemisphere centered about the site meridian, the so-called near side. In particular, data corresponding to the Matuyama-Brunhes reversal, for which there presently exist records from five site localities scattered about the northern hemisphere, well illustrate this claim when all paths are plotted with respect to a common site longitude (Fig. 1). The absolute path locations along with the associated sites (S1 through S5) are shown in figure 2.

The behavior depicted in figure 1 indicates that the field configuration during this R→N transition was controlled by low-order zonal component(s). Moreover, such a situation may be roughly modelled by considering the reversal process to start either at low latitudes or in the southern hemisphere of the core and subsequently extend axisymmetrically until the entire core has been affected (1,2).

These zonal "flooding" models predict for the Matuyama-Brunhes (or, for that matter, any R→N transition) VGP paths which run along the site meridian for the case when the site is located in the northern hemisphere. They cannot explain, however, why most paths (although on the near side) are not found in close proximity to the site meridian (see Fig. 2).

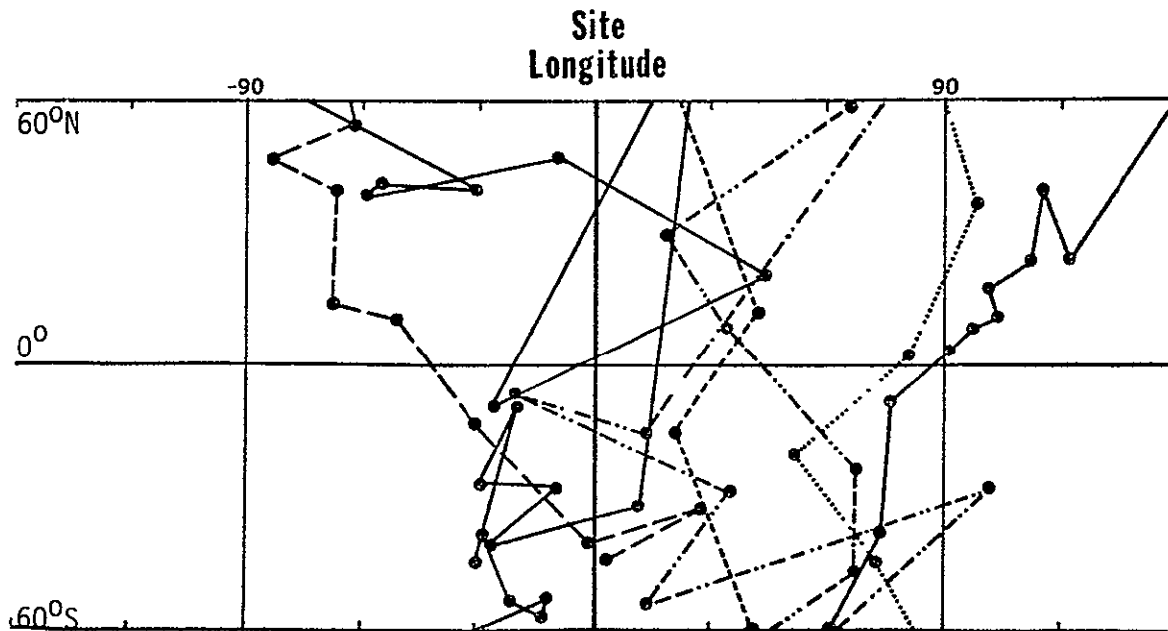


Figure 1: VGP paths corresponding to seven Matuyama-Brunhes reversal records from five northern hemisphere site localities (see Fig. 2), and plotted with respect to a common site longitude. The "near side" lies between -90° and $+90^{\circ}$ meridians.

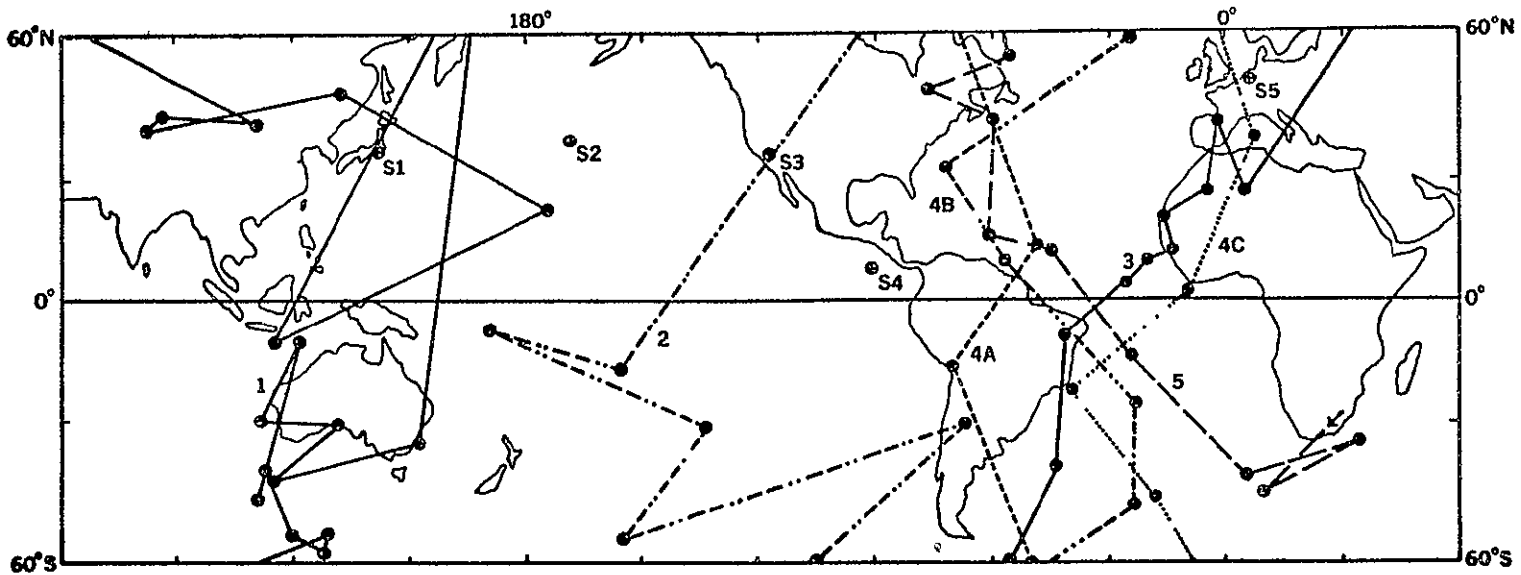


Figure 2: Matuyama-Brunhes VGP paths from sites S1 (4), S2 (5), S3 (6), S4 (7), and S5 (8).

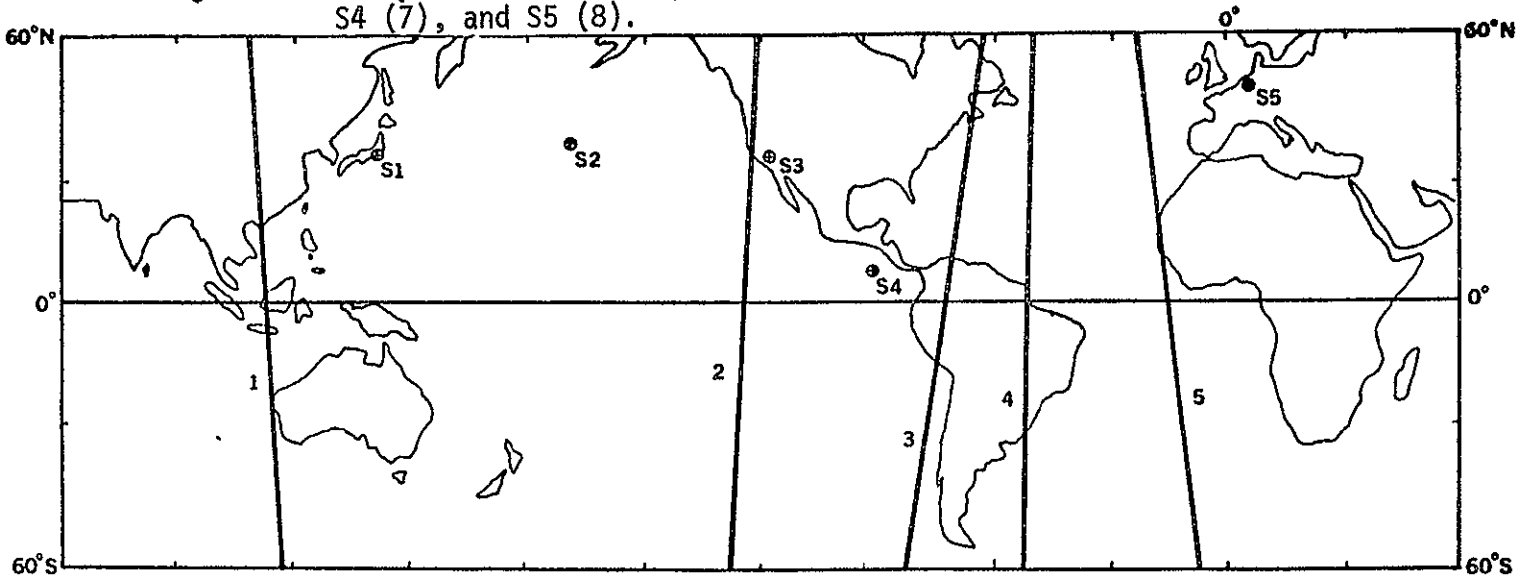


Figure 3: VGP paths predicted by the model for site localities S1 through S5.

BEHAVIOR OF THE REVERSING GEODYNAMO

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In order to account for this non-axisymmetric behavior, the flooding approach has been extended so as to incorporate both azimuthal as well as latitudinal proliferation of magnetic field of opposite sense throughout the core from a point of origin. That polarity transitions may initiate locally in the core has been previously suggested (3).

For the present model cylindrical geometry of the magnetic source region has been chosen for computational simplicity. Starting from a point on the periphery of the cylinder, flooding is assumed to occur symmetrically in the azimuthal $+\phi$ and $-\phi$ as well as along the $+z$ and $-z$ directions. The five variables associated with the model include the dimensions of the source region, the coordinates of the point at which the reversal originates, and the ratio of the flooding velocities. By constraining the transition to start at the equator of the core and applying two other reasonable assumptions reduces the number of fitting parameters to two; namely r , representing the apparent equatorial depth of the magnetic sources, and ϕ_p , the longitude at which the reversal originates.

The predicted paths associated with the site localities S1 through S5 are shown in figure 3 for the case when $r=0.13 r_e$ (r_e =radius of the earth) and $\phi_p=338^\circ\text{E}$. Comparison of figure 2 with figure 3 reveals a surprisingly good fit of the observed Matuyama-Brunhes reversal data. Taking as a quantitative measure of the goodness of fit for each case the absolute value of the angular distance between the observed and predicted equatorial crossings, one finds a range from 0° to 39° with an average of only 23° .

Validity of the modelled solution also depends on its ability to accurately reflect the observed variation in field intensity. In this regard, only one acceptable intensity record (from site S3 (6)) is presently available and the predicted intensity variation for this site is found to be in close agreement with that observed.

Hence, these results show that the non-axisymmetric flooding approach to the geomagnetic reversal process provides a close simulation of observed world-wide transitional field behavior, at least for the presently available records of the Matuyama-Brunhes.

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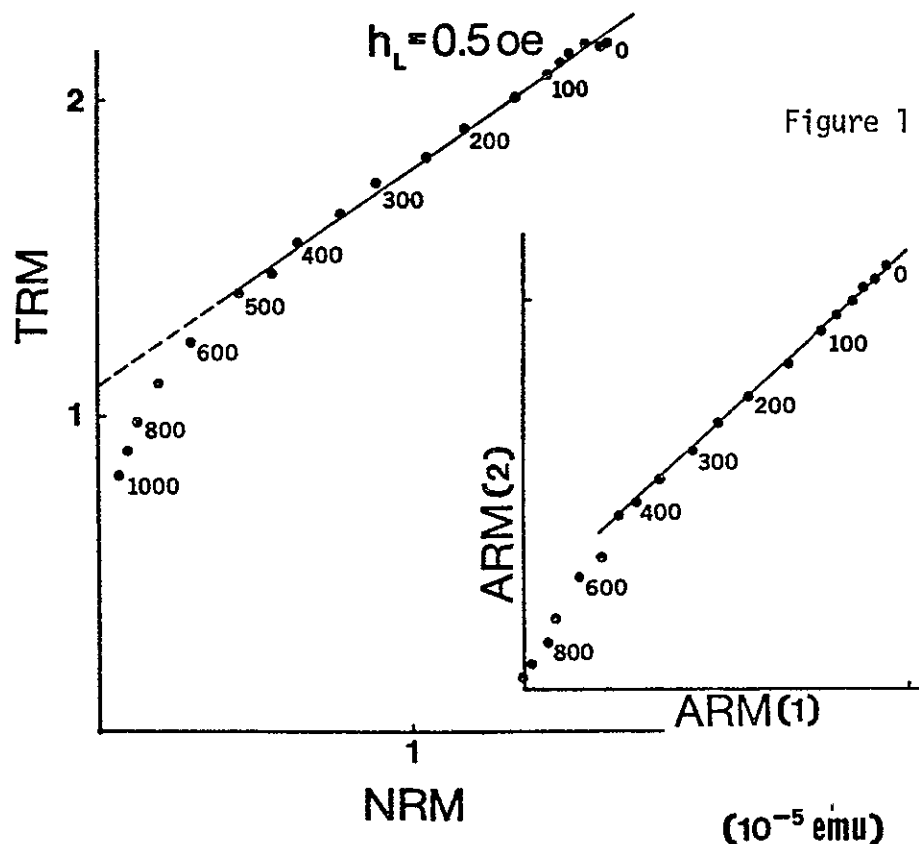
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COMBINING PALEOINTENSITY METHODS: A DUAL-VALUED DETERMINATION ON
SAMPLE 10017,135, K.A. Hoffman*, J.R. Baker and S.K. Banerjee, Dept. of Geol.
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The Shaw method (1) for determining paleointensity involves the comparison of demagnetization characteristics of a sample's natural remanent magnetization (NRM) with that of a laboratory induced thermoremanent magnetization (TRM). In contrast, the method developed by Stephenson and Collinson (2) relies on a similar comparison of NRM with an anhysteretic remanence (ARM). A single heating of the sample to above its Curie point is required, however, by both methods since for the latter case the ratio of TRM to ARM efficiency f' must be known. Accounting for the alteration of the magnetic carriers during heating is accomplished for the Shaw method through the comparison of ARM's induced prior to and after heating, however, no such correction procedure has been proposed for the "ARM-method".

We suggest that the NRM versus TRM plot necessary for the Shaw method determination may be used to make the needed correction to f' , provided that an unaltered coercivity range is recognizable. Hence, if this condition is satisfied, two primarily independent paleointensity values may be determined for a particular sample, one serving as a check of the other.

Data corresponding to sample 10017,135 are shown here for illustration. The NRM for this fine-grained vesicular basalt sample is found to be stable and well-behaved in direction during AF-demagnetization similar to the behavior previously reported for other chips (3,4). The plot of NRM demagnetization versus that of a 0.5 oe TRM is shown in figure 1. The straight line drawn through the data corresponding to AF steps between 0 and 500 oe strongly suggests at most a minimal amount of alteration to this coercivity range occurred during heating.



COMBINING PALEOINTENSITY METHODS

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Experiments conducted on synthetic multi-domain Fe indicate that ARM is invariably more sensitive to alteration than is TRM. The plot of an 0.5 oe ARM induced in 10017,135 before and after heating (Fig. 1), however, indicates behavior which closely resembles the NRM-TRM plot thus confirming the lack of perceptible alteration to the lower coercivity range upon heating. The Shaw-type determination renders a paleointensity of 0.71 oe, similar to a previously reported value (4).

The ARM-method determination may be made by noting that an extrapolation of the linear segment to the TRM intercept (Fig. 1) makes possible the recovery of the magnitude of TRM which would have been acquired by the sample if no alteration took place. (Indeed, the NRM-TRM plot indicates that the principal effect of heating this sample was to produce highly stable carriers. Hence, the total TRM value, if uncorrected, would result in an inflated value of f' .) With the value of ARM acquired in an infinite alternating field, determined from a plot of 0.5 oe ARM acquisition versus $1/H$, f' is found to be equal to 0.94. Theoretically, this value should be no less than unity (5); however, a study of magnetite-bearing samples (6) casts doubt on this prediction for the case of multi-domain grains.

Figure 2 shows ARM acquisition versus NRM lost for the indicated AF's for sample 10017,135. The plot is remarkably linear and its slope coupled with the determined value of f' render a paleointensity of 0.93 oe. Thus, the average paleointensity corresponding to this 3.6-3.8 AE lunar basalt from these two primarily independent determinations is 0.82 oe, a value which lends

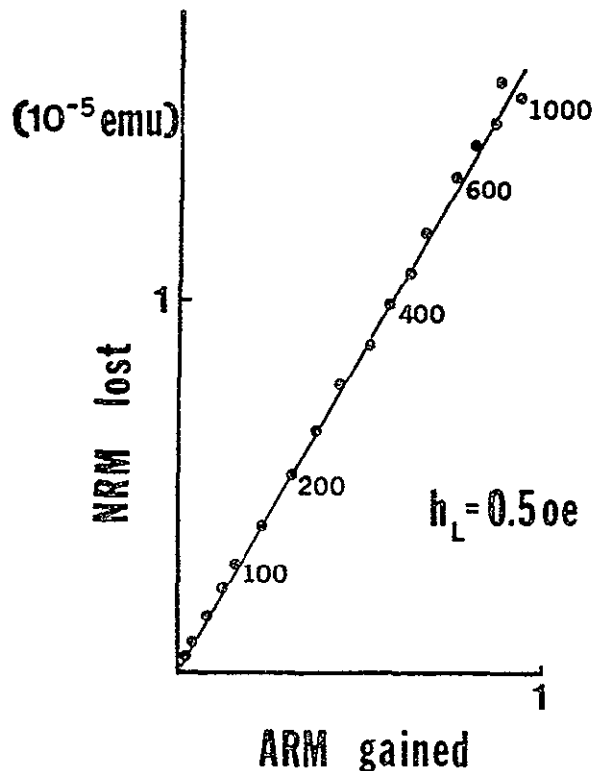


Figure 2

COMBINING PALEOINTENSITY METHODS

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support to the earlier suggestions of relatively strong paleofields experienced on the lunar surface.

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APOLLO SUBSATELLITE MAGNETOMETER DATA: CONSTRAINTS ON THE
INTERPRETATION OF LUNAR MAGNETISM, L. L. Hood, Institute of Geophysics
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The major result pertaining to lunar fossil magnetism of Apollo orbital magnetometer experiments is the detection of widespread magnetic anomalies (Coleman et al., 1972) whose invariance on successive, closely spaced orbits and with externally applied field strength and orientation suggests crustal sources of coherent magnetization. Additional results established to date include:

- (a) An upper limit for the global permanent dipole moment of $\lesssim 0.2\gamma R_m^3$ ($1\gamma = 10^{-5}$ Gauss; $1 R_m = 1$ lunar radius = 1738 km) (Russell et al., 1975). Thus although coherent remanent magnetization must exist on a local scale to explain measured surface anomalies ranging from 3 to 300γ in magnitude, a uni-directional magnetization of anomalous source regions (or any significant portion of the crust) is ruled out.
- (b) No general correlation of orbital anomaly maxima with prominent surface morphology such as the larger craters (Coleman et al., 1977b; Hood et al., 1978b). By default, compositional contrasts rather than topographic variations are the indicated causes of most observed anomalies. Several lines of evidence suggest that concentrated deposits of crater and basin impact-generated ejecta materials may be expected to possess a stronger than usual remanent magnetization and could explain the majority of anomalies detected from orbit (Strangway et al., 1973). The linear magnetic anomaly apparently associated with the rille Rima Sirsalis (Anderson et al., 1977) is difficult to explain on this basis, however.
- (c) A depletion in number and amplitude of anomalies detected across the front side maria relative to those detected across the far side highlands (e.g. Coleman et al., 1977a). This is consistent with returned sample studies which show that mare basalts typically possess a lower stable remanence ($\sim 10^{-6}$ e.m.u./g) than some classes of breccias ($\sim 10^{-4}$ e.m.u./g).
- (d) A correlation of remanent magnetism and crustal geology on the central far side (Russell et al., 1977). In particular, a strong dominance of Imbrian aged crater melt and ejecta debris is present in the Van de Graaff-Aitken region roughly coincident with the occurrence of large magnetic anomalies. One interpretation is that the ancient magnetizing field was significantly stronger during the Imbrian period than during earlier epochs. An alternative interpretation is that highly susceptible ejecta was preferentially deposited at that location (nearly antipodal to the present Imbrium basin) at the time of the Imbrium impact event (Hood et al., 1978b). The second possibility is rendered more plausible by the existence of a second region of strong magnetic anomalies visible on the high resolution electron reflectance maps (frontispiece, Proc. Eighth Lunar Sci. Conf., Vol. 1) and located nearly antipodal to the Orientale basin. Further, some lunar geologists (e.g. Stuart-Alexander, 1976) have independently noted the antipodal deposition theory in discussing the origin of peculiar grooved terrain found in both areas as well as an area antipodal to the Caloris basin on Mercury. "Gardening" of older magnetized deposits by subsequent bombardments could also have contributed to the relative magnetic strength of younger Imbrian and Orientale ejecta materials.

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- (e) Inferred directions of magnetization for major source regions across the central far side are generally highly inclined with respect to one another but seldom possess a strong north-south component (Hood et al., 1978a,b). The latter characteristic is difficult to explain in terms of entirely local, non-global causes of the magnetizing field and implies that any assumed internal magnetizing dipole or external linear magnetizing field must have been oriented mainly perpendicular to the present spin axis. The lack of alignment for nearby inferred magnetization vectors suggests that reorientation with respect to the Moon of the magnetizing field may have typically occurred on time scales less than the mean time between the formation of nearby source regions. This characteristic may also represent an explanation for the negligible global dipole moment (Hood et al., 1978a).

Only a fraction of usable magnetometer data (that collected in the geomagnetic tail lobes) has been mapped. A large amount of excellent unmapped data exists for times when the Moon was in the solar wind and the subsatellite was at a relatively low altitude in the lunar wake. We are in the process of mapping all usable data using methods recently developed for mapping the geomagnetic anomaly field. An exact accounting for variations of the subsatellite altitude is one feature of the new mapping procedure. Figure 1 shows a prototype map constructed from ~30 Apollo 15 lunar wake orbit segments. The large anomaly is evidently due to a relatively isolated source region whose bulk magnetization vector is nearly radially inward. Its amplitude (~2.7γ at 88 km) is comparable in magnitude to the Van de Graaff-Aitken anomalies and its location (235° E. Long., -23° N. Lat.) is approximately antipodal to that of the Crisium basin.

