Magnetic evidence concerning a lunar core

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Abstract—Measurements of the magnetic field induced in the moon while the moon is in the geomagnetic tail lobe field have been interpreted in terms of magnetic permeability due to the lunar free iron content. However, if the moon has a highly conducting core, the core will exclude magnetic fields with an apparent diamagnetic effect. Discussion of lunar chemical and thermal models establishes limits for plausible values of lunar magnetic permeability. These limits and the long period (3 hr to 4 days) magnetic response determine a maximum possible core radius. Measurements by two techniques are reported: (a) comparison of Explorer 35 and Apollo 12 data, and (b) observation of the dipole field configuration by orbiting subsatellites. Results are in moderate disagreement. The subsatellite observations (method b) require the existence of a core, but this measurement conflicts with measurements (method a) that allow, but do not require a small core. Until the discrepancy is resolved the only firm conclusion is that a maximum radius for a lunar core is about 580 km. A scenario for lunar core formation compatible with observations of atmospheric $^3$Ar and recent cool thermal history models is discussed; the core forms late in lunar history by slow descent of FeS. Remanent magnetism is attributed to subsurface dynamos rather than a lunar core.

INTRODUCTION

The moon’s induced magnetic moment, usually attributed to its ferro/paramagnetic properties, has been estimated from orbital and surface magnetometer measurements. Such studies (Parkin et al., 1973, 1974; Dyal et al., 1975; Russell et al., 1974a,b, 1975; hereinafter referred to as PDD73, PDD74, DPD75, R74a, R74b, R75) use data obtained in the high latitude tail lobes of the earth’s geomagnetic field. In the lobe regions the magnetospheric field is considerably larger in magnitude and less fluctuating than in other plasma environments, providing favorable conditions for observation of the long time scale lunar electromagnetic response. Interpretations of such data in terms of lunar magnetic permeability and free iron content have not considered the effects of a highly conducting lunar core. A core of Fe–Ni–S (Brett, 1973) composition would almost completely exclude magnetic field for times typical of major changes in the tail field with an apparent “diamagnetic” effect. Also, a partially molten asthenosphere could exclude magnetic fields to some extent for periods as long as an hour or two.

As two quantities (exclusion by the core and bulk permeability outside the

core) determine one observable property (the induced magnetic moment), we discuss chemical and thermal constraints on the global permeability. Observations of the induced magnetic field then place limits on the radius and conductivity of the lunar core. An additional problem arises in that measurements of the induced magnetic field obtained by different techniques are in disagreement. We report results using two different methods (subsatellite observations and comparison of Apollo 12 surface magnetometer data with Explorer 35 magnetometer data), discuss possible causes of the discrepancy, and determine what conclusions concerning the lunar interior are possible despite the observational difficulties.

THE LUNAR ELECTROMAGNETIC RESPONSE

For the purpose of analysis, we initially assume that the moon is immersed in a uniform external magnetic field, i.e., the geomagnetic tail lobe field. The magnetic permeability of the external environment is taken to be that of free space; plasma effects will be discussed later. The moon is modeled as a spherical, multi-layered body with the different layers having different values of magnetic permeability. In this case, the induced field observed external to the moon is a dipole field. If the external magnetizing field is denoted by $\vec{H}$, and the total magnetic field observed at the lunar surface due to the external and induced fields is denoted by $\vec{B}$, then

$$\vec{B} = H_x(1+2G)\hat{x} + H_y(1-G)\hat{y} + H_z(1-G)\hat{z} ; \quad G = G_\mu - G_{TE} \quad (1a,b)$$

In this equation ALSEP coordinates are used ($\hat{x}$ is upward from the lunar surface, $\hat{y}$ and $\hat{z}$ are tangent to the lunar surface). $G$ is a parameter (notation of PDD74) for the total induced dipole with contributions from both lunar magnetic permeability ($G_\mu$) and transient exclusion of the external magnetic field from the conducting lunar interior ($G_{TE}$, transverse electric mode wave). $G_\mu$ can be approximated by summing the contributions from each spherical shell (PDD74, Eq. (3)) $\tilde{\mu}$ the relative magnetic permeability, $\mu_i$ of each layer is close to one. Thus,