Acknowledgment

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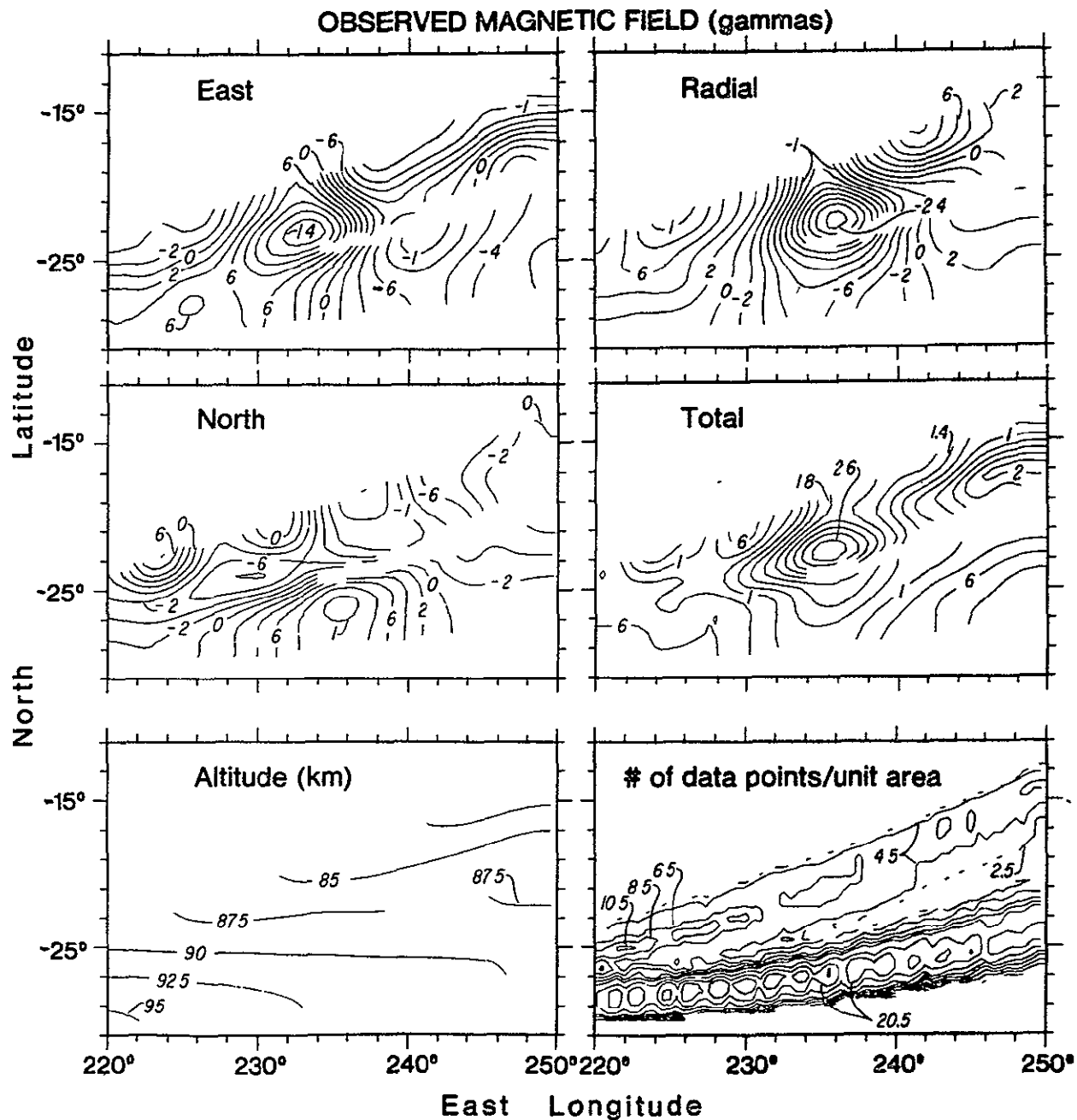


Figure 1

A Tidal-Precessional Dynamo for the Terrestrial Planets H. HOUBEN, Laboratory for Planetary Studies, Cornell U.- Recent developments in the exploration of the terrestrial planets including the discovery of Mercury's intrinsic magnetic field (Ness et al., 1975) and the possibility of an ancient lunar magnetic dynamo (Runcorn, 1975) require the reexamination of theories of planetary magnetism, including the geodynamo. In this light it is interesting to note that a dynamo driven by the combined action of diurnal tides and precession-similar to that proposed by Boudi and Lyttleton (1953) and Mallaas (1968)-can be shown to be viable for the terrestrial planets and to predict correctly the magnetic moments of Mercury, the ancient Moon and (with certain provisos) the Earth.

The magnetic dynamo equation is converted into two equations with very different timescales: 1) an equation for the diurnally varying magnetic field (for which the frozen field approximation is valid) and 2) an equation for the slowly varying average field (with time variations on the order of the free decay time of the field) in which the $V \times H$ term is the time-average product of the diurnally varying fields. The problem is solved for a spheroid of small ellipticity in conjunction with the fluid dynamic momentum equation for an incompressible fluid. A large toroidal field-satisfying the boundary conditions for a conducting sphere surrounded by an insulating mantle-is assumed to exist. It is shown that such a field can be sustained by dynamo action in the sphere. A steady solution is found and the magnitudes of resultant toroidal and poloidal field components are calculated.

The following are important intermediate results.

1) If the toroidal fields have the correct radial dependence to satisfy the boundary conditions-spherical Bessel functions (Backus, 1958)-the Lorentz force cannot be balanced by pressure gradients and a differential rotation is driven (cf. Busse, 1975). When a poloidal field is present, this differential rotation (acts to regenerate the toroidal fields (Gubbins, 1974). The energy for this process comes from the planetary rotation, as the spinup time to uniform rotation is short in conducting cores.

2) The diurnal tides and precession are nearly resonant with the spheroid's toroidal Spinover mode (Greenspan, 1968). The strange motion of this mode does not lead to dynamo action by itself, because it has no radial component (Rochester et al., 1975). However, the motion does lead to a diurnally varying magnetic field. As a result, there are diurnally varying Lorentz forces in the conducting core. These Lorentz forces, in turn, can drive motions which have radial components and which are candidates for dynamo action.

3) When the frozen field approximation is valid, the interaction of oscillatory velocities and an axisymmetric magnetic field leads to an oscillatory magnetic field which is parallel to the velocity field (Mallaas, 1967). Thus, there can be no dynamo action from the $V \times H$ term. However, in the presence of a non-axisymmetric toroidal field, there can be a net $V \times H$ term.

4) A steady solution to this problem can be found for the eight magnetic field components given by combinations of poloidal-toroidal, dipole ($\ell=1$) - quadrupole ($\ell=2$), and axisymmetric ($m=0$) - antisymmetric ($m=1$). In fact, these are two such solutions given by the relative poloidal field strengths at the core surface listed in Table 1.

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Table 1.

	Solution 1	Solution 2
g_1^0	1.0	0.23
g_1^1	0.3	0.8
g_2^0	0.4	0.21
g_2^1	0.13	1.0

The first solution corresponds to the normal arrangement of planetary magnetic fields—an axial dipole tilted by 16° with a fairly large relative quadrupole moment. The second solution is a nearly equatorial dipole—which may be appropriate for the ancient moon (S.F. Dermott, personal communication).

The results of the calculations for the terrestrial planets are given in Table 2.

Table 2.

	Mercury	Earth I	Earth II	Moon
R core (cm)	1.5×10^8	3.5×10^8	3.5×10^8	3×10^7
Vr (cm/sec)	10^{-2}	2×10^{-6}	1.3×10^{-2}	2×10^2
surface g_1^0 (G)	0.015	0.005	0.23	100.
Maximum toroidal field (G)	1.6×10^3	180.	9×10^3	2×10^4

The results for Mercury show good agreement with those of Ness et al. (1975). The results for the Moon (at an assumed distance of 10 Earth radii) are also quite good (Runcorn, 1978). In the case of the Earth, good agreement is only obtained if it is assumed that the Earth's core response to the precessional force is enhanced by a factor of about 700 (Earth II). This is possible because the Earth's inner core, being rigid (Jacobs, 1975), should be quite spherical since it is loaded by a fluid of almost equal density (Houben, 1978). The near resonance of the spinover mode is then enhanced by several orders of magnitude over that for a spheroid of ellipticity $1/400$ (Earth I).

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THE EARTH'S CORE

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The major source of our information about the Earth's interior comes from seismology. Travel-time curves can be used to construct velocity-depth curves and from them the variation of density and other physical properties within the Earth. More recently, travel-time curves have been supplemented by free oscillation data. Considerable advances have been made in dealing with this inverse problem. Many Earth models have now been constructed and bounds placed on the variables. However the broad division of the Earth into crust, mantle and core has been known for some time - Oldham deduced the existence of a core in 1906. It must not be forgotten that seismology only gives us a "snapshot" of the interior of the Earth as it is today and gives no information about its structure in the past nor of its evolution.

In 1936, Miss Lehmann suggested that the liquid core contained an inner core which is solid. More recent evidence that the inner core is solid is reviewed. The problem of the origin of the core is then discussed - which question cannot be divorced from the much broader and more difficult problem of the origin of the solar system and the constitution of all the planets. This leads naturally to the question of the thermal evolution of the Earth and a discussion of possible sources of heating. Apart from heating due to accretion and the impacts of large falling bodies, possible heat sources include the decay of long and short lived radio nuclides, adiabatic compression, tidal dissipation, induced electric currents and the release of gravitational energy due to the formation of the core. In addition to initial conditions and heat sources, the thermal regime of the Earth depends on the adiabatic and melting (or liquidus) temperatures. Our knowledge of both these temperatures is very rudimentary.

Additional information about the Earth's core comes from its magnetic field. It is generally believed that the geomagnetic field is generated by some form of dynamo action in the core. Since rocks as old as $\sim 3,000$ m. yr have been found possessing remanent magnetism, a fluid (outer) core must have been in existence at least that long ago. A fundamental question in geomagnetism is what is the driving mechanism, of the geodynamo. If some form of convective motion in the outer core is demanded, this places further constraints on the thermal regime of the core and its chemical composition. Since dynamo theory is considered in other sessions, it is not discussed in any detail here. However the stability and possible stratification of the core is important in this connection and this question is examined more fully. Finally, the chemical composition of the core is briefly reviewed.

THE ELECTRICAL AND THERMAL CONDUCTIVITY OF VARIOUS TERRESTRIAL CORE MATERIALS TO 1.8 MB[†] by R. Norris Keeler, George M. Matasov*, A. C. Mitchell and William J. Nellis, University of California, Lawrence Livermore Laboratory, Livermore, CA 94550

Electrical conductivity measurements have been extended to 1.8 Mb for the core materials Fe, Fe-Si and Fe-Ni. New measurements have been made for Fe-O, Fe-S and Fe-C. All conductivities fell within the currently accepted Magnetic Reynolds-Number criterion for dynamo action. The wide limits and uncertainty of this criterion points to the need for better quantitative theories of the earth's dynamo to make high pressure conductivity measurements more meaningful in selecting the most probable core materials.

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[†]Work performed under the auspices of the U.S. Department of Energy by the Lawrence Livermore Laboratory under contract number W-7405-Eng-48.

FUTURE PLANETARY MISSIONS; Margaret Galland Kivelson, Department of Earth and Space Sciences and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, CA 90024

For almost four centuries after Gilbert recognized that the magnetic field of the earth approximated that of a magnetized sphere, the study of planetary magnetism was limited to the study of the earth. Terrestrial measurements demonstrated the complex and unsteady nature of magnetic processes and defined an observational framework for theoretical analysis. The formulation of a complete and self-consistent theory of planetary magnetism in the future will undoubtedly be stimulated by comparative studies based on knowledge of the magnetic properties of other objects in the solar system including such diverse bodies as major planets, their satellites, comets and asteroids. Simultaneously, problems connected with remanent magnetism should be elucidated. From closeup spacecraft observations, we have already begun to establish the properties of Mercury, Jupiter, the Moon, Venus and Mars, although the interpretation of the magnetic properties of the latter two planets has been somewhat controversial. Indeed, there are many questions whose resolution must rely on future extraterrestrial missions.

The table below lists missions which are expected to provide data on the magnetic properties of extraterrestrial solar system bodies in the near future. The list is chronologically ordered by launch date, with missions near the end of the table not yet approved but well into the planning phase.

LAUNCH	SPACECRAFT	TARGET (S)	COMMENTS
1973	Pioneer 11	Jupiter (12/74), Saturn (9/79)	Flybys
1977	Voyager 1	Jupiter (3/79), Saturn (11/79)	Flybys
1977	Voyager 2	Jupiter (7/79), Saturn (8/80)	Flybys
1978	Pioneer Venus	Venus (12/78)	Orbiter
1982	Galileo	Mars (82), Jupiter orbit (85)	Satellite flybys
1983	VOIR	Venus (83)	Magne- tometer?
1983	Solar Polar (2)	Jupiter (84), Solar pole (86)	Flybys
1985	Comet Encke	Encke (87)	Flyby
1986	Mars Geochemi- cal Orbiter	Mars (87)	Orbiter
1987	Saturn Orbiter Dual Probe	Saturn (92)	Titan, rings
1988	Mars Sample Return	Mars (89)	Magne- tometer?

All of these missions will contribute to our understanding of planetary magnetism. Flights by Saturn will test the models we use to infer the surface field strengths of remote objects. Similarities and differences between the fields of Saturn and Jupiter will provide valuable insight into the properties of planetary dynamos. Orbits of Mars and Venus should help settle the question of whether either or both of these planets have intrinsic magnetic fields and clarify the nature of their interactions with the solar wind. Only peripherally noted in the table are the Galilean satellites which are objects of intense interest to anyone studying the formation of the solar system. These near Moon-sized bodies form a mini-solar system circling Jupiter. Close flybys planned for Voyager and later for Galileo will provide data on their interactions with the Jovian plasma and their intrinsic magnetic properties.

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Ultimately, similar studies of the moons and rings of Saturn will supplement our knowledge of such bodies. However, it should be recognized that detailed mapping of surface anomalies familiar from Apollo investigations require multiple passes and will not result from flyby measurements.

Not entered in the table are several missions on the "wish list" which have already been studied carefully. Examples include: a multi-asteroid reconnaissance mission which would look for remanent magnetic fields, a Galilean satellite lander, a flyby of Uranus whose magnetic dipole is expected to be almost in its orbital plane resulting in unique interactions with the solar wind, and a Saturn orbiter with probes. Finally, support for a lunar polar orbiter which could provide magnetic maps of the entire surface of the Moon is still strong in the scientific community and the European Space Agency is now considering the support of such a mission.

DYNAMO MODELS FOR THE EARTH
 BASED ON MEAN-FIELD MAGNETOHYDRODYNAMICS

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Some new results on dynamo models are represented which can be applied to the Earth. The models consist of a sphere of electrically conducting matter surrounded by vacuum. As usual in mean-field magnetohydrodynamics both the magnetic field and the motion are understood as a sum of a mean and a fluctuating part. According to the assumptions on the mean motion two different types of models are considered: one with a mainly rigid body rotation and one with a non-uniform rotation. As far as the mean electromotive force caused by the fluctuating motions is concerned in addition to the α -effect also other effects are included.

Models with rigid body rotation (which operate due to an α^2 -process) make allowance for mean magnetic fields with different symmetry properties with respect to the rotational axis and different time behaviour. In addition to axisymmetric stationary fields there are non-axisymmetric fields in the form of waves travelling around the equator. Also superpositions of different fields may occur. This leads to a possibility of not only explaining the existence of the dipole-like magnetic field of the Earth but also the deviation of this field from symmetry with respect to the rotational axis and its westward drift.

In models with non-uniform rotation (which operate due to $\alpha\omega$ or $\int\omega$ -processes) axisymmetric magnetic fields are favoured, which may be either stationary or oscillatory. This provides for an interpretation only of that part of the magnetic field of the Earth which is symmetric with respect to the rotational axis. The behaviour of the remaining part can, e.g., be understood in terms of MAC-waves.

PLANETARY DYNAMOS--THE KINEMATICAL THEORY, E. H. Levy, Department of Planetary Sciences and Lunar and Planetary Laboratory, University of Arizona, Tucson, Arizona 85721.

The primary magnetic fields of most large astrophysical objects, including planets, are thought to be generated by a hydromagnetic dynamo process. In this process motions of an electrically conducting fluid interact with a magnetic field to generate electrical currents which produce the field. The dynamo phenomenon is described physically and mathematically by Maxwell's equations and Newton's laws, along with appropriate constitutive relations to describe the material properties of the fluid. Complete solution of the problem, including simultaneously both the electromagnetic and mechanical behavior, is a formidable task which has so far resisted definitive attack. However, tentative advances are being made against this problem based on the known solutions of simpler fluid dynamical problems and the results of kinematical dynamo analysis of magnetic field generation.

Essentially all of our understanding of the kinds of fluid motions which are capable of generating magnetic fields comes from the kinematical theory. In this approach, the problem is truncated by assuming that the fluid velocity is given and fixed. Thus Newton's laws are dropped from consideration. Of course, most interest centers on fluid motions similar to those occurring in natural bodies.

We review kinematical dynamo theory and its impact on our understanding of the magnetic-field generation process. Included is a discussion of recent and ongoing work. In this context we briefly review our ideas about the reversal phenomena displayed by magnetic astrophysical objects.

MEASUREMENTS OF LUNAR MAGNETIC FIELDS BY THE ELECTRON REFLECTION METHOD by R. P. Lin and K. A. Anderson, Space Sciences Laboratory, University of California, Berkeley, California 94720.

The electron reflection method for remote sensing of lunar surface magnetic fields depends on the fact that charged particles tend to be reflected from regions of increased magnetic field strength. Charged particles in the cis-lunar environment are guided to the lunar surface by whatever ambient magnetic field is present. If the field strength does not increase toward the surface, the particles will be absorbed by the surface material. If fields of near-surface origin are present, the total field strength will increase as the particles approach the surface, causing a fraction of the particles to be reflected back with an intensity that increases with the total field strength. Apollo 15 and 16 subsatellite measurements of the angular distribution of 0.5 and 14 keV electrons incident on and reflected from the lunar surface have been used to obtain high spatial resolution ($\sim 10 - 50$ km) and sensitivity ($\sim 0.1 \gamma$, $1 \gamma = 10^{-5}$ Oersted) maps of lunar surface magnetic fields. It is found that many patches of increased field strength ranging in size from < 10 km to > 200 km are distributed over the lunar surface. We have searched for correlations of these magnetic features to surface geology. In general there is no obvious correlation. The exceptions are as follows:

- 1) Surface fields of $> 100 \gamma$ are present in a region of order 10 km wide and > 300 km long centered on and paralleling the long linear rille Rima Sirsalis. The magnetic feature extends 60 km beyond the place where the rille disappears beneath the lava flows of Oceanus Procellarum. These magnetic results imply that the rille has a strong magnetization ($> 5 \times 10^{-5}$ gausscm³ gm⁻¹) associated with it, either in the form of intrusive, magnetized rock, or in the form of a gap in a uniformly magnetic layer of rock.

- 2) In lunar mare regions a statistical correlation is found between the surface field strength and geologic age of the surface as determined from crater erosion studies.

- 3) However there is a lack of correlation of surface field with impact craters in the mare. This suggests that the mare do not have a strong uniform magnetization and that these impacts do not in themselves generate strong magnetization coherent over ~ 10 km scale size.

- 4) There appears to be a correlation of strong surface field regions on the lunar backside with ejecta from large craters and basins.

The implications of these findings on the origin and history of lunar magnetism, and the application of this measurement technique to other planetary bodies will be discussed.

ARE PLANETARY DYNAMOS DRIVEN BY GRAVITATIONAL SETTLING?

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Several of the planets are known to possess magnetic fields. It is generally believed that these fields are sustained by dynamo action within the planetary interiors. One of the most challenging problems of planetary physics concerns the energy source for these dynamos. The two energy sources mentioned most often in the literature are thermal convection and precessionally induced motions, but each of these mechanisms encounters difficulties in efficiently converting the available energy into mechanical motion: in a metallic fluid such as the Earth's core, thermal convection may be short-circuited by thermal conduction while the energy of precession appears to be dissipated in boundary layers. In this paper we shall elaborate upon a third possible energy source which is not subject to these problems of efficiency: the energy released by gravitational settling. This idea was first proposed by Braginsky (1963) and has recently been discussed by Gubbins (1977) and Loper (1978 a, b). A complete theory of the operation of dynamo driven by gravitational settling is currently being developed (Loper and Roberts, 1978).

The basic features of gravitationally driven convection may be summarized as follows. Consider an initially homogeneous self-gravitating sphere of molten binary alloy representing a planetary core, the alloy being composed of a heavy metal (principally iron) and a light nonmetal in the case of a terrestrial planet or of helium and metallic hydrogen in the case of a Jovian planet. As the planet cools over geological time, a phase change is likely to occur, either to a new liquid phase immiscible with the first or to a solid phase. Assuming the adiabatic gradient to be smaller than the liquidus gradient, the phase change first occurs at the center of the sphere. It is known from metallurgy that the new phase which forms from a binary melt does not in general have the same composition as the liquid. To simplify the discussion we shall assume the melt to contain a larger fraction of the heavy component than does the eutectic. In this case the new phase has a larger fraction of the heavy component and hence is more dense than the initial liquid even in the absence of a density change associated with the phase change. The denser phase will accumulate at the center, forming an inner core. Associated with the growth of this inner core, there is a net motion of heavy material downward and light material upward, releasing gravitational potential energy. More specifically as the inner core grows by accretion of the denser phase, a residue of light material is left in the liquid near the inner core. Since diffusion of matter is ineffective over planetary length scales, this excess of light material must distribute itself throughout the outer core by means of convective motions driven by compositional

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bouyancy. The fact that the material diffusion coefficient is typically much smaller than the thermal diffusion coefficient accounts for the inherent efficiency of the gravitationally driven dynamo first pointed out by Braginsky (1963) and qualified by Gubbins (1977).

The power available from gravitational settling depends upon the density contrast $\Delta\rho$ between the two phases and may be crudely estimated by the formula

$$P = \Delta\rho V g L / t$$

where V is the inner core volume, L is the outer core radius g is a typical value of the local acceleration of gravity and t is the time which has elapsed since the inner core began to form. Typical estimates for the Earth's core give $P = 4 \times 10^{11} W$, which is sufficient to sustain a large field (see Loper, 1978a). In light of the uncertainty in composition and structure of the planetary cores, it is difficult at present to estimate the viability of gravitational settling as a power source for the other planetary dynamos.

If gravitationally driven convection occurs, the outer core is in effect mechanically mixed and Loper (1978b) has shown that several interesting thermal regimes are possible. For example, if the liquidus gradient is less than the conduction gradient (i.e., that gradient which would occur in the absence of motion), a slurry must occur in the fluid at the bottom of the outer core. A general thermodynamic theory for the motion of a binary-alloy fluid containing a slurry has been developed by Loper and Roberts (1978). Also, compositionally driven convection can occur even if the actual temperature gradient is less than the adiabatic gradient, making the fluid thermally stably stratified. In this case the fluid motions convect heat radially inward--up the temperature gradient. Hence the heat conducted down the adiabat cannot be taken as a lower bound on the heat transferred out of the core.

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"MACRODYNAMICS OF α DYNAMOS IN ROTATING STRATIFIED FLUIDS"

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Convective motions in the sun are presumed responsible for its magnetic fields, yet models for the solar stability require that these motions are confined to the outer sixth of its radius. In contrast, the magnetic fields and electric currents pervade the entire sun, and may be largest in the stable layer just beyond the lower boundary of the convection zone. Recent models for the stability distribution in the earth's core also suggest that the dynamo producing motions are confined to a small fraction of the electrically conducting volume. Theories for the geodynamo, e.g. Busse 1978 Moffatt 1978, deal with magnetic fields generated in and, in turn, limiting the fluid motions responsible for them. No current theory deals with the role of an adjacent stably stratified electrically conducting fluid as the significant amplitude determining factor in a dynamo. Three earlier papers Greenspan 1974, Malkus and Proctor 1975, Proctor 1977) do discuss the equilibration of global magnetic fields by the large scale fluid motions resulting from the Lorentz forces. When ohmic loss is the principal dissipative mechanism and inertial forces play a secondary role, the cylindrical integrals of the geostrophic components of the Lorentz force must vanish in a homogeneous fluid (Taylor 1963). For magnetic fields resulting from averaged small scale motions (the α -effect), the Taylor constraint leads to an unusual eigenvalue problem in which the determination of a finite amplitude velocity field precedes the determination of the eigensolution for the magnetic field B . However the Taylor constraint is both inappropriate in a stratified flow and suspect in any intense regions of generation such as solar convection. Here the following idealization is proposed: consider a thin spheroidal shell with intense fluid motions characterized by a local α ; determine the appropriate reformulation of the Taylor-like constraint for a rotating stable stratified fluid; formulate the eigenvalue problem for B (α) and for the two-dimensional velocity field (eigenflow) required by the Taylor-like constraint; formulate the finite amplitude problem, including the reduction in the static stability by the finite B fields and also by large thermal conduction. First conclusions are presented for the finite-amplitude eigenflow and the structure of axisymmetric initial B fields for particularly simple α . Eigenflows scale with the magnetic diffusion velo-

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city as anticipated from previous studies. The amplitude dependence of B on stability and conductivity is not yet determined.

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"TURBULENT CORE FLOW DUE TO PRECESSIONAL AND TIDAL FORCES" Willem V.R. Malkus, Dept. of Mathematics, M.I.T., Cambridge, Mass. 02139

The secular tidal deceleration of the earth is observed to be approximately three milliseconds per day per century^{1,2}. In consequence, the secular loss of rotational energy by the liquid core of the earth is greater than $5 \cdot 10^{18}$ ergs/sec.. Can the magneto-fluid dynamical tidal spin-down, or the interrelated precessional spin-over^{3,4}, of the core fluid transfer this rotational energy to the fields and currents of the geodynamo? Laboratory studies to be reported here suggest that either core spin-down or core spin-over separately could power the geodynamo if the flows are fully turbulent. How turbulent are the magneto-fluid dynamical processes in the earth's core? We may never establish the amplitude or time dependence of the small scale flows, however, the large scales exhibit considerable disorder. An example is the observed ten to thirty year random transfer of momentum between the core fluid and mantle. This process involves an exchange of energy with an r.m.s. value ten to one hundred times larger than the secular loss of rotational energy. Hence this turbulent exchange would have to be more than ninety percent (or ninety-nine percent) efficient to dissipate less than 10^{19} ergs/sec.. An electromagnetic core mantle coupling sufficient to explain these observed non-secular changes in the length of the day remains a matter of debate⁵, to be briefly assessed here. The central purpose of this study is the determination of turbulent onset and transport "efficiency" in flows induced by precession and tides in laboratory rotating spheroids. The external and fluid parameters determining the transition of these motions from laminar to turbulent are found for boundaries of varying roughness. The turbulent momentum transport is observed to relax the constraints on the flows imposed by rotation and the ellipsoidal cavity. In homogeneous or convectively unstable fluids the induced turbulent regimes occupy the entire spheroid. Initial experimental results on cylindrically stratified fluids suggest a host of new instabilities but reduction in the r.m.s. turbulent amplitudes. Over much of the parameter range estimated for the earth's core fluid there appears to be little doubt that both tidal and precessional forces could sustain turbulent flows akin to those observed in the laboratory. More complete dynamo theory, or an operating laboratory dynamo, may be needed to establish that these flows actually act as energy sources for the geodynamo.

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AN OUTLINE OF SOME OF THE PROPERTIES OF THE GEOMAGNETIC FIELD

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The magnetic field measured at the earth's surface originates from three sources: external sources that arise mostly from ionospheric currents, crustal remanent magnetization sources, and hydromagnetic sources in the earth's core. This talk deals with the latter, which can be well described outside the earth's core in terms of a magnetic scalar potential, ψ given by:

$$\psi = a \sum_{\ell=1}^{\infty} \sum_{m=0}^{\ell} \left(\frac{a}{r}\right)^{\ell+1} P_{\ell}^m(\cos\theta) (g_{\ell}^m \cos m\phi + h_{\ell}^m \sin m\phi)$$

where we use spherical coordinates (r, θ, ϕ) , a is the earth's mean radius (6371 km), $P_{\ell}^m(\cos\theta)$ are the Schmidt polynomials and g_{ℓ}^m and h_{ℓ}^m are referred to as the Gauss coefficients. The magnetic field can be obtained, by taking the negative gradient of ψ .

Primarily through the use of measurements from magnetic recording stations, one finds that approximately 90% of the present magnetic field at the earth's surface can be described by a geocentric dipole inclined $11\frac{1}{2}^{\circ}$ to the axis of rotation, while 10% of the field is described by higher order ($\ell > 1$) non-dipole field components.

The field constantly changes with time, with the dominant feature being a westward drift of the non-dipole field at approximately 0.2° per year (Bullard *et al.* 1950). It is clear that this westward drift does not account for all the secular change in the field, since zonal harmonics ($m=0$) also change significantly with time. A considerable amount of controversy exists concerning how to best describe the changes in the field that are not drifting westward. Part of the problem is certainly the fact that the historical record represents only a very small time window through which the magnetic field can be observed.

Palaeomagnetic studies allows one greatly to extend this window in time, but at the expense of considerable loss in resolution. Even when one confines work to rocks of the last 5 million years, for which tectonic problems are not too serious, one must deal with the fact that the palaeomagnetic record is discontinuous in both time and space.

In spite of these problems, a crude first order, approximation for the "mean palaeomagnetic field" can be obtained. To do this we exclude times when the field is undergoing transitions between normal and reversed polarity states or times when the field is undergoing large changes in direction that are not associated with polarity transitions (excursions). The primary reason why one can obtain a mean palaeomagnetic field is that, to a first order approximation, the time averaged field turns out to be axisymmetric, probably reflecting the averaging done by the westward drift of both the non-dipole and dipole fields. Because of this, one can average measurements from sites of a common latitude to surmount the problems that

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result from having incomplete time averages of the field at any one given site

The time averaged palaeomagnetic field obtained in this way in described by the zonal harmonics given in Table 1, as taken from Merrill and McElhinny (1977). This field exhibits north-south asymmetry. In addition, the time averaged normal polarity field is significantly different (at the 95% confidence level) from that of the reversed polarity field.

Phillips (1977) has used an entirely different approach that also leads to the conclusion that there are significant differences between the normal and reversed polarity states. Phillips has shown that both the lengths of normal and reversed polarity intervals can be modelled by a gamma distribution, but that the shape of this distribution is significantly different (at the 95% confidence level) for the two polarity states.

One can model the difference between the normal and reversed polarity states non-uniquely by a standing zonal magnetic field. One can then test the adequacy of such modelling by seeing if this standing field produces systematic virtual geomagnetic pole (VGP) paths during transitions between polarity states. Hoffman (1977) pioneered work on the systematics of VGP paths during transitions and showed that the observed systematics are consistent with kinematic dynamo models for polarity transitions developed by Parker (1969) and Levy (1972). Hoffman and Fuller (1978) have subsequently examined 8 quadrupole and octapole models that in a generalized sense are consistent with the Levy-Parker reversal mechanism. Of these, 2 models appear to be most consistent with the VGP transition data. The first is a Quadrupole model in which reversal occurs first in the southern hemisphere and the second is an Octapole model in which reversal occurs first at low latitudes. These are precisely the two models that are predicted from the use of the standing zonal field discussed earlier.

TERM	POLARITY		
	NORMAL	REVERSED	PRESENT FIELD
$\frac{0}{g_2}$			
$\frac{0}{g_1}$	+0.050	+0.083	+0.063
$\frac{0}{g_3}$			
$\frac{0}{g_1}$	-0.017	-0.034	-0.043

Table 1. Gauss coefficients for the time averaged palaeomagnetic field.

METEORITE MAGNETISM AND THE EARLY SOLAR SYSTEM
MAGNETIC FIELDS (A REVIEW): Takesi NAGATA, National
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Since the first systematic studies on meteorite magnetism by Stacey et al. (1961), a number of research works on the magnetic properties of meteorites and their implication for the early solar system magnetic fields have been reviewed, particularly by Brecher (1971, 1977) and Gus'kova (1972). The main problems in meteorite magnetism may be classified into (a) basic magnetic properties in relation to their chemical and petrographical compositions and (b) characteristics of natural remanent magnetization (NRM) of meteorites in association with the early solar system magnetic field.

(a) Basic magnetic properties of meteorites

Intrinsic magnetic properties of meteorites such as saturation magnetization (I_s) and magnetic transition temperature (H) represent contents of ferromagnetic constituents and their compositions. Corresponding to the Urey-Craig diagram, I_s values of ordinary chondrites and E-chondrites decrease in the order of sequence expressed by $E \rightarrow H \rightarrow L \rightarrow LL$, whereas I_s values of achondrites are extremely small ($I_s < 1$ emu/gm) because of a very small content of metallic phase. I_s values of C-chondrites are mostly due to spontaneous magnetization of magnetite or magnetite plus a small amount of native iron (e.g. Larson et al., 1974; Watson et al., 1975; Herndon et al., 1976).

The composition of ferromagnetic phase in meteorites is reasonably well represented by thermomagnetic characteristics which separately give magnetizations of thermally irreversible α -phase, a thermally reversible γ -phase and a thermally unstable $(\alpha + \gamma)$ -phase of Fe-Ni-Co-P alloy grains and thermally reversible magnetites. Corresponding to the Prior rule, ratio of saturation magnetization of α -phase component, $I_s(\alpha)$, to the total I_s value decreases in the order of sequence given by $E \rightarrow H \rightarrow L \rightarrow LL \rightarrow C$.

In Fig. 1, observed values of I_s and $I_s(\alpha)/I_s$ are plotted on an I_s vs $I_s(\alpha)/I_s$ diagram, which gives rise to a reasonably clear magnetic classification of stony meteorites corresponding to their chemical classification.

The structure-sensitive magnetic properties of meteorites such as magnetic susceptibility (χ_0), saturation remanence (I_R), coercive force (H_C) and remanence coercive force (H_{RC}) also have been examined. A significant characteristic of meteorite susceptibility is its considerably large anisotropy ($\chi_{max}/$

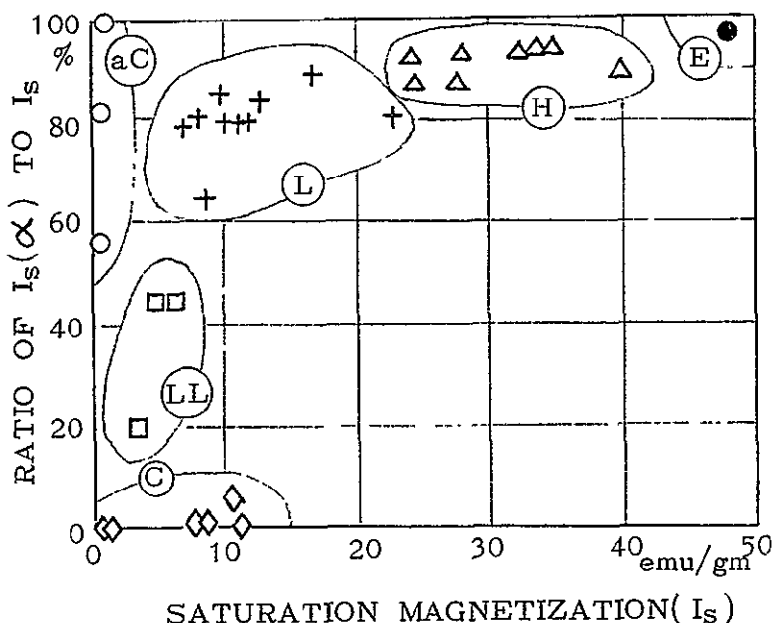


Fig. 1

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$\chi_{\min} = 1.5 \sim 2.0$), which suggests a possibility of their aggregation in the early solar system (e.g. Stacey et al., 1961; Brecher and Arrhenius, 1974).

(b) Natural remanent magnetization

Among various kinds of meteorites, C-chondrites generally maintain stable NRM (Banerjee and Hargraves, 1972; Butler, 1972; Larson et al., 1973; Brecher and Arrhenius, 1974). Since it is believed that C-chondrites were formed in the early stage of solar system formation, NRM of C-chondrites may represent the interplanetary solar wind magnetic field at the stage. The paleointensities (H_0) of C-chondrites derived from their NRM analysis are summarized in the following table, where T_0 denotes a critical temperature, during a cooling process from which the remanent magnetization was acquired.

Meteorite	H_0 (Oe)	T_0 ($^{\circ}\text{C}$)	Investigators
Orgueil (C ₁)	0.67	120	Banerjee & Hargraves
Renazzo (C ₂)	2.3	(250)	Brecher
Murray (C ₂)	0.7	(250)	Brecher
Murchison (C ₂)	{ 0.18	90	Banerjee & Hargraves
	{ 0.4 ~ 3.0	(250)	Brecher
	{ 1.09	130	Banerjee & Hargraves
Allende (C ₃)	{ 1.3	150	Butler
	{ 0.15 ~ 0.95	-	Brecher
	0.5	-	Nagata & Sugiura
Leoville (C ₃)	0.97	350	Nagata & Sugiura

Achondrites also maintain stable NRM in most cases. Since it is believed that achondrites represent the upper mantle part of meteorite parent planets, their paleointensity may suggest the magnetic field of parent planets. Paleointensities of Yamato-7307 (howardite) and Yamato-74013 (diogenite) are 0.13 and 0.093 Oe respectively.

NRM of E-, H-, L- and LL-chondrites are comparatively unstable in most cases. However, some of these chondrites maintain reasonably stable NRM, from which the paleointensity can be estimated. Estimated paleointensities are 0.10 ~ 0.68 Oe for L-chondrites, 0.15 ~ 0.39 Oe for H-chondrites and 0.33 Oe for an E-chondrite (Stacey et al., 1961; Gus'kova, 1963, 1970; Nagata and Sugiura, 1977). However, Brecher and Ranganayaki (1975) have stated that paleointensity of H-, L- and LL-chondrites is much smaller, ranging only from 0.01 to 0.08 Oe.

Iron meteorites also have stable NRM in most cases. It seems, however, that their stable NRM is due to spontaneous magnetization of $\alpha\{110\}$ plates nucleated and developed along the octahedral $\gamma\{111\}$ crystallographic plane (Brecher and Albright, 1977).

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THE ELECTRICAL CONDUCTIVITY AND EQUATION OF STATE OF SEVERAL CONSTITUENTS OF THE MAJOR AND OUTER PLANETS by William J. Nellis, Arthur C. Mitchell, Maxim I. Kovel*, Ronald S. Hawke and R. Norris Keeler, University of California, Lawrence Livermore Laboratory, Livermore, California 94550

The primary constituents of the major and outer planets are He, H₂, N₂, CH₄, NH₃ and H₂O. Equation of state measurements have been made on N₂ to 600 kb; CH₄ to 400 kb; NH₃ to 700 kb, and H₂O to 2.2 Mb. Electrical conductivity has been measured for H₂ at 2 Mb; NH₃ to 280 kb; and H₂O to 450 kb. None of these materials exhibited anomalous behavior within the limits of the measurements made with the exception of N₂, which shows behavior consistent with transition to an atomic (metallic) phase. The implication of these results is that the level of conductivity which exists in hydrogen above its transition pressure (2 Mb) does not exist in the most abundant constituents of the outer planets at core pressures; therefore, it may be anticipated that neither Uranus nor Neptune will have a magnetic field, unless it is associated with a small residual metallic core.

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THE MAGNETIC FIELD OF MERCURY-Norman F. Ness, Laboratory for
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Direct observations of the magnetic field of Mercury and its magnetosphere, formed by the interaction of the solar wind, were performed twice by the Mariner 10 spacecraft, in March of 1974 and again in March 1975. From these data, it is clear that there exists an intrinsic magnetic field of the planet, sufficiently strong at present to deflect the solar wind flow around the planet and to form a detached bow shock wave in the super Alfvénic solar wind. Three different methods have been used to analyze the magnetic field and derive quantitative values for the description of the planetary field:

- (1) Direct spherical harmonic analysis of the data,
- (2) Modeling of the magnetosphere by an image dipole and infinite current sheet in addition to the planetary field,
- (3) Scaling of a mathematical model for the terrestrial magnetosphere.

The results obtained yield dipole moments $\sqrt{(g_1^0)^2 + (h_1^1)^2 + (g_1^1)^2}$ ranging from 2.4 to 5.1×10^{22} Gauss cm^3 , with the lower values associated with certain models using partial quadrupole (g_2^0) and octupole (g_3^0) terms to improve the least squares fitting of models to observations. Because the data set is incomplete, in the mathematical sense, no unique representation of the planetary field multipolar representation can be derived by method (1). The use of only 1 of the 5 quadrupole moment terms and 1 of the 8 octupole moment terms corresponds to a displacement of the dipole along its axis. These terms, used in methods (2) and (3), yield equivalent offsets of the dipole by approximately $0.2 R_M$. The selection of only those higher order terms having axial symmetry ($m=0$) has not been justified on any physical grounds. The fields of Earth and Jupiter do not show axial symmetry. Such a large offset may reflect the limitations of the models used to represent the external current systems. Because of the relatively short radial excursion of the data, the g_2^0 and g_3^0 terms may also be spatially aliased with the g_1^0 term.

Analyses by method (1) of subsets of data from the third encounter, taken near closest approach, yield a convergent series of dipole moment values which are believed to best represent the intrinsic planetary field. These provide a mean moment of $330(+18) \gamma R_M^3 = 4.8 \times 10^{22}$ Gauss cm^3 at a tilt angle of $14^\circ \pm 5$ and at longitude of $148^\circ \pm 21^\circ$. This means that the surface field at Mercury is about 1% of Earth's, while the moment is 6×10^{-4} of Earth's.

The origin of the field cannot be uniquely determined. It may be due to an active dynamo, a remanent magnetic field or a combination. Consideration of remanence as the source leads to some difficulties although definitive knowledge of the planetary interior structure and thermal state is lacking sufficient to categorically eliminate this source. Some success in attempting to explain the field as due to an active dynamo has encouraged these efforts. This paper will critically review the quantitative analyses of the Hermean field and discuss its origin.

TABLE I

MERCURY'S MAGNETIC MULTIPOLE MOMENTS*

Source	Data	g_1^0	g_2^0	g_3^0	External Terms	Offset	Tilt**
Ness et al (1974a)	I	227	0	0	0	$0.47 R_M$	30_E°
Ness et al (1975)	I	350	0	0	n=2	0	10_E°
Ness et al (1976)	III	342 ± 15	0	0	n=1	0	$11^\circ \pm 1_0$
Whang (1977)	I & III	266 165 166	0 117 75	0 0 48	2-Dimensional Tail sheet + image dipole	0 0 0	2.3_0°
Jackson and Beard (1977)	I & III	257 170	0 114	0 0	Scaled terrestrial analogue	0 0	$10^\circ - 17_0^\circ$
Ng and Beard (1978)	I & III	190	0	0	Scaled terrestrial analogue	(0.033, 0.026, 0.189)	1.2_0°
Ness (1978)	III	330 ± 18	0	0	n=0 and 1		$14^\circ \pm 5_0$

*Using dipole aligned coordinates. Polarity sense is same as at Earth.

**Relative to normal to ecliptic or orbital plane, indicated by subscript E or 0.

THE METALLIC IRON IN LUNAR SOILS AND BRECCIAS, G.W. Pearce,
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It has for some time been known that the characteristic materials of the lunar regoliths -- soils and soil breccias -- differ markedly from local rocks in the amount and grain size of constituent metallic iron, the only important magnetic mineral of the lunar surface. Thus at mare sites the regolith material is enriched by factors in the order of 10 in metallic iron over the local basaltic rocks (1, 2, 3). Much of the excess iron metal is very fine grained (dimensions less than 300 Å) as determined magnetically. Mare soils are quite similar to each other both in the quantity and grain sizes of their metallic iron components. At highland sites the local rocks have perhaps, on the average, more metallic iron than the regolith constituents (4). However, as with the mare regoliths the metal in the highland soils and soil breccias is generally fine grained whereas highland crystalline rocks contain mostly coarse iron (dimensions greater than 300 Å) (4). The highland soils are also quite similar in most cases to each other in their metal contents.

The highland and mare soils appear to differ in how they obtained the metallic iron they possess. Using a combination of magnetic analysis and trace element analysis, Chou and Pearce (5) were able to determine the relative contributions of iron from external sources such as meteorites and of iron produced by reduction of indigenous silicates and oxides. Thus highland soils appear to have obtained almost all metal from meteorites since they contain the proper meteoritic abundances of nickel and other trace elements. Mare soils, on the other hand, have a considerable contribution (approximately one half) from reduced iron.