$$G_\mu = \sum \frac{1}{3}(\mu_i - 1)(\lambda_i^1 - \lambda_i^3), \quad (2)$$

where $\lambda_i$ (or $\lambda_{ni}$) is the ratio of the outer (or inner) radius of a shell to the lunar radius and the sum is computed over all layers. If we define $G_c(\tau)$ to be the conductive lunar response to a step function increase in the external magnetic field ($0 \leq \tau < 0$, $H$ for $\tau \geq 0$ where $\tau$ is time after the step function increase), then the value of the induced field is approximated by integration over past changes in the magnetic field

$$B_x(t) = (1+2G_\mu)H_x(t) - \int_0^t 2 \frac{dH_x(t-\tau)}{dt} G_c(\tau) d\tau \quad (3)$$

This approximation is valid if both magnetic permeability and TE mode induction fields are small; squares of small quantities can be ignored. Magnetic permeability effects are small and at the low frequencies investigated in this study the induced fields are also small.
From Eqs. 1 and 3 the value of $G$ is determined by

$$G(t) = (B_r(t) - H_r(t))/2H_r(t) = \int_0^{\tau} \frac{1}{H_r(t)} \frac{dH_r(t - \tau)}{d\tau} G_R(\tau) \, d\tau$$  \hspace{1cm} (4a)$$

where

$$G_R(\tau) = G_\mu - G_c(\tau)$$  \hspace{1cm} (4b)$$
is the total lunar step function response including the effects of magnetic permeability.

For the case of a two-layer model with a non-conductive mantle and a core of finite conductivity, $\sigma$, and a ratio of core radius to the lunar radius of $\lambda$, $G_c(0) = \lambda^2/2$ whereas for $\tau \to \infty$ the value of $G_c(\tau)$ approaches zero. For $\sigma_c$ less than $10^{-1}$ mho/m, exclusion decreases rapidly with time; thus, magnetic permeability measurements obtained more than a few hours after major changes in the external magnetic field would truly represent lunar magnetic permeability. Alternatively, $\sigma_c$ as large as $10^4$ mho/m would result in virtually complete exclusion of the tail field from the lunar core for the duration of a tail crossing.

**Observations**

As the induced lunar fields are typically quite small, considerable care must be devoted to data analysis. As previous measurements of lunar magnetic permeability have differed (PDD74, R74b), we employ two different methods of determining the lunar electromagnetic response: (a) comparison of Apollo 12 and Explorer 35 magnetometer data; and (b) observation of the dipole field configuration by orbiting Apollo 15 and 16 subsatellites. Our study confirms that the methods do give significantly different results. We now describe our analysis methods.

The technique used in this and prior studies of lunar magnetic permeability is to choose magnetically quiet periods (less than $1^\circ$ change during an orbit for the subsatellite technique) in the tail lobes, wait some time (3 hr in our case) for the short period response ($G_c(\tau)$ for small $\tau$) to decay, and then obtain an estimate, $G_{obs}$, of the induced field. This method greatly reduces the short period contribution to $G_{obs}$ from $G_c(\tau)$. Also, if field values immediately prior to observation periods were equally likely to have been smaller or larger, the average from the short period response is zero. Averaging all measurements of $G$ over observing periods, $\Sigma$ (Eq. (5a)), substituting Eqs. (1b) and (4a) into Eq. (5), and interchanging order of integration, a weighting function, $W(\tau)$ is defined such that

$$G_{obs} = \int_\Sigma G(t) \, dt \int_\Sigma dt = \int_0^{\tau} G_R(\tau) W(\tau) \, d\tau$$  \hspace{1cm} (5a,b)$$

$$W(\tau) = \int_\Sigma \frac{1}{H_r(t)} \frac{dH_r(t - \tau)}{d\tau} \, dt / \int_\Sigma dt.$$