In a detailed analysis of grain size separates of an Apollo 15 core (6), it was found that the sub-45 µm fractions are always considerably richer in metallic iron than fractions of larger grain size. The concentration of metallic iron in these sub-45 µm fractions showed some variation as a function of their depth below the regolith surface, but showed remarkable similarity in these magnetic properties (such as coercivity and the ratio of saturation remanence to saturation magnetization) which are sensitive to the grain size distribution of the iron particles themselves. This thus suggests that there was a common source during the period of time represented by the core or that there was a process acting on the metal modifying its grain size towards a constant distribution. The similarity of mare soils to each other and of highland soils to each other further suggests that this common source or equilibrating process may be widespread geographically as well as temporally.

The major portion of the external source of iron metal is the rain of vaporized and melted material arising from meteorite, micrometeorite and solar wind particles that hit the lunar surface. Most of the metal in the rain would be due to vaporization and melting of meteoritic metal although a small component due to vaporization and melting of indigenous metal would be present. The rain would consist of the vaporized and melted material returning to the lunar surface as solid particles of small but variable size. Meteoritic iron which remained solid during impact and fallback must be a small component of the rain since the iron metal in meteorites is generally much coarser than that found in lunar soils. Vaporized metal would now be found on the surfaces of small soil particles whereas melted metal might sometimes be found inside silicate particles which had simultaneously been melted and ejected. If the distribution of impacting particles is constant

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in time and space then the grain size spectrum of the metal in the fallback might be expected to be constant from place to place and from time to time. Thus there is a possible common source for iron in highland sites.

In the case of mare sites the indigenous metal must be considered. Reduction of iron from the ferrous state on the lunar surface must be accomplished through the agents of heat from impacts and reducing gasses such as H₂ and CO implanted into regolith particles from the solar wind. To get the very fine grained metal particles observed in soils, the reduction must take place under constrained conditions of temperature and heating time (7), since at high temperatures (above 600-700°C) grain coarsening of the iron particles occurs through diffusion (8).

Whether the iron metal results from meteoritic addition or reduction of indigenous material the latter process of grain coarsening will continue to occur as long as impacts produce heating of the regolith. Thus to get the uniform distribution of grain sizes of iron particles observed or expected, an equilibrium of between the processes of addition of very small particles and the grain growth of these particles would have to exist.

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Magnetic Properties of Asteroids

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Abstract

The aim of this research is to develop a first order set of magnetic models of asteroids based on the magnetic properties of meteorites determined from laboratory analysis. Although asteroids have been suspected as the source of meteorites by many investigators (Opik, 1963; Anders, 1964; Arnold, 1965; Wetherill and Williams, 1968), the association of asteroid mineralogy with certain corresponding meteorite types has been accomplished through telescope spectral data (Mc Cord et. al., 1970; Mc Cord and Gaffey, 1974). Further studies of the reflectance spectra of asteroids and meteorite optical properties has indicated that some meteorite types have reflectance curves which may be interpreted in terms of the type, composition, relative abundance and distribution of the constituent mineral phases. Such matching of meteorite and asteroid surface reflectivity data suggests the following associations: Vesta-basaltic achondrite; Ceres-carbonaceous chondrite; Toro-chondrite; and Juno-iron with a silicate admixture. We assume, for the first order models, that the surface material of these asteroids is representative of their internal constitution.

We consider asteroids composed of one or more layers of meteoritic material which possesses its own magnetic moment and which is subjected to an external (solar) field. The total magnetic potential, Φ , consists of a natural magnetic remanence term, Φ_r , and an induction term, Φ_i , such that $\Phi = \Phi_r + \Phi_i$. Laplace's equation is solved in the region where $\underline{B} = \mu \underline{H}$ and where there are no currents. In this region the magnetic field $\underline{H}$ and induction $\underline{B}$ are derived from the (scalar) magnetic potential. In regions where ferromagnetic behavior is exhibited, the permeability, μ , and magnetic induction become so large as to permit simplifying assumptions. Our approach does not take into account any details of the thermal remanent magnetism acquired by a body cooling through the Curie point, but rather uses the measured values of the natural magnetic remanence and so the problem is simplified.

We use the laboratory results of Guskova (1972) and Brecher (1975, 1977) who have documented the magnetic properties of numerous stony and iron meteorites. Associations between the magnetic properties of several meteorite groups has been established by Wassilewski (1974).

Results based on remanence data indicate homogeneously layered stony asteroids possess surface fields from 12 - 2400 gamma at the poles

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6 - 1200 gamma at the equator. Asteroid models composed of iron meteoritic material give unrealistically high values starting from 6000 gamma to well above 50,000 gamma at the equator, and double at the poles. In addition, the inductance field is vectorially added to the remanence field.

There are several criticisms of this model: the magnetic remanence is assumed to have a single orientation throughout the body, the meteorites past history affects the remanence properties, the asteroid is probably not uniformly magnetized -- we may be only observing (asteroid) crustal magnetic remanence, and the overall population of meteorite classes is not the same as that of observed asteroid spectral data.

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SOME COMMENTS ON METEORITE MAGNETISM: M.W.Rowe and M.Hyman, Dept. of Chem., Texas A&M Univ., College Station, TX 77843.

Primordial Magnetic Field- During the past decade, the consensus of opinion for the origin of the magnetization of stone meteorites has shifted from fields associated with meteorite bodies to fields associated with the solar nebula during the early period around the formation of the solar system (e.g., Stacey and Banerjee, 1974). Evidence toward this conclusion rests primarily on magnetic studies of carbonaceous chondrites (Banerjee and Hargraves, 1972; Brecher, 1972; Butler, 1972; Brecher and Arrhenius, 1974). Geochemical evidence from carbonaceous chondrites indicates that they are primary condensates which accreted at less than 200°C and have not been heated significantly since accretion (Anders, 1971). Thermomagnetic analyses on all known carbonaceous chondrites at the time (Larson, et al., 1974; Watson, et al., 1975; Herndon, et al., 1976) indicated that some of the carbonaceous chondrites are more suitable for paleointensity determination than others. All C1 chondrites contain only magnetite as a magnetic constituent and their magnetic analysis may be more straightforward than the carbonaceous chondrites studied to date, which contained troilite. Troilite, under the conditions generally used for paleointensity estimates, seriously complicates the analysis. Other C2 and C3 chondrites contain magnetite only, whereas some others contain primarily magnetite and iron. This information should be useful in designing magnetic experiments. Unpublished thermomagnetic analysis of some ureilites are presented here as well. Studies on carbonaceous chondrites, especially those on carbonaceous chondrite chondrules (Lanoix, et al., 1978) yield information about the fields associated with the solar nebula at very early times in the history, i.e., about the first 10 million years (Sonnet, 1978). Estimates of the intensity of this solar nebula field range for ~ 1 - 16 Oe.

Later Magnetic Fields- Rowe (1974 a,b) suggested that in some heavily shocked meteorites sufficient heating was experienced relatively recently in solar system history to have erased all evidence of any magnetization acquired prior to the collision. Careful magnetic studies on such meteorites may lead to interesting information about magnetic fields of more recent origin.

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Bogard et al., (1976) have shown by ^{40}Ar - ^{39}Ar dating that at least nine chondrites show ages of $< 0.6 \times 10^9$ years which resulted in 89-98% argon loss. Rowe (1974 a,b) previously argued that such loss probably requires temperatures in excess of any expected Curie point involved. Of particular interest is the Ucera chondrite which Bogard, et al., found to have a ^{40}Ar - ^{39}Ar age consistent with its cosmic ray exposure age. Magnetic studies, if properly conducted (Brecher and Ranganayaki (1975), on Ucera and other shocked meteorites should yield considerable information on the duration of the magnetic field(s) which have been recorded by meteorites which have undergone severe shock.

Magnetic and Troilite Contents of Selected Carbonaceous Chondrites- Hyman, et al., (1978) have conducted large numbers of measurements on the magnetite contents of the C1 chondrites. Those measurements, along with a few troilite estimates, will be presented here.

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THE ANCIENT LUNAR DYNAMO; S.K.Runcorn, Institute of Lunar & Planetary Sciences, University of Newcastle upon Tyne, England.

The discovery of lunar palaeomagnetism was a surprise because (a) Luna II failed to detect a general lunar magnetic field, (b) the mean density of the Moon (3.34) was rather generally held to exclude the existence of an iron core, and (c) the absence of indications of tectonic movements, which had been generally interpreted by a Moon with a cold accretional origin, a minimal later differentiation or no internal dynamical history. Some carry over of these early ideas has contributed to scepticism concerning the interpretation of the facts of lunar palaeomagnetism. As in the early development of terrestrial palaeomagnetism, and indeed of any other historical process, hypothetical ideas based on limited observations to explain the origin of the phenomena are hard to test or to exclude with finality even if physically implausible and the correct interpretation is that which explains best the sum total of the evidence. Just as terrestrial palaeomagnetism was seen to contain a record of the variations in the geomagnetic field, so we think that the remanent magnetization of rocks records the presence in the Moon's early history of a magnetic field of internal origin.

Stable remanent magnetization ranging from 10^{-4} emu/gm to 10^{-6} emu/gm has been found in crystalline rocks and high grade breccias from all the Apollo missions, the magnetization being carried by multidomain and single domain iron grains. From this random sampling of the Moon, widespread general magnetization of its crust must be inferred and fits rather well the discoveries of magnetic anomalies from Explorer 35 studies of the solar wind, by the magnetometers landed on the surface during the Apollo 12 and 15 missions and the surveys made by the astronauts during Apollo 14 and 16 missions, the magnetic surveys made from the sub-satellites of Apollo 15 and 16 and the magnetic anomalies inferred by the reflected electron technique. An essential idea in the interpretation of these anomalies is that their magnitude and scale, 1-100 γ and tens of kilometers are consistent with an origin in edges, such as are produced by cratering, in horizontal strata magnetized with intensities of those of the returned samples.

The Moon's present dipole moment is extremely small (the surface field is 0.05 γ). This is strong evidence against a uniform magnetized crust of more than negligible thickness: thus this null result excludes magnetization in a roughly uniform external field such as might be produced by a close rapid approach to the Earth or some primeval magnetic field perpendicular to the ecliptic. The crustal magnetization of the Moon is, therefore, either random or is the result of magnetization by a field of internal origin, for it has been proved that if a spherical shell became permanently magnetized in the same direction and proportional to the intensity of any field of internal origin, this magnetizing field later disappearing, then, induced magnetization effects being neglected, there is no field whatsoever outside it. On this view the lunar anomalies arise from small scale edge effects or demagnetization by impact. Recent analyses of the limited region of the Moon where magnetic anomalies have been carefully determined support the hypothesis of magnetization by an internal field rather than the hypothesis of random magnetization such as would result from impact magnetization.

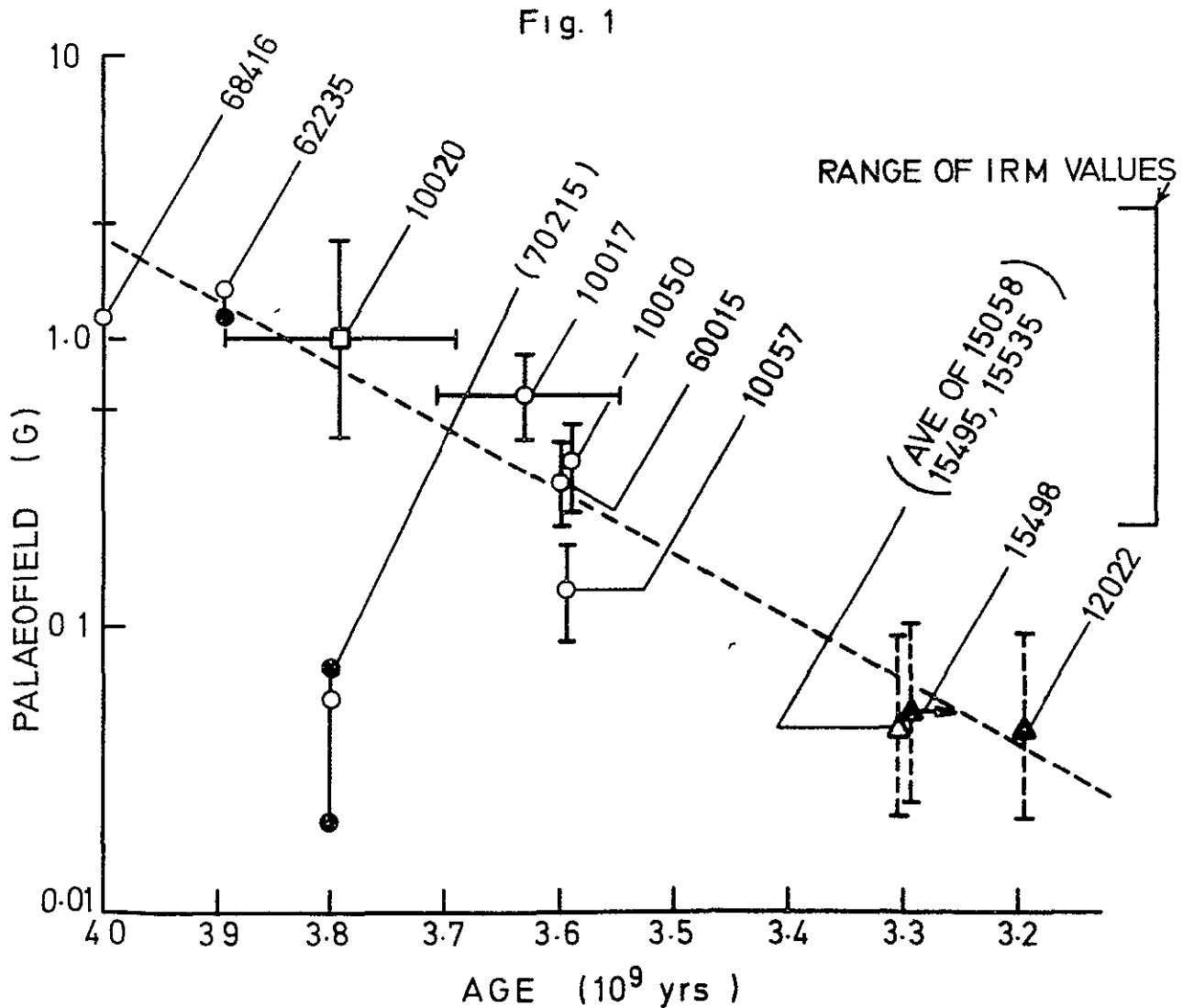
Palaeointensity measurements have been made with apparent success on a number of Apollo rocks but in other cases inconsistent results have been obtained which suggest that either secondary magnetization or chemical changes

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during experimentation prevent the original magnetic field intensity from being determined. Fig. 1 shows results by the Thellier-Thellier method and the ARM method which appear to be reliable and the range of those found by the IRM normalization method. Both the high fields of around 1 gauss for $4.0 \cdot 10^9$ yrs ago and the suggested exponential decrease with time are significant. This later trend, if substantiated, is extremely hard to explain on the impact hypothesis. The high field values also exclude a primeval permanent magnetization of the Moon as being the source of the magnetizing field. The hypothesis of a core dynamo is, therefore, the most adequate explanation of the whole lunar palaeomagnetism data.

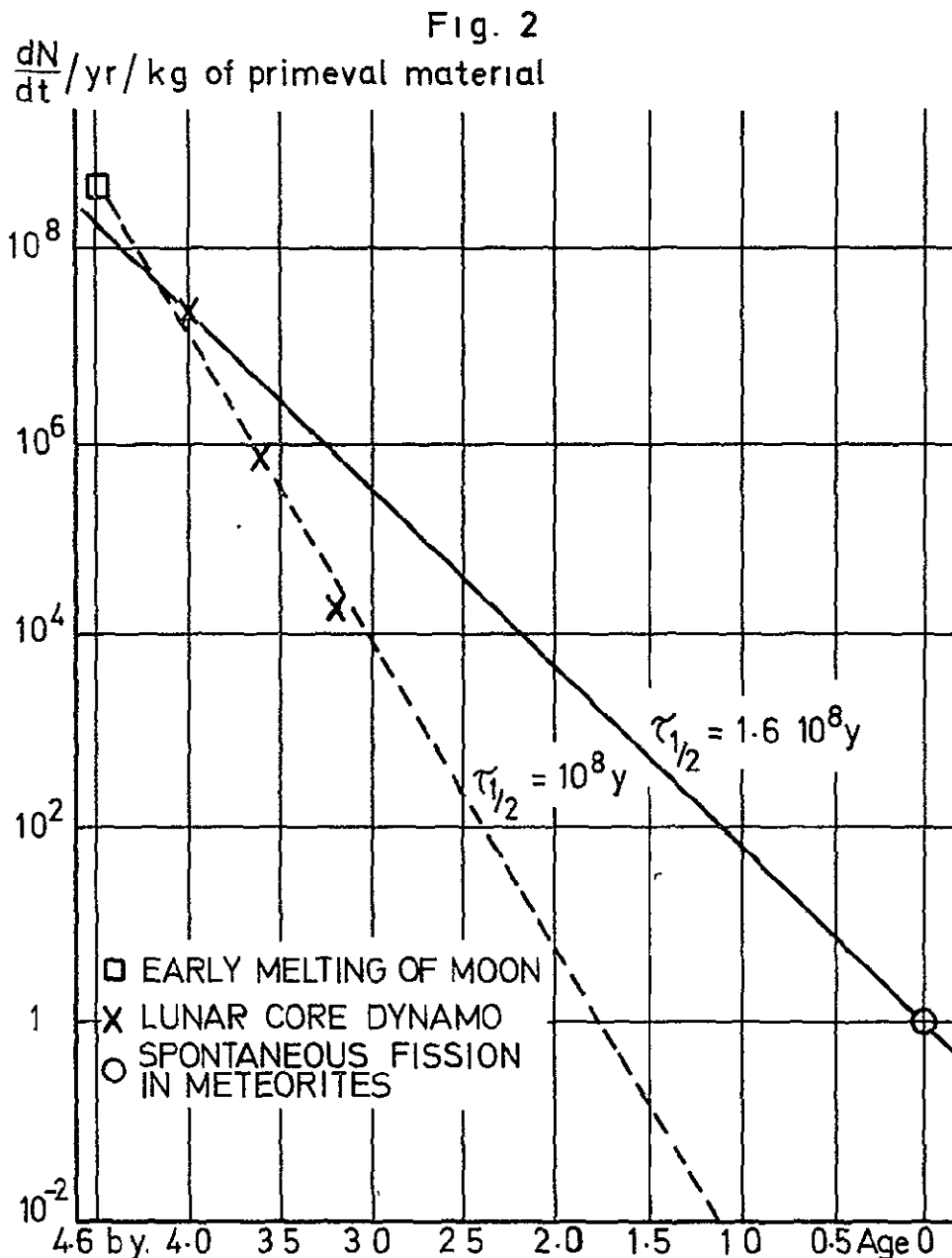
The evidence for a present lunar core is not conclusive but comes from three lines of evidence: A delay in the P-wave travel time passing through the core from a meteorite impact on the far side, electrical conductivity measurements indicating a possible high conductivity region in the deep interior and recent determinations of the moment of inertia factor as 0.391 ± 0.002 . It can be shown, from the absence of changes in the dimensions of the Moon since about $4.0 \cdot 10^9$ years, that the present core must have formed



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prior to this time. The formation of a core in the very early history of the Moon requires a powerful heat source and the melting of the Moon to the centre can not be done by accretion or eddy current heating. The hypothesis has been made that superheavy elements are responsible for the complete melting of the Moon. It is likely that these are siderophile and, therefore, could provide the heat source which generates the ancient lunar magnetic field. Thermodynamic arguments enable a minimum value of the heat source in a core of 500 km in radius necessary to generate the fields observed is too high (at $4 \cdot 10^9$ yrs the heat source must be 10^{12} W) to be generated by the conventional methods suggested for the terrestrial dynamo. Fig. 2 shows the rate of fission of SHE nuclei needed for different periods and the results give a half life of 10^8 years (dotted line).



THE MARTIAN MAGNETIC FIELD, C.T. Russell, Institute of Geophysics and Planetary Physics, University of California, Los Angeles, California 90024

The first spacecraft to Mars, Mariner 4 in July 1965, carried a magnetometer as well as other particles and fields instrumentation that would have detected a Martian magnetosphere had it encountered one. However, at its rather distant 'closest' approach of 3.9 Rm, it detected no magnetosphere. Mariner 4 did encounter a bow shock similar to the earth's¹. This observation led to some controversy over the existence of a planetary field due to the uncertainties in determining the size of the obstacle needed to cause a shock at the observed position^{2,3}.

The next set of observations came with the successful orbiting of Mars by the Soviet Mars 2 and 3 spacecraft in November and December 1971⁴ and later by the Mars 5 spacecraft in February 1974⁵. These spacecraft had periapsides of 1100 km for Mars 2 and 3 and 1800 km for Mars 5. From the Mars 2 and 3 data, Dolginov and colleagues immediately concluded that Mars had an intrinsic magnetic field⁴. They confirmed this conclusion with the Mars 5 data⁵. The identification of a planetary field by Dolginov and colleagues is essentially based on three different types of observations: the location of the bow shock crossings identified by a variety of instruments; a direct penetration (or two) into the dayside magnetosphere; and the constancy of the direction of the magnetic field in the wake region of Mars. Any one of these observations would be very good evidence for an intrinsic planetary field, if it were unambiguous. However, all three observations have been called into question.

When the totality of bow shock observations are considered, and the terrestrial scaling between magnetopause and shock location is used to scale between Martian shock and obstacle height, the effective obstacle height is in the Martian ionosphere^{6,7}. Some of the distant "shock" encounters may have been confused with upstream effects on the solar wind^{8,9}.

If the scaled magnetopause height is in the ionosphere at a height of 300 km then it is very improbable that Mars 2 and 3 entered the magnetosphere proper above the dayside of the planet since their closest approach to the planet was 1100 and the claimed encounters were even more distant than this. In fact, if the two reputed distant shocks were upstream effects, then the magnetopause encounters were in fact simply shock crossings^{8,9}. Further, the field in the putative magnetosphere had the characteristics of shocked solar wind field lines draped around the planet⁸.

Finally, the reputed Martian magnetotail observations resemble neither those of the Earth nor of Venus. The magnetosheath field points directly into the wake of the planet as if there were no obstacle to such penetration as a magnetotail would provide¹⁰. The field in the wake region is claimed by Dolginov et al. to be independent of the radial component of the interplanetary field. However, this is not the component of the solar wind magnetic field to which induction effects are sensitive. In fact all tail observations are consistent with induction effects¹⁰.

The estimates of the Martian magnetic moment are all much less than one would expect if Mars had an active magnetic dynamo¹¹. In the absence of any theory one would expect since the earth and Mars have similar rotation rates and since Mercury and Mars are expected to have similar core sizes, the Martian moment should be intermediate between those of the Earth and Mercury, i.e., between 8×10^{25} and 3×10^{22} Gauss-cm³. If the dependence on rotation rate were linear we would expect the Martian moment, as scaled from Mercury to be about 3×10^{24} Gauss-cm³. This estimate is far larger than that reported by Dolginov et al.^{4,5} of 2.5×10^{22} Gauss-cm³, or the upper limit based on

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rejecting the reported Mars-5 magnetotail encounters¹⁰ of 2.1×10^{21} Gauss-cm³.

If we use the similarity law for planetary dynamos derived by Busse¹² and assume a core of about 1750 km¹³ and that internal core conditions are similar, we obtain a predicted moment of about 5×10^{24} Gauss-cm³ a factor of 200 greater than Dolginov et al.'s estimate and 2000 greater than our estimated upper limit. We note that Busse's model when applied to Venus in a similar way is only about a factor of 3-4 too high. Another way to approach this calculation is to use the similarity law to calculate the dynamo radius. Using Dolginov et al.'s estimate we obtain a radius of slightly under 500 km. Using our best guess upper limit we obtain a maximum dynamo radius of 250 km. Thus, at most only a small fraction of the Martian core, as determined from gravity studies, can be undergoing dynamo action today if Busse's model of the geodynamo is correct, or applicable to Mars. Even without the guidance of theory, the possible strength of a Martian dynamo is far below expectations. Hence, it is most likely, that if Mars ever had an active dynamo, it is not operative today.

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FREEZING THE CORES OF TERRESTRIAL PLANETS BY SUBSOLIDUS WHOLE MANTLE CONVECTION. G. Schubert, Dept. of Earth & Space Sciences, Univ. Calif., Los Angeles, Calif., 90024, P. Cassen and R. E. Young, NASA Ames Research Center, Moffett Field, Calif. 94035.

It is all too easy for subsolidus whole mantle convection to freeze the core of a terrestrial planet thereby destroying its magnetic field. This was first demonstrated quantitatively by Schubert and Young (1976) who calculated temperatures in a convecting, constant viscosity, internally heated fluid model of the Earth's mantle. They showed that the temperature at the core-mantle boundary would lie significantly below the iron melting point if the mantle viscosity were less than $10^{24} \text{cm}^2/\text{s}$. Cassen et al. (1976) showed that convection in the relatively thin mantle of Mercury could freeze its core in a billion years or less.

Here we report additional computations using the Earth model of Schubert and Young (1976) which reveal that the low core-mantle boundary temperatures are established by the convecting mantle on a time scale less than the age of the Earth. If the viscosity of the Earth's mantle is indeed nearly uniform with the value of $10^{22} \text{cm}^2/\text{s}$, as inferred from glacial rebound data (Cathles, 1975; Peltier, 1976), how could the outer core still be liquid? The existence of a liquid outer core in the Earth places a significant constraint on the efficacy of subsolidus convective cooling during the thermal history of our planet. While we must find a reason why overly efficient mantle convection has not frozen the Earth's core, this problem may not exist for one or more of the other terrestrial planets if future seismic observations should reveal a solid core or if future magnetic observations should confirm the absence of a planetary field.

For the Earth, there are several ways in which core solidification by subsolidus convective cooling can be prevented. One way is to have a significant source of radioactive heating in the core. Another way is to prevent convection from reaching the lower mantle for a portion of the Earth's history, particularly during the initial period of cooling after core formation when convection should be especially vigorous. This may be accomplished by chemically or viscously stratifying the lower mantle or by hypothesizing that the lower mantle geotherm is subadiabatic. In view of both the inference from glacial rebound data that mantle viscosity is essentially uniform and the argument of Sammis et al. (1977) against large viscosity jumps across the major mantle phase transitions, a subadiabatic lower mantle accessible only by some form of weak penetrative convection may be the more likely explanation. The models of Sharpe and Peltier (1978) rely on an assumed subadiabaticity of the lower mantle to prevent core solidification by solid state convective cooling.

In addition to the numerical results mentioned previously, we present a simple model of core cooling by subsolidus convection which has an analytic solution and which is therefore particularly well-suited for illustrating the efficiency of the solid-state cooling mechanism. The model assumes a vigorously convecting, essentially isothermal mantle with thermal boundary layers at the core-mantle and lithosphere-mantle interfaces. The heat flow across each of these boundary layers is by conduction; accordingly, the heat flux is directly proportional to the temperature difference across the boundary layer and inversely proportional to its thickness. The thermal boundary layer thickness is in turn assumed to be proportional to a fractional power (about $1/3$) of the Rayleigh number of the mantle. For constant temperatures at the upper and lower boundaries of the mantle, a simple energy balance provides a differential equation for the cooling history of the mantle. The model in-

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incorporates a temperature-dependent mantle viscosity which has a controlling influence on the thermal evolution. The analytic solutions clearly show that subsolidus convection can remove all the latent heat of solidification from the core of a terrestrial planet on time scales short compared with geologic time.

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A THERMAL HISTORY MODEL FOR THE TERRESTRIAL PLANETS WITH PARAMETERIZED CONVECTION AND IMPLICATIONS FOR PLANETARY MAGNETISM

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With satellite data now available for the magnetic field intensities of all the terrestrial planets, it is of interest to consider this data from the viewpoint of the thermal histories of the planets. One naturally expects a relationship between the nature, and even the presence of an externally observable magnetic field and the thermal state of the interior. The strong, mainly dipolar geomagnetic field is believed to be a consequence of fluid motions in the outer, electrically conducting iron core. On the other hand, if strong magnetic fields were present in the early solar system, then 'cold' planets could have been magnetized, with the result that if portions of these planets have never warmed beyond the Curie Point, a present day remanent magnetization may be possible. In this paper we propose to study how the thermal histories of the terrestrial planets may 'set the stage' for determining the nature of their magnetic fields.

Traditionally, thermal history models for the planets have neglected the convective transport of heat. The resultant 'diffusion' models were characterized by very long thermal time constants and generally proposed conduction profiles which were highly unstable to convection. Hence these models should not be physically realized. In the present study we have developed a technique which parameterizes the energetically important aspects of convection without requiring an explicit solution for the dynamics. The scheme accounts for the enhancement of the radial flow of energy by modulating the thermal conductivity in the diffusion equation with an estimation for the Nusselt number, and is shown to be valid provided viscous dissipation is negligible. This latter requirement is satisfied for Mars and Mercury, but may be only marginally valid for Venus and the Earth, depending on the extent of convection. A power law dependence of the Nusselt number upon the Rayleigh number, as deduced from theory and experiment, is employed which maximizes the assumed efficiency of convection.

To commence the thermal history calculations it is necessary to assume a specific model for planetary formation. Assuming the moon accreted homogeneously, in an isolated environment, and interpreting the lunar crust as a product of late-stage accretional melting, it is possible via a simple energy balance argument to compute the minimum conversion efficiency and corresponding rate of accretion required to initiate melting of the incoming material. For the model which we have employed for the accretion dynamics, this provides sufficient information to permit the characteristics of the accreting 'cloud' to be found. Assuming the terrestrial planets formed from 'clouds' of similar characteristics it is then possible to construct the initial thermal states of these planets. Such calculations show that virtually all of Venus and the Earth would pass through an initially molten stage, while only the outer half of Mars and the outer third of Mercury would melt. Additional modifications to the initial profile due to adiabatic compression, the heat released during iron-silicate separation, and finally the action of fluid convection are also considered.

It is a characteristic feature of the formation models for Venus and the Earth that their interiors are totally depleted in radioactive heat sources by the end of accretion, all of these sources being concentrated in an outer 'crustal' layer. This removes an important but poorly constrained parameter. While we do not wish to suggest the interiors actually are depleted in heat sources, we may consider that for Venus and the Earth at least, we are studying

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an important end member in a class of possible thermal history models - namely the consequences of 'primordial' heat associated solely with the formation process itself.

Earth : The initial thermal structure of the Earth is essentially identical to the present day state, so that the study of the Earth's thermal evolution reduces to a consideration of the conditions under which it is possible to preserve the initial structure. This in turn leads to the identification of several 'critical' physical parameters. The most important parameter is the mean mantle viscosity, which controls the effectiveness of convection in cooling the interior and in particular, the molten outer iron core. Assuming that the mantle deforms as a power law fluid, it is possible to find a class of models which preserves the dominant features of the initial state until today and which, in addition, satisfies the mean mantle viscosity as inferred from the glacial rebound data (10^{22} poise), and the observed mean surface heat flux (60 mW/m^2). It is interesting to note that the corresponding Rayleigh number of 10^7 results in an effective thermal time constant for the entire mantle which is actually less than that for the moon, considering heat transport in the moon as purely diffusive.

Venus : Applying the identical thermal model to Venus except for specifying a surface temperature of 750°K instead of 300°K , we found that the higher mean mantle temperatures resulted in an enhanced cooling of the interior due to correspondingly lower viscosities. The model suggests that if the presently observed high surface temperature of Venus has persisted through most of its history, the enhanced vigour of mantle convection would have resulted in core solidification about 2×10^9 years ago. At present the Venusian geotherm would be colder than that for the Earth and the corresponding mean surface heat flux would also be diminished by about 30%. This result could explain the observation that while the Earth and Venus have similar physical properties, the magnetic field of Venus is nevertheless very weak.

Mars : The thermal evolution of Mars follows a significantly different course than that of the larger terrestrial planets. Shortly after accretion the small, molten outer iron core solidifies. The next 2×10^9 years are relatively quiescent, characterized by a general warming of the 'primitive' inner half of the planet. At about 2×10^9 years ago this 'primitive' core would have overturned, melted and differentiated following the onset of solid-state convection and an assumed subsequent Rayleigh-Taylor instability. The significant heat pulse associated with the formation of the molten iron core would have lead to the resumption of mantle convection, although it is unlikely that a form of plate tectonics could have ensued. Nevertheless, the surface would have undergone significant tensional deformation at this time. The model suggests that core solidification should again have occurred about $.5 \times 10^9$ years ago and that at present the mantle may be only weakly convecting. These results are in qualitative agreement with the observation that the Martian magnetic field is extremely weak.

Mercury : While the early thermal history of Mercury is similar to that of Mars, the assumed absence of potassium, in accordance with proposed cosmochemical models, prevents the interior from warming up to where the onset of convection would occur. Consequently the inner 2/3 of Mercury would remain as 'primitive', undifferentiated material until the present day. In this sense its thermal history would most closely resemble that of the moon. This raises the interesting question as to the nature of the observed Mercurian magnetic field. We propose that the field is a fossil remanence of an early, strong solar system magnetic field, and is carried in the outer 300km of the planet. The required intensity of magnetization is such that the remanence

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could easily be carried by material with a composition similar to that which has been observed to carry the lunar remanence.

In this study we have attempted to understand how the thermal history of a planet may determine the nature and even the existence of its magnetic field. While the size of the planet is the single most important parameter, there are several other parameters which also enter the evolutionary study in an important way. These include the surface temperature, the concentration and distribution of radioactive heat sources and the presence and vigour of convection. This is only a preliminary investigation and rests on several ad hoc assumptions the most important of which surround the formation model. However because of the dominating influence which solid-state convection seems to have had on the thermal histories of Mars, Venus and the Earth, we feel that the present study is a significant step forward in understanding the energetics of planetary evolution.

ON THE DETERMINATION OF THE HERMAEAN MAGNETIC MOMENT.

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Mariner 10 passed through the near tail portion of Mercury's magnetosphere on March 29, 1974 and March 16, 1975. In addition to these encounters, termed MI and MIII respectively, a second pass on September 21, 1974 did not bring the spacecraft near enough to observe either the magnetosphere or bow shock. Subsequent studies (7,8,12,13,14,15) have found that magnetic field and particle observations to be consistent with Mercury possessing a permanent terrestrial-type magnetosphere as modified by both this planet's proximity to the sun and large size relative to its magnetospheric cavity. These two factors result in the magnetospheric current systems being nearly tantamount to the planetary magnetic moment as source of the observed magnetic field during part of the Mariner 10 encounters.

As noted by Ness (8,10) a unique determination of the intrinsic magnetic moment by means of spherical harmonic representations of both the internal and external scalar magnetic potentials would require the determination of two or more components of the field on a surface enclosing the planet. However, the Mariner 10 mission returned only ~30 min of magnetic field data from the night-side magnetosphere so that it becomes necessary in modeling the observations to make assumptions regarding the position and intensity of both the magnetopause and neutral sheet currents. This has been done in various ways by Ness et al (8,9), Whang (17), Jackson and Beard (6), and Ng and Beard (11) with different results. When the intrinsic field is assumed dipolar, the values inferred for the magnetic moment range from $2.9\text{--}5.2 \times 10^{22} \text{G-cm}^3$ depending upon the portion of the data considered and the method used. Dipole moments of $2.4\text{--}2.9 \times 10^{22} \text{G-cm}^3$ are found when both MI and MIII are modeled together and the intrinsic field is allowed to possess higher order moments and/or offset positions within the planet (17,6,11). However, there are serious doubts concerning the physical reality of higher order moments inferred from this spatially and temporally limited set of observations (8,10,11).

An alternate approach to determining the Hermaean magnetic moment is to infer values of the solar wind stand-off distance, r_s , from the positions of the Mariner 10 bow shock and magnetopause encounters. The bow shock shape and position are determined by the forward portion of the magnetopause surface which is expected to have the same blunt shape at both the earth and Mercury. Hence, with a knowledge of the shape of the terrestrial bow shock (1) and the ratio of the distance to the nose of the shock to r_s (3) the solar wind stand-off distance may be calculated at the time of each of the 4 bow shock crossings. The earth's magnetopause out to sun-earth-satellite angles of $\sim 110^\circ$ is known to be elliptical with the earth at one focus (4). If the solar wind is indeed deflected around Mercury by an intrinsic dipole moment as found by Ness et al (9), then the shape of the magnetopause out to a sun-Mercury-satellite angle where the tail field begins to dominate will also be elliptical. Because the Hermaean magnetotail current density is greater than in that of the earth (8), this angle will be less than 110° . For the purposes of this study the Hermaean magnetopause has been assumed elliptical out to sun angles of $\sim 100^\circ$ so that the values of r_s were inferred from the two outbound crossings at 83° and 101° , but not for the inbound crossings at 135° and 121° . In this way the solar wind stand-off distance has been determined on 6 occasions during MI and MIII when the satellite encountered these boundaries.

The stand-off distances found above are a function of the solar wind dynamic pressure, the amount of magnetic flux transferred to the tail by erosion, and the intrinsic planetary magnetic moments. Due to an instrument failing to

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function after launch, no in situ solar wind ion observations were made by Mariner 10. Solar wind dynamic pressure during MI and MIII may be inferred from the electron plasma observations (12,13), the observed magnetic field with assumptions to determine the state of compression, and solar wind observations made near the earth and scaled back to Mercury at the time of the encounters (16). In this study the pressure during MI and MIII is assumed to be 1.3×10^{-7} and 2.6×10^{-7} dynes/cm $^2 \pm 30\%$ respectively, on the basis of the magnetic field observations (16). Whereas an increase in solar wind pressure produces a compression of the dayside magnetosphere, erosion transfers magnetic flux to the magnetotail and alters the magnetospheric currents in such a way to produce a contraction of the dayside magnetosphere. Slavin and Holzer (16) have scaled the effects of erosion on r_s at the earth to Mercury and found that the magnetopause may be found within $0.2R_m$ of the Hermaean surface a significant portion of the time. In addition, this increase in tail flux enhances the drag on the tail by the solar wind which is then balanced by the neutral sheet moving closer to the earth (e.g. 2). This results in decreases in the magnetic field strength of the near tail regions as evidenced during the outbound portion of MI. The stand-off distance at which a balance between the solar wind and the magnetic pressure is obtained is also dependent upon the strength of the planetary magnetic moments. As the models of Whang (17) and Jackson and Beard (6) which attribute significant portions of the fields observed by Mariner 10 to quadrupole and/or octupole moments are not consistent with the solar wind pressures and values of r_s found above, it is assumed that the Hermaean magnetic field is dipolar as found by Ness et al (9). Pressure balance at the stagnation point then requires $M_0^2 = 2\pi\rho v^2 R_s^6 k/f^2$ where M_0 is the dipole moment, ρv^2 is the solar wind dynamic pressure, R_s is the stand-off distance in the absence of erosion (i.e. ground state magnetosphere), and k/f^2 is a constant. With this relationship all 6 values of r_s may be scaled to a single solar wind pressure as has been done in the table and the dipole moment calculated. Since the IMF was observed to be northward only prior to the MIII bow shocks, the stand-off distances inferred from these crossings should correspond approximately to uneroded stand-off distance, R_s .

Crossing	r_s	As shown the mean stand-off distances for MI and MIII are 1.5 $1.9R_m$ respectively. The smaller values of r_s inferred for MI are consistent with both the southward IMF and substorm signatures (15) observed during this encounter and not during MIII. In addition, this evidence for magnetic flux transfer during MI
MI Inbound BS	$1.33R_m \pm 14\%$	
Outbound MP	$1.76R_m \pm 13\%$	
Outbound BS	$1.43R_m \pm 20\%$	
MIII Inbound BS	$2.08R_m \pm 11\%$	
Outbound MP	$1.62R_m \pm 15\%$	
Outbound BS	$1.98R_m \pm 17\%$	
(All scaled to 6.0×10^{-8} dynes/cm 2)		

explains the finding of smaller dipole moments for MI relative to MIII from modeling the observations with the magnetopause and neutral sheet currents assumed to have the same position and intensity during both encounters. The uneroded stand-off distance inferred from the boundary crossings is the largest value in the table, $2.1 \pm 0.2R_m$. With this stand-off distance and the previously stated assumptions and uncertainties a Hermaean dipole moment of $6 \pm 2 \times 10^{22}$ G-cm 3 is found from the pressure equilibrium condition. This value is in substantial agreement with the values of 4.7 – 5.2×10^{22} arrived at by Ness et al (9). However, final determination of Mercury's magnetic moment must await an orbiter mission of sufficient duration to study the time dependent contributions of magnetospheric currents and possibly planetary induction currents (5).

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THE MAGNETIC FIELD OF JUPITER: Edward J. Smith
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The main properties of Jupiter's intrinsic magnetic field are known from radio astronomy observations and from the Pioneer 10, 11 measurements. Qualitatively, the field is that of a reasonably well-centered dipole. A good quantitative approximation is provided by a dipole with moment $4.2 \text{ Gauss } r_J^3$, tilted 10° relative to Jupiter's rotation axis and displaced from the planet's center by $0.1 r_J$ in an equatorial direction. These parameters, as well as the longitude of the magnetic pole, are consistent with both the remote and the in situ observations. However, both sets of observations provide clear evidence that the jovian field is actually more complex than that of a simple dipole. Spherical harmonic analyses of the spacecraft data have been carried out in the region between 1.6 and $8 r_J$ and reveal significant contributions from a centered quadrupole and octupole. Non-dipole structure is revealed in the radio astronomy data through asymmetric variations in brightness, in the position angle of linear polarization, etc. In principle, it should be possible to obtain a better description of Jupiter's field from both sets of data than from either set taken alone. Furthermore, the radio astronomy data acquired over many years also make it possible to look for secular variations. By a fortunate coincidence, Sam Gulkis and I have just completed a review of spacecraft and radio studies of Jupiter's field for a forthcoming volume of Annual Review of Earth and Planetary Sciences. This talk will be based to a large extent on that review.

MAGNETIC FIELDS OF THE GIANT PLANETS, R. Smoluchowski
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Ever since the two Pioneer spacecraft provided us with spectacular data about the Jovian magnetosphere (1) much effort has been devoted to the search for magnetic fields of the other giant planets. Although the larger distances and weaker fields of these planets prevent obtaining data of an accuracy and detail comparable to what we knew about the Jovian field before the Pioneer flights, it seems possible to draw certain general conclusions and make comparisons apart from the tentative linear relationship between magnetic moments and angular momenta of all the planets. In this paper the internal fields and their origin will be discussed with only parenthetical mention of the external fields and of magnetospheres although these provide the main clues for the study of the former. The analysis of the origin of the magnetic fields of the giant planets is somewhat facilitated by the fact that, in contrast to the terrestrial planets, the chemistry and structure of the giant planets is relatively simple. Unfortunately the various parameters which enter into the necessary conditions (2,3) (no sufficient conditions have as yet been unambiguously derived) for the generation of an internal magnetic field are so sensitive to subtle details of the models that no rigorous predictions can be made at this time.

The generation of a magnetic field by the hydromagnetic dynamo mechanism (2) requires a sufficiently high magnetic Reynolds number in which the electrical conductivity of the liquid plays the major role. For this reason it was generally accepted that on Jupiter the field is generated (4) only in the metallic hydrogen interior ($r < 0.75 R_J$). It turns out, however, that the high pressures and temperatures prevailing on this planet lead to the presence of metallic conductivity also in the inner part ($0.75 < r < 0.92 R_J$) of the H_2 mantle (5) thereby increasing the high conductivity volume by about a factor of two and bringing it closer to the surface. The latter aspect seems important from the point of view of recent theories (6) which indicate that the field is generated primarily in the outer parts of the high conductivity region. It may also help to understand the unusually high quadrupole and octopole components of the field just above the surface (5).

In contrast to the early negative results recent terrestrial observations (7) indicate that there is indeed a magnetic field on Saturn but that the rings suppress the formation of a strong radiation belt (1) so that only hectametric pulses and no decimetric radiation are observed. Four recent models of Saturn have been examined (8) and it turns out that the volume of high conductivity H_2 is 3 to 6 times larger than that of the metallic hydrogen so that it is in this region ($0.4 < r < 0.7 R_S$) that the field is generated. The top of this region is, however, more than twice as deep below the surface of the planet as on Jupiter so that the higher components of the field are probably weak. There is also the possibility that the magnetic field of Saturn is generated by the thermoelectric effect or by the precession currents induced by Tethys (3).