In the magnetosphere only a few time periods were suitable for obtaining accurate measurements of $G$ due to the presence of various error sources such as plasma diamagnetism; small errors in $H$ or $B$ result in large errors in determining
\(G\) (which depends upon the difference between \(\bar{B}\) and \(\bar{H}\)). On the other hand, when using Eqs. 5 and 6 to estimate the contribution to \(G_{\text{obs}}\) from \(G_R(\tau)\) as a function of \(\tau\), it is possible to tolerate considerable error in \(\bar{H}\) provided that the errors occur at large values of \(\tau\) (several hours or more) for which \(G_R(\tau)\) is small and slowly varying. Using the solar magnetospheric \(X\) component of the field at two hour intervals, making no corrections for changes in field orientation or plasma sheet diamagnetism, and replacing missing data by previous values, we have calculated the weighting function, \(W(\tau)\), for the particular observing periods, \(\Sigma\), used to estimate \(G\) from subsatellite data (see Fig. 1). The value in the interval 0-2 hr is \(-0.011\), indicating that the short period response should be suppressed and that there was a slight average tendency for the field magnitude to decrease within 2 hr before observing periods. A slight decrease in the external magnetic field on the order of 1 hr before observations will result in the mantle opposing the decrease in the external magnetic field. The effect due to the recent decrease is in opposition to the effect due to the increase in magnetic field upon

\[
W(\tau) = \int \frac{1}{H(t)} \frac{dH(t - \tau)}{dt} dt / \int dt
\]

\[
G_{\text{obs}} = \int_0^\infty G(\tau) W(\tau) d\tau
\]

Fig. 1. The integral of the weighting function, \(W(\tau)\), as a function of lag time.
entering the tail lobes at some earlier time in the past. As recent history dominates the contribution from the mantle, the expected bias due to mantle conductivity for these particular subsatellite observations would be in the sense of making a core look smaller (rather than larger as might intuitively be expected). However, this conclusion is not absolutely certain due to the relatively large time intervals used in calculating $W(\tau)$.

Our first method of estimating $G_{obs}$ is based on comparison of Explorer 35 and Apollo 12 magnetometer data. The Explorer 35 spacecraft measures only the external magnetospheric tail field, $\vec{H}$, while the Apollo 12 surface magnetometer measures the sum of external field, induced field, and local remanent field; both instruments may at times measure fields due to nearby plasma currents. The data were provided by B. Smith of Ames Research Center in the form of low-pass filtered averages at ten minute intervals. The data base was restricted to include only quiet periods within geomagnetic tail lobes; any periods during which the Solar Wind Spectrometer observed plasma or were obviously contaminated by plasma sheet diamagnetism were discarded. (The plasma detector was not sufficiently sensitive to detect typical plasma sheet particle fluxes, but did eliminate some data near the boundary between the solar wind and geomagnetic tail.) Any data following within 3 hr of entering the tail lobes were eliminated, as were any data immediately following large field transients. A 3 hr delay was chosen so that for a low conductivity mantle and an asthenosphere of 700 km radius with conductivities of $10^{-1}$ mho/m or less the observations would represent primarily magnetic permeability effects. In particular, the following procedures were followed in an attempt to eliminate plasma diamagnetism and the short period response.

First, preliminary estimates of the local remanent magnetic field were obtained from tail lobe data by a straight line fit of Apollo 12 to Explorer 35 data; the intercept at the value of zero for the Explorer 35 field was taken as the local Apollo 12 site field. The values obtained were within 0.05 $\gamma$ of those reported by Dyal et al. (1973). Then, averages and standard deviations of the 10 min data were computed at 1 hr intervals. Next, averages of both these quantities were computed at 2 hr intervals; the difference of the 1 hr averages for each 2 hr period was also computed. The following restrictions were then applied to the 2 hr averages of Explorer 35 data and Apollo 12 data after removal of the remanent field contribution. (1) All data for which the sum of the standard deviations of all three components was greater than 12% of the field magnitude at Explorer 35 were discarded; similarly, if the sum of the absolute values of the 1 hr differences of all three components was greater than 24% of the field magnitude at Explorer 35 the data were discarded. (2) The field magnitudes at both instruments were required to be greater than $6\gamma$. (3) The field direction was required to lie within 15$^\circ$ of the plane of the lunar equator. Despite these selection criteria, some data that obviously did not fit the model of constant external field plus aligned dipole field remained. (4) To eliminate such points the $y$ and $z$ components of the two data sets were compared; values differing by more than 1.9$\gamma$ were discarded (only the $x$ component was used to estimate magnetic permeability). (5) For the same reason,
periods during which the field components for both instruments did not have a similar temporal behavior (based on visual comparison of plotted data) were eliminated.