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One treads on much shakier ground when the magnetic fields of Uranus and Neptune are considered. Although now we know that Uranus has a field but probably no internal heat while the opposite is true for Neptune (9) one cannot claim that these facts are in agreement or not with what we know about the interiors of the two planets. The absence of a heat source on Uranus makes the hydromagnetic dynamo mechanism somewhat questionable. The thermoelectric mechanism in the H_2 layer is a possibility and one should not forget either, that, in contrast to the satellites, Uranian rings (especially the biggest ϵ -ring) have probably substantial inclinations to the equatorial plane and could induce the Malkus precessional mechanism (3). In spite of huge differences between the models of these planets as proposed by various authors (10) one can conclude that, unlike Jupiter and Saturn, neither one has much, if any, metallic hydrogen and that both have large cores made of "rocks" and "ices" although metallic ammonia has also been suggested (10). On the whole the pressure and density in Uranus are lower than those in Neptune and depending upon the equations of state used one might expect much of the cores to be solid. The magnetic could be thus generated in the liquid H_2 layers and in the liquid parts, if any, of the cores. Neptune, having a presumably colder core with lower electrical conductivity and a thinner H_2 layer would be less likely to sustain a hydromagnetic dynamo than Uranus. These conclusions are clearly based on hindsight rather than on foresight. Best estimates indicate that the magnetic moments of Jupiter, Saturn and Uranus are in the ratio 1000:100:1.

Finally one should mention that while the Jovian magnetosphere, or rather the curious magnetodisc, has been investigated in considerable detail (1) all one can do about Saturn is to scale down the Jovian magnetosphere (12) and take into account the perturbing role of the rings. Uranus with its unique enormous tilt of the axis must have an interesting magnetosphere and magnetotail (1,8). In fact one may suspect that if there is a substantial radiation belt then this belt and the radiation reaching Earth will show seasonal variation with a period of 84 years. One does not expect significant interaction of this magnetosphere with the rings because their width is several hundred times lower than that of the rings of Saturn (13).

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LUNAR THERMAL HISTORY VERSUS LUNAR MAGNETISM, OR
HOW TO BOX IN A HYPOTHESIS AND BUILD A COFFIN; Sean C. Solomon,
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All theories for internally generated magnetic fields for the Moon involve explicitly or implicitly some assumptions about the lunar thermal history. It is of interest as a test of such theories to explore fully the range of possible thermal histories for the Moon consistent with geological, geophysical, and geochemical data independent of the evidence from lunar paleomagnetism. The range of possible thermal models on these grounds is surprisingly narrow, and in particular poses severe problems for an early (first 0.5 b.y.) lunar core dynamo.

The essential constraints of a thermal history model are (a) that the current heat flow [1] and relative abundances of radioactive heat sources be satisfied, (b) that a 70 km thick, feldspar-rich crust have formed by crystal-liquid differentiation [2] early (first 0.2 b.y.) in lunar history [3,4], (c) that a distinct lunar core be no larger than 400 km radius at present [5,6], (d) that major surface volcanism have lasted until at least 3.0 b.y. ago [7], (e) that the thermal stress in the lunar lithosphere never have accumulated to high enough levels (≥ 1 kbar) to cause global-scale failure [8], but (f) that the global thermal stress provided an important modulation to the local stress due to mare basalt loading in mare basins so as to affect the dominance of local extensional or compressive failure [9].

Constraint (b) excludes models with a cold initial state in favor of models with at least the outer few hundred kilometers of the Moon initially at or near melting temperatures. A necessary condition for (d), that partial melting persisted in the mantle until that time, is satisfied for most models consistent with (a) and (b). Constraint (c) dictates that core formation on the Moon is at most a very modest thermal event, releasing 10^{27} J of potential energy as heat, or only a 12°C equivalent mean temperature rise [10].

Constraints (e) and (f) are quite powerful. The absence of global-scale compressional or extensional tectonics on the Moon greatly restricts possible thermal models to those with nearly constant volumes and small accumulated lithospheric thermal stresses since the time of heavy bombardment. In particular, models with the outer 300 ± 100 km of the Moon initially melted and the deep interior of the Moon initially cold are possible [11], but models with completely molten initial states [e.g., 12] may be excluded. This conclusion includes consideration of heat transfer by subsolidus convection [13] and of models of the type advocated by Tozer [14]. The requirement that the change in the stress regime near mare basins from one dominated by graben formation to one dominated by mare ridge formation at about 3.6 b.y. ago [15] be due to a change in sign of incremental global scale thermal stress [9] restricts possible thermal histories still further (figure 1).

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Thus while metal-silicate fractionation in the outer 300 km may have occurred quite early on the Moon, formation of a central core close to final size should have awaited warming of the deep interior sufficient for relative flow of mantle and core material. Such a time should be close to the time of peak lunar volume, or about 1 b.y. after origin [9]. Models for magnetic field generation involving a fully developed fluid core with an active dynamo 4 b.y. ago [12] are thus unlikely. Models with a permanently magnetized, initially cold interior [16, 17] are acceptable on thermal and geological grounds.

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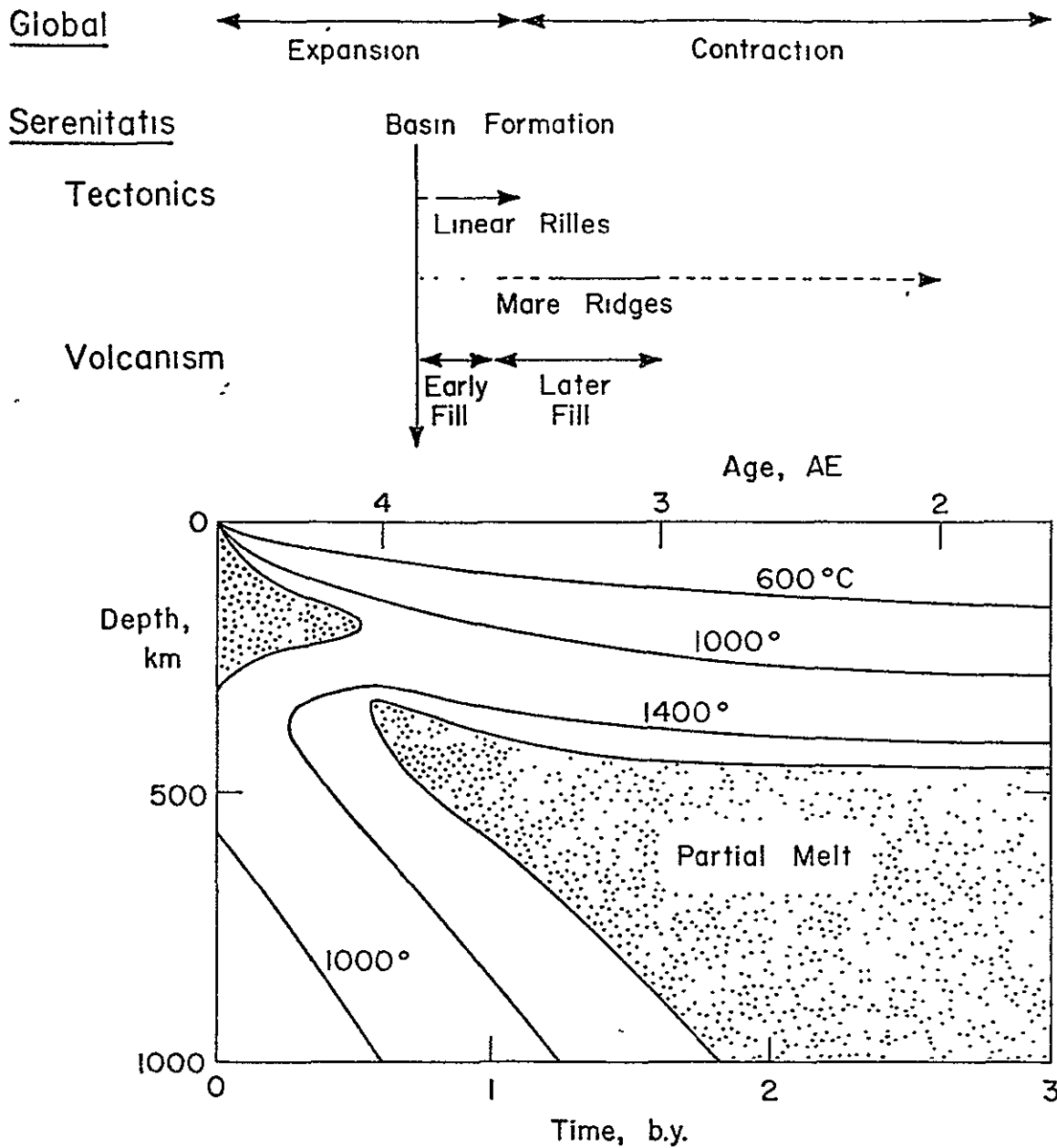


Fig. 1. Summary of temporal relationships among lunar thermal history, global expansion and contraction, and the tectonic and volcanic history of a typical mare basin, Serenitatis [9]. Isotherms are shown versus depth and time for a thermal history model with initial melting to 300 km depth and an initial central temperature of 300°C. Regions of partial melt are shown as stippled.

THE MAGNETIC FIELD OF THE EARLY SOLAR SYSTEM. C.P. Sonett, Lunar and Planetary Laboratory, Univ. of Arizona, Tucson, AZ 85721 (*also Dept. of Planetary Sciences).

The strongest evidence of the presence of a magnetic field during the formative phase of the solar system comes from observation of the presence of ancient thermoremanent magnetization possibly joined with chemical remanent in meteorites [1,2]. The requirement for redistribution of angular momentum in the proto-solar nebula also infers the existence of a solar spin braking mechanism, which can reasonably be driven by magnetic torques [3,4]. Magnetization of the Moon and Mercury, the most critical planetary examples provides less conceptual connection to the early field problem because, for example, the maria basalt magnetic fields were imprinted between 3.9 and 3.2 aeons ago, rather later than primordial. The source for Mercury's magnetic field is generally thought to reside in a dynamo; a less favored source is thermoremanence of the crust which could be primordial [5].

Recent work by Lanoix *et al* [6] on chondrules from Allende shows evidence for a primordial magnetic field strength as large as 16 oersted, a figure substantially higher than previous values of 0.18-2 oersted. This high value may reflect a different magnetic field at the time of acquisition of TRM by Allende chondrules than that for the matrix. The primordial magnetic field appears most likely to reflect the presence of either a magnetic field extended from the early Sun or a nebular dynamo [2].

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CONVECTION-DRIVEN DYNAMOS A.M. Soward, School of Mathematics, The University, Newcastle upon Tyne, U.K..

The fluid motions responsible for driving the geodynamo are generally believed to be the result of convection, thermal or otherwise. The detailed nature of the processes involved is still unknown. Nevertheless owing to its relative simplicity theoretical dynamo models have been developed on the basis of thermal convection.

A model, which incorporates the main features of the Earth's liquid core is as follows. A self-gravitating, Boussinesq fluid is confined in a spherical container and rotates rapidly about an axis with angular velocity Ω . The fluid contains a uniform distribution of heat sources, which leads to an unstable radial temperature gradient β . A dimensionless measure of β is the modified Rayleigh number

$$R = \alpha g \beta L^2 / \kappa,$$

where α is the coefficient of expansion, g is the acceleration due to gravity, L is the core radius, κ is the thermal conductivity. The critical value R_c of the Rayleigh number at which the onset of convection occurs depends on the Ekman number E and the Prandtl number σ , where

$$E = \nu / \Omega L^2,$$

$$\sigma = \nu / \kappa$$

and ν is the viscosity. A detailed numerical study of this model has been made by Roberts (1968). When the fluid is electrically conducting, the stability characteristics of the system are altered in the presence of magnetic field. Eltayeb and Kumar (1977) have considered the special case of an azimuthal magnetic field, which increases linearly with distance from the rotation axis. They find that R_c depends critically upon the sizes of

$$\Gamma = \sigma B^2 / \rho \Omega,$$

$$q = \sigma \mu \kappa,$$

where B/L is the magnetic field gradient, σ is the electrical conductivity, ρ is the fluid density and μ is the magnetic permeability.

For geophysically relevant values of the parameters, q is small and for a magnetic field of about 30 gauss Γ is of order unity. The corresponding value of R_c is of order one also and in this case the primary force balance is between the Coriolis and Lorentz forces. As the magnetic field is decreased, R_c increases and when the field vanishes R_c is proportional to $E^{-1/3}$. According to the model convection occurs most readily when the magnetic field is large enough to release the rotational constraint but not so large that strong Lorentz forces inhibit convection (i.e. $\Gamma = O(1)$). The result suggests that the geodynamo may operate in this parameter range but owing to the mathematical complexity no theoretical model has been developed. Instead attention has been directed mainly towards quasikinematic models for which the character of the motion is largely (but not completely) uninfluenced by the Lorentz force.

Busse (1975) has proposed a weak field model ($\Gamma < 1$) of the geodynamo, which he developed in an annulus geometry. A dynamo model, which is similar in many respects, was also developed by Childress and Soward (1972) in the simpler plane layer geometry. For this model Soward (1974) found stable hydromagnetic dynamos with very weak magnetic fields, $\Gamma = O(E^{-1})$. By contrast Childress (1976) discovered that, when the magnetic field is stronger, $\Gamma = O(E^{-2/3})$, the dynamo process is unstable. This is simply because increasing the field strength releases the rotational constraints. On the other hand, Busse's model is stable with substantially larger magnetic fields. This is explained by the additional constraints imposed by the sloping boundaries of the annulus. Nevertheless, for small but moderate field

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strengths there remains the danger of the instability isolated by Childress. If this occurs, it is reasonable to suppose that the dynamo will equilibrate with an order one value of Γ .

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CRUSTAL THERMOREMANENCE AND THE ORIGIN OF MERCURY'S PRESENT
MAGNETIC FIELD; L. J. Srnka, Lunar and Planetary Institute,
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Mercury is the only terrestrial planet in addition to the Earth which is known unequivocally to have an intrinsic global magnetic field. If the origin of this field could be shown to be due to dynamo action in a mercurian liquid core, the planet could serve as a crucial test for dynamo theories which have been developed primarily to understand geomagnetism.

The discovery from the Apollo program that lunar crustal rocks have magnetizations which date from the early history of the Moon, and the growing body of data on meteoritic magnetism, suggest that many solid objects in space are remanently magnetized. If this remanence is ordered, it may be possible to account for substantial planetary magnetic fields without invoking currently active dynamo processes. Models of this kind have been proposed for Mercury by Stephenson (1976) who studied crustal remanence acquired in a mercurian dipole field produced by a once-active dynamo, and by Sharpe and Strangway (1976) who considered the possibility of fossil magnetization in Mercury acquired from a strong primordial solar magnetic field. The resolution of this present dynamo *vs.* remanent field question is important for the further understanding of dynamo theory and (through the physical link of convective processes) the thermal evolution of Mercury and the other terrestrial planets.

The acquisition of global thermoremanent magnetization (TRM) in Mercury's lithosphere is studied here using a planetary model which incorporates both the thermal evolution of the planet and the possible characteristics of a mercurian dynamo. The thermal evolution model of Solomon (1977) is used to track the evolution of the Curie isotherm depth (for Fe) with time. Mercury's lithosphere is assumed to consist of an unspecified paramagnetic mineral matrix containing dispersed, equidimensional Fe inclusions which are multidomain and are present in volume concentrations f of $10^{-2} \leq f \leq 10\%$. A core-generated dipole field is assumed to have existed, with the general form:

$$P(t) = P_0 t^n \cos(2\pi t/T) \exp(-t/\tau)$$

where T is the reversal period, τ is the decay constant for the field envelope at late times, and the factor t^n models the rise of the dipole moment during core formation. As a test case, the parameters $n = 2.47$, $\tau = 5 \times 10^8$ yr, and $10^5 \leq T \leq 10^8$ yr were used. These result in a dipole field which peaks at 700 myr after the "origin" of the planet, after Mercury's core has fully formed (Solomon, 1977). The field is "turned off" 100 myr ago, after which time the interior changes very little.

The results of this calculation are given in Fig. 1. These show that if the ancient magnetic dipole moment of Mercury's core-generated field did not exceed $\sim 5 \times 10^{29}$ G cm³, if the dipole field reversed at least once every 10 myr, and the volume fraction of Iron in the planet above the Curie isotherm depth is less than 10% by volume, then the present upper limit on Mercury's dipole moment of $\sim 4.8 \times 10^{22}$ G cm³ (Ness, 1978) is too large to

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be attributed to crustal TRM acquired in an ancient mercurian dipole field. On this basis, a currently active dynamo is favored as the source for Mercury's field. On the basis of the lunar evidence, primordial magnetization in Mercury due to an early, strong solar field is unlikely to be the source of its present magnetic dipole moment.

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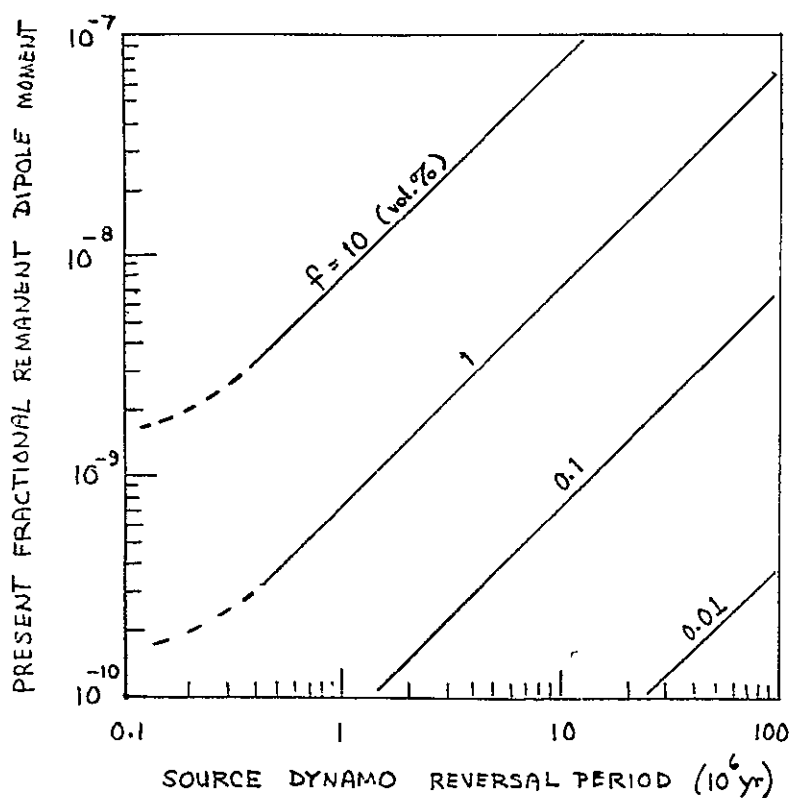


Figure 1

A STUDY OF THE RIMA SIRSALIS LUNAR MAGNETIC ANOMALY;
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The discovery of a magnetic anomaly associated with the lunar grabenlike nonsinusuous rille Rima Sirsalis by Anderson *et al.* (1976, 1977) is the first positive identification of a lunar surface feature which has a local magnetic field. In this paper we study various source models for the anomaly which we believe are compatible with the geology of the rille and its surroundings.

The source of this linear anomaly has been modelled using exact solutions for the scalar potential and vector field intensities due to arbitrary uniform remanent magnetization in a finite rectangular prism. The earlier work of Anderson *et al.* (1977) used a line dipole model in the form of a circular cylinder of infinite length, which has the disadvantage of poorly representing the grabenlike structure of the rille and hence the higher order magnetic moments of such an anomaly (i.e., fields due to sharp corners) near the surface. As a result, the required source magnetization has been underestimated.

Three classes of source models were studied: 1) rectangular prisms with magnetizations limited to those found in lunar samples; 2) a gap in a uniformly magnetized layer; and 3) a system of narrow, parallel dikes intruding the rille and its surroundings. The following general results were obtained: if the field source is confined to the width of the rille and the magnetization does not exceed that found in lunar samples, the body responsible for the anomaly must have a vertical extent of tens of kilometers. More specifically, if the field source is a number of thin (100 - 200 m wide) dikes beneath the rille, the intruded material must have a uniform downward component of remanence of at least 10^{-5} emu g^{-1} over at least 30 km of depth. We also find that all reasonable gap models for this anomaly require surface magnetizations of at least 10^{-3} emu g^{-1} if the gap dimensions are those of the rille. All of the source models require a total magnetic moment (all orders) of $3 - 8 \times 10^{15}$ G cm^3 for this anomaly if the Apollo 16 subsatellite magnetometer data ($B_z \approx -0.1\gamma$ at 155 km altitude) is to be matched. This moment is comparable to the magnetic dipole moment derived for the Van de Graaff lunar magnetic anomaly by Russell *et al.* (1975).

If the model magnetization is constrained to those characteristic of the returned lunar samples, we favor a massive dike swarm as the principal source of the Rima Sirsalis anomaly. A system of 20 parallel dikes, each 200 m wide, lying within 5 km of the axis of the rille, and extending from near the surface to a depth of 50 km or so (the base of the crust) would produce the required field at 155 km altitude if the dikes have a downward magnetization of $2 - 3 \times 10^{-5}$ emu g^{-1} . The surface fields in such a model are irregular, with vertical component peaks of about 350γ over the individual dikes, which is consistent with the electron reflection data (Anderson *et al.*, 1977). However, this remanence is considerably greater than that found in lunar basalts (by a factor of 3 or more) so that it is necessary to invoke

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unusual magnetic properties for this region of the moon.

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THEORETICAL CONSTRAINTS ON THE MAGNETISM OF GASEOUS AND ICY PLANETS.

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This paper is in part a compilation of the present theoretical understanding of the interiors of gaseous and icy planets, and how this knowledge relates to the existence of a global planetary magnetic field (c.f. Stevenson, 1978a). However, the emphasis is on three new results: (i) A global constraint on the Jovian dynamo which demonstrates its turbulent character. (ii) An estimate for the outer boundary of dynamo cessation in these planets from a consideration of dynamo generation in a strong conductivity gradient. (iii) A consideration of high pressure water and/or ammonia, indicating that dynamo generation in icy bodies (Uranus, Neptune; perhaps even Ganymede, Titan, Pluto ...) is conceivable.

The Jovian Dynamo: The total internal heat flux of Jupiter is $Q \sim 5 \times 10^{24} \text{ erg/sec}$. In a dynamo driven by thermal convection, the Ohmic dissipation should be of order $Q_c d/H_T$ where Q_c is the core flux, d is the thickness of the dynamo region and H_T is the temperature scaleheight (Hewitt et al., 1975). The dissipation could be even larger if the convection is non-thermal, but the formation of helium rain, which has been suggested as an energy source for Jupiter and Saturn (Salpeter, 1973; Stevenson and Salpeter, 1977b) is still a predominantly thermal source because of the small size ($\sim 1 \text{ cm}$) of the raindrops. Since Q_c is almost as large as Q , and $d \sim H_T$, the Ohmic dissipation is comparable to the total heat flux. However, a global field of 10 Gauss in Jupiter would only dissipate of order 10^{16} erg/sec for a magnetic diffusivity of $2 \times 10^2 \text{ cm}^2/\text{sec}$ (Stevenson and Salpeter, 1977a). It is highly improbable that the eight orders of magnitude discrepancy is due to an enormous (10^5 Gauss) toroidal field in Jupiter. Instead, the discrepancy is best explained by a turbulent dynamo in which the dissipation is primarily at small lengthscales (i.e. kilometers). The known properties of Jupiter, together with a model of turbulent thermal convection (Stevenson, 1978b) suggest a magnetic energy spectrum consistent with that expected for helical turbulence (Pouquet et al., 1976).

Dynamo Generation in a Conductivity Gradient: These planets are characterized by no rigid outer boundary at which dynamo generation ceases (such as the core-mantle boundary of the Earth) but by a gradual cessation of dynamo generation because of a magnetic diffusivity which increases rapidly in the radial direction. An example of this is molecular hydrogen at megabar pressures, where the band gap between valence and conduction electron states becomes sufficiently small for the electrical conductivity to be almost metallic (Smoluchowski, 1975). A consideration of α -effect dynamos suggests that the outer boundary of dynamo generation can be defined crudely in terms of the region in which the toroidal field becomes small relative to the poloidal field. This region is characterized by $V l_G / \lambda \sim 1$ where V is the convective velocity, λ is the magnetic diffusivity and $l_G \equiv (d \ln \lambda / dr)^{-1}$ is the conductivity scaleheight. Since l_G is much less than the planetary radius, dynamo generation ceases at greater depths than is sometimes supposed. The outer boundary of dynamo generation is about 0.8 of the radius in Jupiter and 0.5 of the radius in Saturn. These results also have implications for possible magnetic secular variations in these planets.

Dynamo Generation in Icy Bodies: Infrared observations (Lowenstein et al., 1977) suggest that Neptune has an internal energy source but Uranus does not. Actually, this is misleading because Uranus is closer to the Sun so its smaller internal energy source is more difficult to detect (Hubbard, 1977). In fact, the deep interiors of both planets should be warm ($\sim 3000^\circ \text{K}$), fluid and convective (Stevenson, 1978a). Water does not become metallic until

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megabar pressures, but it is already an excellent ionic conductor (with a magnetic diffusivity of order $10^6 \text{ cm}^2 \text{ sec}^{-1}$) at about 200kbar (Hamann and Linton, 1966). Addition of NH_3 aids the conductivity further. The possibility of metallic ammonium seems unlikely (Stevenson, 1975) but an adequate magnetic Reynolds number for dynamo generation seems likely (Stevenson, 1978b). The icy satellites and Pluto may also have convective, fluid H_2O - NH_3 layers, but the pressures are lower and dynamo generation is marginal at best.

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THE ROLE OF ROTATION IN PLANETARY DYNAMOS

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Dynamo generation is usually discussed in terms of a magnetic Reynolds number R_M . Since rotation evidently plays a crucial role, the value of the Rossby number R_O (representing the ratio of inertial to Coriolis effects) should also be considered. The basic conjecture in this work is that R_M and R_O together determine necessary and sufficient conditions for dynamo existence. Rotation also determines the symmetry of the resulting field and its magnitude. To support these assertions, a turbulent thermal convection model is constructed in which nominal values of R_M and R_O are directly determined by F_C , the convective heat flux. (The basic principles are unaltered if a compositional gradient drives the convection.)

Relevant and Irrelevant Parameters: The flux F_C , convecting fluid depth d , rotation rate Ω , magnetic diffusivity λ , boundary conditions and thermodynamic properties are sufficient to specify the system. Viscosity ν and thermal diffusivity κ are excluded since $Vd \gg \nu, \kappa$ for any realistic convective velocity V . (In the Earth's core, $Vd \sim 10^6 \text{cm}^2 \text{sec}^{-1}$, $\nu \sim 10^{-2} \text{cm}^2 \text{sec}^{-1}$, $\kappa \sim 10^{-1} \text{cm}^2 \text{sec}^{-1}$). Turbulent convection is inevitable, and any dynamo model which explicitly depends on ν or κ is either unrealistic or relies crucially on the effects of boundary layers.

The Convective Model (Flasar and Gierasch, 1978; Stevenson, 1978) considers the growth rates of linear, Boussinesq convective modes in a planar geometry for a superadiabatic fluid with imposed uniform rotation and magnetic field. Velocity amplitudes are estimated by assuming that these modes are stabilized by non-linearities, such as shear instabilities, which cascade the kinetic and thermal energies into smaller scale motions. The dominant mode is assumed to be that which minimizes the temperature gradient for a given heat flux. For no field or rotation, the predictions are equivalent to those of mixing length theory, and agree with observed convective velocities in stellar atmospheres (Böhm-Vitense, 1977) and thermally-driven turbulence in the earth's atmosphere (Monin and Yaglom, 1971). For rapid rotation and no field, the preferred modes are rolls aligned along the horizontal component of Ω (c.f. Busse and Cuong, 1977). The predicted convective velocity is

$$V \approx (F_C d / \rho H_T)^{6/5} (|\Omega_Z| d)^{-1/5} \quad (1)$$

where Ω_Z is the vertical component of Ω , ρ is the fluid density and H_T is the temperature scaleheight. The wavevector of the preferred mode is essentially azimuthal and has magnitude $\sim R_O^{-1/2} d^{-1}$ where $R_O \equiv V/2|\Omega_Z|d$. Provided $R_O \lesssim 1$, the flow possesses a non-zero horizontally averaged helicity $\langle \mathbf{V} \cdot \nabla \times \mathbf{V} \rangle$ and an α -effect is predicted which can transform toroidal field B_T to poloidal field B_P (c.f. Childress, 1977):

$$\left(\frac{\partial}{\partial t} - \lambda \nabla^2 \right) B_P = \nabla \times (\alpha B_T) \quad (2)$$

$$\alpha \approx \frac{V^2 R_O^4 \Omega_Z d}{2\lambda |\Omega_Z|} \sin(2\pi z/d) \quad (3)$$

where z is the vertical coordinate. This α -effect does not transform B_P back to B_T , but this can be done by:

The Thermal Wind. The convective model predicts a latitude dependent temperature gradient for a given heat flux. For realistic boundary conditions, a horizontal temperature gradient is unavoidable. As in the Earth's

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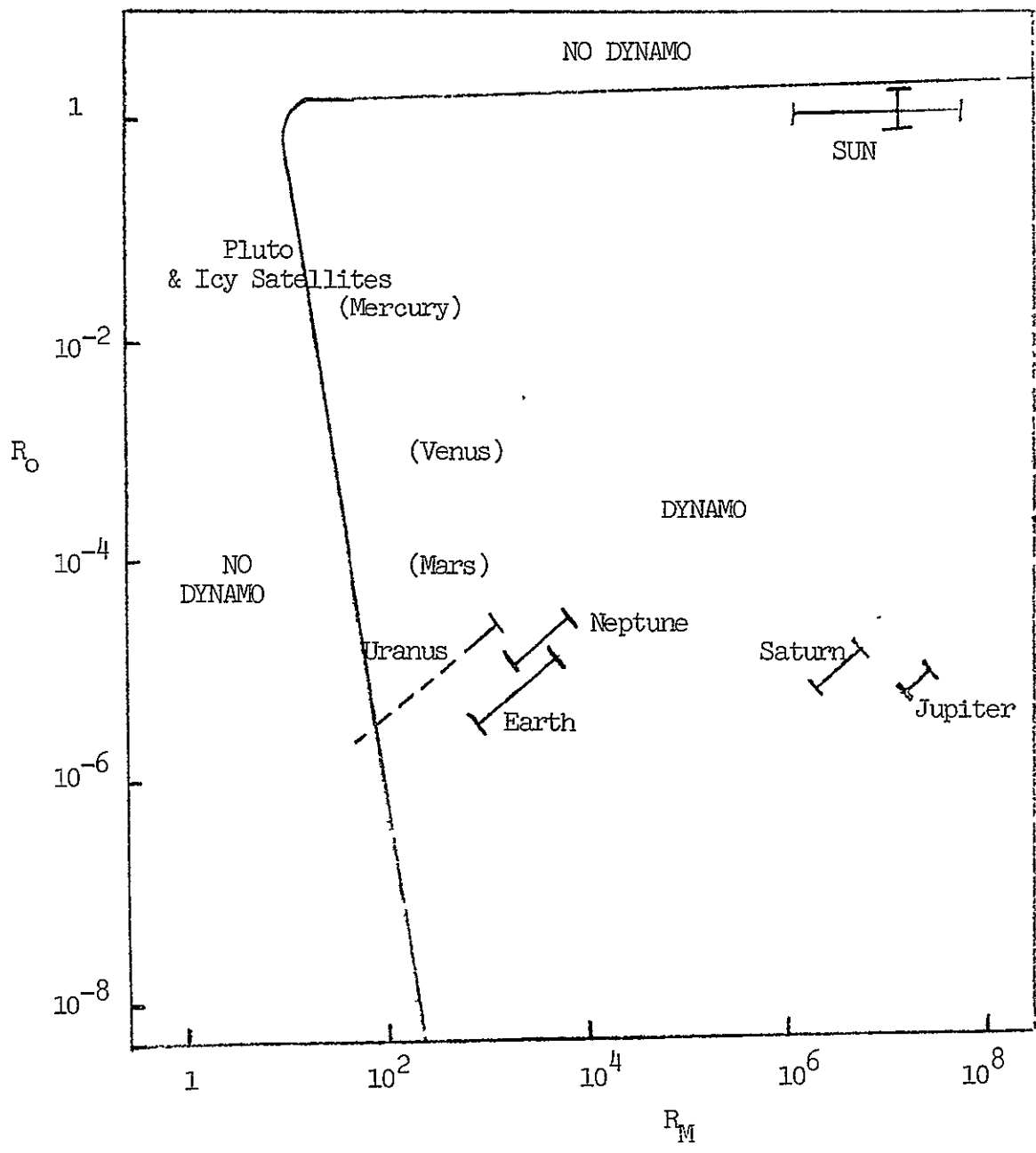


Fig.1: The Rossby number - magnetic Reynolds number plane for convective dynamos, showing the curve which separates dynamo from non-dynamo. Positions of Earth, Jupiter, Saturn, Neptune and the Sun are reasonably well established. The dashed error bar for Uranus extends into the non-dynamo region because of the large uncertainty in the heat flux. Mercury, Venus and Mars are shown bracketed at the positions they would occupy for $F_c \sim 10$ c.g.s. In fact, F_c may be zero in all three. Pluto and the icy satellites lie at the boundary and are tests of the theory.

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atmosphere, a thermal wind is expected (c.f. Eady, 1949). This provides the "w-effect" to complete an $\alpha\omega$ dynamo.

The Zero-Field $\alpha\omega$ -Dynamo: The model predicts a thermal wind U_{TW} of order V and $\alpha \sim VR_0^{-1/2} R_M$ where $R_M \equiv Vd/\lambda$. Dynamo onset therefore requires (Moffatt, 1978):

$$\left(\frac{U_{TW}d}{\lambda}\right) \left(\frac{\alpha d}{\lambda}\right) \geq 100 \quad (4)$$

(the precise numerical value is unimportant). Necessary and sufficient conditions for dynamo generation are thus $R_M \geq 10R_0^{-1/6}$ and $R_0 \leq 1$. This is illustrated in Fig.1 and can be calculated for planets if F_C is known.

Finite-Field Equilibration: If $10R_0^{-1/6} \leq R_M \leq 10R_0^{-1/4}$ then the field equilibrates by reducing α and U_{TW} so as to satisfy eq.(4). (The finite field enhances convective efficiency, and reduces convective velocity and temperature gradients). However, if $R_M \geq 10R_0^{-1/4}$ then a bifurcation is possible and two convective states (one primarily geostrophic, the other primarily magnetostrophic) can coexist. The magnetostrophic solution prevents further field amplification, but no truly steady-state can be achieved. Poloidal and toroidal fields are comparable, and the bifurcation may be related to reversals. The ultimate field amplitude is around $(8\pi\rho\lambda\Omega)^{1/2}$ or $(4\pi\rho V^2)^{1/2}$, whichever is larger; but complications arise if $R_M \gg 1$. In Fig.1, an attempt is made to locate the possible positions of planets and satellites on the $R_M - R_0$ diagram (see figure caption).

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ROCK MAGNETISM OF EXTRATERRESTRIAL MATERIALS, D.W. Strangway, Dept. of Geology, Univ. of Toronto, Toronto, Ont. Can. M5S 1A1

Access to the Apollo samples has suddenly brought a new focus to rock magnetism. In most terrestrial environments we are dealing with the iron-titanium oxides and with some of the sulphides. The principles of the type of magnetism carried by these have been extensively studied. The lunar samples however have required us to examine the magnetic properties of iron and iron-nickel phases and of troilite in particular. The properties of the iron-nickel-cobalt phases were studied extensively in the 1930's and it has been a question of reviewing old literature.

Pure iron has a Curie point of 780°C and is extremely magnetic. The Curie point decreases slightly with increasing nickel content but the phase change between α and γ phase is magnetic while the γ -phase is not so this transition becomes the effective Curie temperature. Thus kamacite ($\sim 7\%$ nickel) becomes magnetic on cooling through a sluggish phase change. Taenite ($\sim 40\%$ nickel) on the other hand is γ -phase but at this composition it is magnetic below a few hundred $^{\circ}\text{C}$ and becomes magnetic on cooling. These pure ferromagnetic substances have their bulk magnetic properties controlled largely by grain size and grain shape.

Extremely small grains $<100 \text{ \AA}$ are superparamagnetic and carry no remanence. Above $100 \text{ \AA}$ they behave as single domain grains and hence are extremely stable magnetically. At grain sizes of 1μ or so they become multi-domain and hence appear to be magnetically soft.

The lunar igneous rocks contain mainly grains of kamacite in the few micron size range while the soils and breccias contain in addition abundant super-paramagnetic iron as a result of steady reduction of iron silicates. The magnetic properties of meteorites are dominated by kamacite taenite and plessite (intergrowths of the above two) with some contributions from magnetite and possibly pentlandite.

Troilite is antiferromagnetic at 320°C and so its magnetic effects can be detected, but they contribute little to the overall remanent magnetism.

PALEOINTENSITY STUDIES ON LUNAR SAMPLES, D.W. Strangway and
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In spite of considerable efforts to determine satisfactory values of the ancient lunar field, there are still very few acceptable determinations. It is therefore not yet possible to draw firm inferences about changes in the magnetic field during the history of the moon.

The determinations have proved to be exceptionally difficult for several reasons. These are:

- i) many lunar samples have large, soft components of magnetization that make it difficult to isolate the stable component;
- ii) On heating to simulate the original cooling at about 500°C in spite of studies using vacuum and controlled fugacity environments. There is more work in progress on this problem in several laboratories.
- iii) The intergrowth textures present in the lunar sample magnetic minerals lead to complex interaction which reduce the elegance of the Thellier-Thellier method;
- iv) Attempts to use non-heating methods depend on an assumption of critical ratio relating thermoremanent magnetization (TRM) to anhysteretic magnetization (ARM). These ratios are known to vary considerably especially when interactions are present.
- v) Shock effects can add several complexities. In the absence of a field, simple shock will reduce the magnetization preferentially removing the soft components. In the presence of a field, simple shock may increase the magnetization observed. In the course of shocking a rock, the heating process will be either minor (little thermal effect on NRM), major (if heated over 800°C, NRM will be totally reset) or intermediate. In the intermediate case the samples are reheated in a very inhomogeneous manner so that some parts of a sample will be magnetically reset and others will not. It is not possible to fully reproduce this latter case to create an artificial magnetization for paleointensity determination.

We conclude that while a few samples do reproduce ancient lunar fields it is not yet possible to fully reconstruct the ancient lunar field.

MAGNETIC STUDIES ON THE ALLENDE METEORITE, N. Sugiura and D.W. Strangway, Dept of Geology, Univ. of Toronto, Toronto, Ont. Can. M5S 1A1

Earlier studies on the Allende meteorite showed that the natural remanent magnetism (NRM) behaved like a simple thermoremanent magnetism (TRM) up to 130°C. This magnetization appears to have been acquired in a field of about 1.0 oersted. At this temperature it was found that the remanent magnetization showed a sharp change. This effect could be due to one of several effects i) reheating in a different field at some stage to 130°C; ii) chemical changes due to the laboratory heating or; iii) some property of the material involved. We now know that iron with minor impurities undergoes sharp reversible changes in its magnetic properties at this temperature probably explaining this difficulty.

Subsequent experiments in our laboratories have been done using individual chondrules separated from the Allende meteorite. The chondrules vary from strongly magnetic to moderately magnetic to weakly magnetic chondrules are the light-coloured ones with presumably less iron content. In general, it appears that the most stable magnetic component of the individual chondrules is randomly oriented. This suggests that the chondrules themselves were magnetized by cooling before they assembled into the meteorite. It appears that some of these were magnetized by cooling in fields on the order of 10 oersteds.

The chondrules then have in addition, a softer component of magnetization and the direction of this component while showing some scatter shows a common direction. This is evidence that the meteorite was magnetized at or after the assembly of the meteorite. The matrix material of the meteorite is also quite magnetic. At least two stages and perhaps three stages of magnetization may be present -- a) magnetization of the chondrules before assembly; b) magnetization of the whole meteorite during or near assembly and c) remagnetization, perhaps not uniformly, by subsequent metamorphism.

In either case we conclude that we have a record of early magnetic fields that predate the formation of Allende.

EVOLUTION AND THE STATE OF THE CORES OF TERRESTRIAL PLANETS;

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The internal structure and temperature regimes of a planet play an important role in generating and sustaining its magnetic field. Whether the planetary magnetic fields are due to remanent magnetism or are of magneto-hydrodynamic origin in their cores, may be discriminated by their temperature profiles. The first mechanism requires that the temperature within an appreciable portion of the planet remains below the Curie temperature throughout its history. The second mechanism requires the existence of a conducting liquid core such that dynamo actions can be derived either by internal heat sources or by external interactions. In this paper we calculate thermal evolution models of the inner planets and compare their present day temperatures and the state of their cores. The Earth is used as a reference in these studies.

The Moon: The existence or lack of a lunar core is not known. If an iron-rich core does exist, it is unlikely that it will exceed 400 km in radius. This upper bound on the lunar core size has been established by density constraints, seismic studies¹ and electro-magnetic induction studies². Even if the lunar core has a size equivalent to the maximum allowed, it only represents about 1% of the total volume. Therefore, such a small core will not play any significant role in the thermal history of the Moon. Based on our thermal evolution studies, if the Moon has a homogeneous accretion origin, its core has probably been molten since its formation about 3.5 billion years ago³. Whether this small core is able to induce the remanent magnetism observed in the lunar crust is another matter. If the surface magnetism of the Moon is the remanence of a primordial field⁴, this requires a very special thermal evolution model.

Mercury: Mercury has the second largest magnetic field and the highest iron content among the inner planets. Whether this field is generated by the memory of a primordial field or by a presently active dynamo is still a subject of debate. Most data favor a differentiated model of the planet. Thermal evolution models that can produce differentiation favor an initially hot planet. An iron-rich core can be separated early in the evolutionary history of the planet. In order to have a molten or partially molten core at the present time, it is necessary to have a small amount of heating within the core.

Mars: The existence of a weak Martian magnetic field has been reported⁶ based on the data obtained by the USSR Mars 2 and 3 spacecrafts. The accurate determination of the moment of inertia factor⁷ and density models⁸ favor the existence of an Fe-rich Martian core. The radius of the core varies between 1500 and 2000 km depending on the composition of the core material. If the Martian core is composed of Fe, the formation of the core will take place relatively late in its evolution (approximately 2.5 billion years after its formation). However, it will remain molten throughout its history. On the other hand, if the Martian core is composed of FeS, core formation will take place about 1 billion years after formation. Again, it remains molten even at the present time.

Venus: Venus may actually have a weak magnetic field implied by the re-evaluation of the magnetic observations of the planet⁹. Because of the scarcity of data, the internal structure of Venus is not known. However, if Venus was formed in the present orbit, it probably has an evolution history and structure very similar to the Earth's because of the extreme similarity in size and mean density of the two planets. Based on this assumption, Venus is modelled to have an Fe-rich core similar in size to the

STATE OF THE CORES

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Earth's. Our thermal models indicate that the outer core of Venus is molten at present. The planet may have a solid inner core.

A comparison of the present-day models of the internal structure of the inner planets is shown in Figure 1. It should be noted that these are neither the unique nor the definite models of the interior of the inner planets. They only represent models favored by the available data and by earth comparisons. There appears to be some pattern between the intensity of planetary magnetism and the size and state of the cores of the Moon, Mars, Mercury and the Earth. Venus stands out as an exception, indicating the importance of other factors and the complexity of the origin of planetary magnetic fields.

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STATE OF THE CORES

M.N. Toksöz and A.T. Hsui

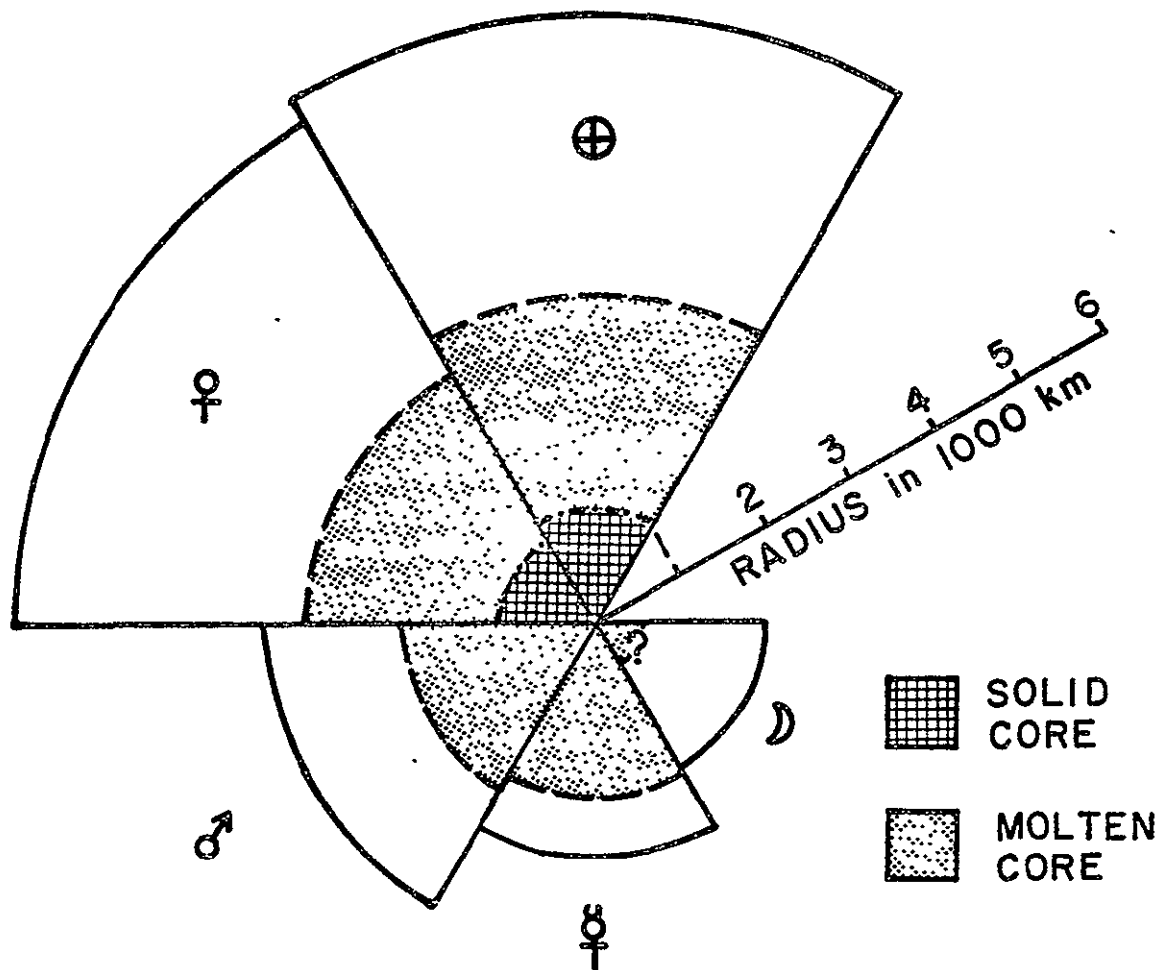


Figure 1. The present state of the cores of the terrestrial planets

MAGNETIC PROPERTIES OF FeNi ALLOYS - MICROSTRUCTURE AND REMANENT MAGNETIZATION MECHANISMS.

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Remanent magnetization in lunar samples and meteorites is carried by FeNi alloys. The lack of understanding of the NRM in these samples is directly attributable to the absence of an experimental data base which would provide basic magnetization data in terms of Ni content and thermal history. This need in terrestrial rocks has been recognized early by Nagata and colleagues and 25 years later, active research on synthetic FeTi oxides continues.

Results of the first systematic study, of weak field remanence acquisition, hysteresis properties, AF and thermal demagnetization, cryogenic cycling in zero field of room temperature acquired remanence, and magnetic viscosity, of the complete FeNi system are presented. The magnetic results are correlated with recognizable microstructure which is composition dependent and related to thermal history. Metallographic recognition criteria are presented which derive from optical and electron optical observations. Each mechanism of magnetization, active during a given thermal or dynamic history has a discrete imprint which is easily identified. These are based on sound physical metallurgy principles. The use of various carry over terms, from oxide experience such as SRM (shock remanent magnetization), CRM (chemical remanence) and pTRM (partial thermo remanence), in describing lunar and meteoritic NRM are incorrect and can lead to serious errors in evaluating the NRM. These objections will be explained in terms of the experimental results.

Specimens of FeNi alloys were acquired as spheres which were solidified in free fall using techniques developed for the NASA-zero-g solidification program. Therefore macroscopic demagnetization factors were constant and the serious problem of shape anisotropy was eliminated by the use of spheres. The FeNi system is divided into three groups: (a) the bcc FeNi alloys ($\text{Ni} \leq 28\%$), (b) the fcc FeNi alloys ($30\% \leq \text{Ni} \leq 50\%$), and (c) the fcc ordering alloys ($50\% \leq \text{Ni} \leq 90\%$). These groups are easily dealt with in terms of mode of occurrence in natural samples, thermal history and shock response, and magnetic behavior. Magnetic hardness is related to thermal and shock histories and is determined by microstructure induced. If an alloy has experienced a specific history which is unknown, but the event is discrete, then examination of the microstructure describes what mechanism of magnetization is operative and therefore specifies the analysis technique. This is possible because unlike oxides the alloy structures are indicative of thermal history, dynamic history, and subsequent thermal history.

For example, in the bcc FeNi alloys we show that the deformation twin density reflects the shock level but beyond the pressure transition-twinning is absent and a special shock structure results (this indicates a demagnetization has taken place) and as the peak impact levels increase the adiabatic decompression produces shock heating and 'self' anneal is possible. At the highest levels we have a case of a quench from $\geq 1000^\circ\text{C}$. All of these shock levels have been studied and the magnetic effects are different for each.

MAGNETIC PROPERTIES OF FeNi ALLOYS

P. J. Wasilewski

Indications are that the 'paleointensity' of a given metal sample in a given test field is a function of its prior history. For example, even though the $\alpha \rightarrow \epsilon \rightarrow \alpha_1$ transformation demagnetizes then remagnetizes a sample on a microsecond time scale in a given external field the NRM acquired may be 1-2 orders of magnitude less intense than that acquired by simple cooling. The misleading point about the remanence is that it appears stable, since the shock is a hardening process. Specific natural and laboratory examples of thermal and shock effects covering the entire FeNi system are presented.

The ordering fcc alloys are discussed in some detail as they are found in the ordered state in serpentinites and meteorites. Since thermomagnetic and slip induced anisotropy, both induced effects, are dominant anisotropies in these alloys, this research is of geophysical importance particularly since ordered and disordered states have such vastly different magnetic properties.

Sources of magnetic hardness in FeNi alloys are discussed and examples presented, e.g., twinning, shock structures, duplex $\alpha + \gamma$ structure, quench of shock heated structures. Results of analysis of remanent magnetization and hysteresis properties of chondrules from Allende, Chainpur, Bjurbole, Allegan, Ochansk and Pultusk are presented. The results indicate that metal size, shape, and distribution are different in different classes of chondrite meteorites. Results of thermal demagnetization in zero field of the NRM in iron meteorites and chondrites and the FeNi alloys are presented to demonstrate the behavior of the NRM in samples with different thermal and shock histories.

This review outlines the previous experimental research on FeNi alloys, presents new calibration for the FeNi alloys, and indicates how to analyze samples which contain FeNi alloys.

SHOCK INDUCED MAGNETIC EFFECTS IN FINE PARTICLE IRON.

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Since fine particles not observable by optical microscopy are difficult to characterize, since particle shape is extremely difficult to handle theoretically and experimentally, and since preparation of dispersed synthetic systems are near impossible we have adopted a model system to handle these problems. The system is based on the Copper (iron) precipitation alloy. Properly handled a narrow particle size distribution with a reasonably well defined interparticle spacing can be obtained. The iron precipitates are initially spherical in shape as long as $d \leq 1200 \text{ \AA}$ and maintain the antiferromagnetic fcc high temperature state when cooled to room temperature. Therefore, we have a metastable non-magnetic array of spherical iron particles waiting for the plastic deformation in the shock pulse to transform them to the ferromagnetic bcc state. Since the fcc $\rightarrow$ bcc transformation is a shock induced first order magnetic transition, an argument by analogy would suggest that the hcp $\rightarrow$ bcc transition is similar, (i.e., the remagnetization step in the bcc $\rightarrow$ hcp $\rightarrow$ bcc' transitions when bcc Fe is shocked at levels $> P_T$ (130 Kb). Therefore we can study magnetic effects during rapid magnetic transitions in controlled external fields.

The model system enabled us to establish the criterion for particle size and shape discreteness, i.e., when the magnetized ensemble is brought from saturation through $H = 0$ to the field $|H| = H_R$ (the remanent coercive force) and then reversed to pass through $H = 0$ to an equal and opposite field $|H| = H_R$ the magnetization hits on the upper portion of the major loop. This effect is not observed in coarse MD materials, Lodestones, packed $1 \mu\text{m Fe}_3\text{O}_4$ powder, coarse grained basalts, iron meteorites, chondrites, and Disko Island and Mt. Fuji basalts containing Fe. This test appears to be a necessary condition for SD behavior in rocks or for dispersions where magnetization reversal is ideally similar from particle to particle. In fact, for $|H|$ in excess of H_R , say H_1 , a return through $H = 0$ to $H = H_1$ should put M on the upper portion of the major loop and so on for H_2 etc.

In the spherical Fe samples this holds until $d \geq 1200 \text{ \AA}$ and true MD behavior is described, i.e., the path of M is somewhere within the upper and lower parts of the major loop. With electron microscopy we have observed substructure in particles from $250 \text{ \AA}$ to $1200 \text{ \AA}$. Maximum magnetic stability is observed in the $400 \text{ \AA}$ to $600 \text{ \AA}$ particles and $800 \text{ \AA}$ to $1100 \text{ \AA}$ particles, those $200 - 350 \text{ \AA}$ particles are least stable to AF demagnetization. The minimum in the R_H ($R_H = H_R/H_C$) vs θ (angle) curve is at 90° , i.e., parallel to the shock direction with the maximum perpendicular to the shock direction.

In the model system particle size, shape and interparticle spacing are controlled by varying the conditions of the precipitation anneal. The amount of transformed fcc iron, of a given size increases as the pressure increases and for a given pressure the amount transformed depends on the particle size.

SHOCK INDUCED MAGNETIC EFFECTS IN IRON

Wasilewski, P.

The maximum anisotropy for a given pressure is observed in the 400-600 Å particle. When transformed, the sign of the axial remanent vector always obeys the sign of the external field indicating that on the microsecond time scale of the transformation the magnetic systems sees the external field. The amount of remanent magnetization is considerably less than would be expected for TRM in the same material. The ratio of remanence after the shock to the saturation remanence is < 0.002 .

In the second generation experiments, we will evaluate directly the properties of shock induced remanence compared to thermo remanence, and evaluate the effects of recrystallization. These combined data should provide us with the experimental basis for evaluating shock effects in lunar samples and meteorites.

MODEL MAGNETOSPHERE OF MERCURY BASED ON MARINER 10 DATA

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From the first and the third encounter of Mariner 10 with Mercury's magnetosphere, a combined set of thirty minutes of directly observed magnetospheric magnetic field data was obtained. Direct observations also provided data on some boundary conditions of the magnetosphere such as the size of the magnetosphere and the magnitude of the magnetic field at the stagnation point of the magnetosphere. This information is invaluable and unique in studying the planetary magnetic field of Mercury. The analysis and interpretation of magnetic field data from Mariner 10 have been performed in two phases. Phase One Analysis used certain selected subsets of the magnetic field data, and usually disregarded boundary conditions of the magnetosphere. These boundary conditions are satisfied in Phase Two Analyses of Mercury's magnetic field. The confinement of planetary magnetic field lines inside a magnetosphere-like boundary requires a proper description of the magnetospheric external field. Prior to Mercury III encounter, the entire set of Mercury I data was used by Whang and Ness to study the model magnetosphere of Mercury using an image dipole representation for the external field. The result was rather surprising in that the inclusion of a magnetosphere-like boundary revised the estimated planetary dipole moment of Phase One Analyses by a factor of 2. After Mercury III encounter, further studies of Mercury's model magnetosphere using the entire combined set of Mercury I and III data revealed that the planetary field is very much distorted from a simple centered dipole, the planetary quadrupole and octupole are quite significant in magnitude compared with its dipole moment as shown in Table 1. The presence of a quadrupole moment repre-

Table 1. Planetary Magnetic Field Calculated From Image Dipole Models

Planetary Field	Data Base	Result
Centered Dipole (Whang & Ness, March 1975)	Mercury I	$D=2.3 \times 10^{22} \text{ G-cm}^3$
Dipole & Quadrupole & Octupole (Whang, March 1977)	Mercury I & III	$D=2.4 \times 10^{22} \text{ G-cm}^3$ $D:Q:O=1:0.4:0.3$

sents an axial offset of the apparent planetary dipole center by a distance

$$d \approx 0.5(Q/D) \times \text{planetary radius}$$

northward from the planetary gravitational center. The presence of an octupole moment indicates that the average radius of the source current system inside

MAGNETOSPHERE OF MERCURY

Whang, Y.C.

the planet is a significant fraction of the planetary radius. A spherical harmonic representation for the external field has also been used to study the model magnetosphere of Mercury. Table 2 summarizes various Phase Two models based on the entire combined set of Mercury I and III data, and their calculated deviations between model field and observed field.

Table 2. Models Based on All Mercury I & III Data

Planetary Field	Result	Deviation	$\sqrt{\sigma_x^2 + \sigma_y^2 + \sigma_z^2}$
Dipole & Quadrupole & Octupole (Whang, March 1977)	$D=2.4 \times 10^{22} \text{ G-cm}^3$ $D:Q:O = 1:0.4:0.3$	$\sigma = 11.4\gamma$ $\sigma_x = 10.3\gamma$ $\sigma_y = 13.0\gamma$ $\sigma_z = 13.0\gamma$	20 γ
Dipole & Quadrupole (Jackson & Beard, July 1977)	$D=2.44 \times 10^{22} \text{ G-cm}^3$ $D:Q=1:0.66$	$\sigma = 12.0\gamma$ $\sigma_x = 9.4\gamma$ $\sigma_y = 18.0\gamma$ $\sigma_z = 18.0\gamma$	24 γ
Offset Dipole (Ng & Beard 1978)	$D=2.37 \times 10^{22} \text{ G-cm}^3$ Offset=0.24 Rm	$\sigma = 13.1\gamma$ $\sigma_x = 8.6\gamma$ $\sigma_y = 19.2\gamma$ $\sigma_z = 19.2\gamma$	25 γ

Whang's quantitative three-dimensional model of Mercury's magnetosphere can identify the various regions of the magnetosphere, for example, the region of closed field lines, field lines connected with the tail current sheet, and field lines connected with polar regions. The trajectories of Mercury I and Mercury III at different regions exhibit very different plasma characteristics. Of particular interest is that the observation of high-energy charged particles are found to be along the field lines connected with the inner edge of the tail current sheet.

In Mercury's magnetosphere, the external field represents a significant fraction of the total field. Along the first half of Mercury I's trajectory, the total field was observed to increase from 40 to 98 gammas; whereas, the external field estimated by Whang's model is in the order of 30 gammas. Thus an accurate model representative of the external field becomes very important in data interpretation. The location and strength of the two current sources for the external field, the magnetopause and the tail current sheet, vary all the time responding to conditions of the solar wind and the magnetosheath. This temporal variation causes the fluctuation of the external field in the magnetosphere particularly in the region where the gradient of the external field is large. Making use of the model magnetosphere, the magnitude for the gradient of the external field is calculated on the trajectory plane of Mercury I and Mercury III. The area of large gradient positively correlates with the large fluctuations in magnetic field observed on Mercury I.

THE LUNAR SURFACE MAGNETIC SURVEY ALONG THE ROUTE OF LUNOKHOD-2

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The magnetic field measurements on the southern area of the Bay Le Monnier along the route of Lunokhod-2 were performed in Februar - April 1973 [1]. The Lunokhod-2 vehicle was not specially adapted to the magnetic measurements. The significant interferences were generated by the moving details of the vehicle, by the electromagnetic deviation and by the thermocurrents as well. As a result of the careful processing of all the body of data it succeeded to eliminate the deviation effects at least for two horizontal components. The measurement accuracy of the horizontal component was 3 ± 6 gammas and of the vertical one - ~ 10 gammas. The relative measurements were made with the resolution of 1 gamma. The contribution of external fields was taken into account as well under analysis of the lunar surface fields.

The magnetic field intensity is on the average 25 ± 30 gammas on the observed area of the Bay Le Monnier. The horizontal component is directed generally to the south-west. The transition region between the floor and the slope of the Le Monnier crater is characterized by the abrupt field gradients and the both field components are markedly decreased at the joint of the floor with the crater rim [2]. It is supposed this anomaly is caused by the edge effect of the steep contact of the magnetized basalt thickness with the non-magnetic (or slightly magnetized) rocks of the crater rim. The edge effect was distinctly observed at the edges of the tectonic break Straight Rille as well (fig. 1).

The magnetic anomalies which were observed when the Lunokhod crossed the small craters (with sizes of 100 ± 450 m) have typically maximum or minimum of the vertical component over the

The Lunokhod-2 magnetic survey

Ye. G. Yeroshenko



Fig. 1

crater center and maximum of the horizontal component over its edges (fig. 2). It is revealed the trend to increasing of the

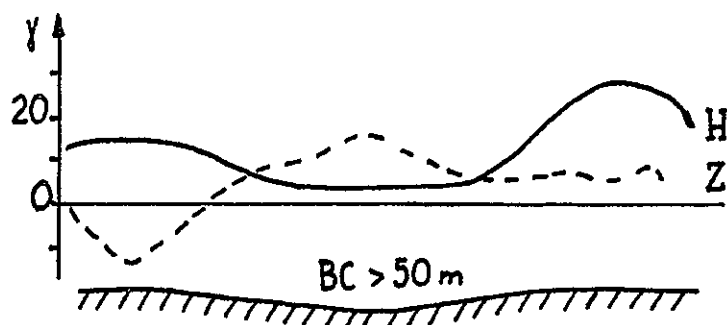


Fig. 2

anomaly amplitude with the increasing of the crater diameter as well. That kind of magnetic field features may be explained by the shock-explosion origin of the crater anomalies.

The magnetic anomalies caused by the edge effects were used for the estimation of the base rocks magnetization provided the uniform magnetization of the selenological structures [2]. The calculations of the edge effects of the ledges on oblique magnetization have indicated that the theoretical curves were in good agreement with the observed data if the basalt magnetization was of $10^{-5} \text{ G} \cdot \text{cm}^3 \cdot \text{g}^{-1}$ and the magnetization vector was

The Lunokhod-2 magnetic survey

Ye. G. Yeroshenko

directed down at the angle $30+45^\circ$ to horizon.

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THE GEOMAGNETIC SECULAR VARIATIONS IN HISTORIC TIMES

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In order to maintain a dipole field by a self exciting dynamo process a proper magnitude of non-axisymmetric motions with non-zero helicity are required in addition to the axisymmetric flows predominant in the liquid core. Although the concept of helicity has been originally exploited in the turbulent dynamo theory, it has been recently revealed that fluid motions with sufficient helicity are capable of regenerating the original axisymmetric field irrespective of the scale of motions. At present no preference seems theoretically attributable to either type of fluid motions, turbulence or a large scale motion, to maintain the geomagnetic dipole field. The type of the fluid motions might possibly differ according to the planetary bodies. This paper attempts to provide basic data, by reviewing the geomagnetic secular variations, for investigating what kind of fluid motion is most plausible within the earth's core.

The asymmetry of the fluid motion accompanies the asymmetric magnetic field through induction process within the axisymmetric field. If the fluid motion is of turbulent type, rapidly changing fields with small scales could be observed in the geomagnetic field. When the dipole field is subtracted from the observed fields, there remain irregular fields, called non-dipole field. If the irregular fields change their patterns in a very short time, the fluid motions in the core may also be regarded as changing their forms as quickly. In fact, there was a time when the non-dipole field was considered to decay in a hundred years or so and therefore it was regarded as the product of turbulent flows near the surface of the core.

During the past tenyears, spherical harmonic analyses have been conducted by several different groups for the historical data going back to the 17th century. Reconstruction of the non-dipole charts from these analyses disclosed that the life time of the non-dipole field was much longer than had ever been considered. The North American continent which is now covered with an intense positive anomaly in the vertical component was already under the positive anomaly in 1600 AD. This suggests that the American anomaly has life time longer than 400 years. Examination of the non-dipole charts indicates two distinct types of non-dipole anomalies. One is the migratory anomaly and the other is the anomaly staying almost at the same location. A negative anomaly existing off the west coast of Africa belongs to the first type. It was located in the Indian Ocean in the 17th century. Since then it has drifted westwards steadily with an average velocity of 0.28degrees/year and reached the present location. On the other hand, an intense positive anomaly that now covers the Asian continent with its center around Mongolia is the second type anomaly. Until 1650, no anomaly was recognizable in Mongolia. In 1700 a small anomaly was created and continued to grow to the largest non-dipole anomaly in the same place. An average rate of increase in the intensity amounted to 50 nT/year. The rapid growth of the anomaly could be interpreted by a development of localized fluid flows in the core. However, this phenomenon is understandable in a completely different way based on steady field models.

SECULAR VARIATIONS IN HISTORIC TIMES

Yukutake, T.

By making use of the spherical harmonic coefficients, it is possible to decompose the asymmetric parts of the geomagnetic field into two types of field, the drifting field and the standing field, with the constant intensity during the period of analyses. The decomposition was made of the coefficients since the 17th century up to degree 4. Reconstruction of the geomagnetic fields from these two types of the fields approximates the observed secular variations fairly well. It was demonstrated that not only the steady drift of the African anomaly but also the rapid growth of the Mongolian anomaly could be well accounted for by a simple superposition of the drifting and the standing fields. These may be an indirect proof of the long persistence of the two types of the fields, at least over several hundred years.

One of the differences noticeable between the two types of fields is lack of the sectorial term in the standing field for the degree two. i.e. $n=2$ and $m=2$. The amplitude of the standing field is only about 700 nT, while the drifting field has the amplitude of about 2000 nT. A similar feature is also seen in the sectorial term of degree 3. The standing component of $n=4$ and $m=4$ is neither large. The question whether or not the standing field generally lacks the sectorial terms except for the equatorial dipole component, which has the largest amplitude among the standing components, will be solved by examination of higher terms in future. Though the data available are insufficient yet, the remarkable lack of the sectorial terms for the degrees of 2 and 3 in the standing field might provide an important clue to the investigation of the physical process of creating the two kinds of fields.

The several centuries length is not sufficient to identify whether "the standing field" is strictly standing at the same locality or it is also subjected to a slow drift that may not be detected for several centuries observation. Regarding the equatorial component, however, observation evidently deviates from the drifting and the standing model even within several hundred years period. It was shown recently that introduction of a component drifting eastwards, instead of a standing field, better approximated the observed secular variations of the equatorial dipole.

Axisymmetric fields expressed by zonal terms of the spherical harmonics are known to have undergone considerable fluctuations in the past. On the other hand, there is no explicit data yet suggesting the variations in the intensity of the drifting and the standing fields. If they have changed their intensities, the observed secular changes in the harmonic components are supposed to deviate from the model computed on the steady field assumption. The deviations around the computed for some of the components amount to about 20 to 30 percent of the amplitude of the drifting field. This seems to place an upper bound on the amplitude variation of the drifting field.

Examination of the geomagnetic secular change during the past several hundred years indicated that the asymmetric fields that can be expressed by lower spherical harmonics had been stable. This might be favourable for considering the existence of considerably stable fluid motions with laterally large scale in the core.

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