Thirty-seven averages survived these criteria (Fig. 2). As random error in both data sets is expected to be approximately equal, the parameter $G_{\text{obs}}$ is determined by linear regression of $B_x$ as a function of $H_x$ and $H_y$ as a function of $B_x$; the estimated values of $G_{\text{obs}}$ were averaged and the standard errors added (same technique as PDD74). An estimated value $G_{\text{obs}} = 0.009 \pm 0.009$ was obtained. (The estimate and error (0.004 ± 0.002) obtained by PDD74 suggests a more successful elimination of periods contaminated by plasma diamagnetism, possibly due to their criteria rejecting data with short period (~1 min or more) fluctuations. However, problems involving statistical independence of error estimates may arise if a source of error (i.e., plasma diamagnetism) remains present during several short period averages.) The value of $G_{\text{obs}}$ and the estimated errors depended considerably upon what criteria were tried. The best estimate that could be obtained was $G_{\text{obs}} = 0.005 \pm 0.005$ based upon 19 averages. The estimated values of $G_{\text{obs}}$ were in the range 0.005 to 0.009; inclusion of periods contaminated by plasma diamagnetism increased $G_{\text{obs}}$. In addition, smaller values of $G_{\text{obs}}$ were associated with smaller error estimates; in no case was the one standard deviation

![Fig. 2](image_url)

Fig. 2. The induced lunar magnetic field as determined by comparison of Explorer 35 data, $H_x$, with Apollo 12 LEM data, $B$. The average value of the local remanent field at the Apollo 12 site has been subtracted. The slope of the best fit line is twice $G_{\text{obs}}$. 
lower limit (estimate with error subtracted) less than 0.000. A similar mean standard deviation relation between the estimates of PDD73 and PDD74 also occurs. This strongly suggests that a variable source of bias is present at times. Plasma diamagnetism due to either the plasma sheet or low energy particle events (Hardy et al., 1975) should increase the measured value of \( G_{\text{obs}} \) and thus explain the observed relation between estimate and standard deviation. The apparent increase in \( G_{\text{obs}} \) would be due to decreased plasma density on magnetic field lines connected to the Moon; the counterbalancing magnetic pressure increase would be observed by the Apollo 12 surface magnetometer. As the lunar plasma environment is often unchanging for periods of several hours, an additional problem is that averages taken at much shorter time intervals will not be statistically independent. Thus, even statistical error estimates obtained in prior work may be too small.

Our second method of estimating \( G_{\text{obs}} \) utilized Apollo 15 and 16 subsatellite magnetometer data obtained in lunar orbit at a nominal altitude of 100 km. A study of the global lunar magnetic field (R74b, R75) indicated that the permanent lunar dipole field was negligible (maximum value of \( 1.1 \times 10^{-9} \, \text{T cm}^3 \)) and observed a substantial induced dipole moment (\( G = -0.009 \)) opposed to the external magnetic field (PDD73, PDD74, DPD75, and our first method find an aligned moment). The measurements were obtained in the tail lobes where the magnetospheric field is directed either towards (north lobe) or away from (south lobe) the solar direction. A detailed analysis of the observational techniques, uncertainties, and problems of data analysis is given in R74b; we will review the prior analysis and then discuss the significant differences of our present approach.

For the purposes of analysis, orbits were chosen during which the magnetospheric field exhibited no large sudden transients or time dependent behavior. This criterion did not exclude orbits for which sudden changes had occurred some time before the beginning of the orbit. For each orbit the average value of the magnetic field and linear trends were subtracted from the data. The residual fields were then converted into a local lunar coordinate system where \( B_R \) is directed radially outward, \( B_T \) is tangential to the orbit plane (eastward), and \( B_P \) is perpendicular to the orbit plane (northward). This technique changed the profile of a dipole field as the dipole field has an average component that was removed.

To estimate the strength of magnetic induction, induced dipole field components were normalized by dividing by the magnitude of the average magnetic field component in the orbit plane. Azimuth angle, \( \theta \), was measured eastward from the subsolar projection of the magnetic field onto the orbit plane and data were averaged in 5° bins. To combine the north tail lobe data with the south lobe data the results were averaged by subtraction rather than addition as the geomagnetic tail field in the south lobe is opposite in direction to the north lobe field. This procedure also eliminated any possible contribution from a permanent lunar dipole. A Fourier analysis as a function of azimuth angle was then used to estimate the dipole moment; the cosine component of \( B_R \) and the sine component of \( B_T \) provided two independent determinations. We note that the component of the external magnetic field normal to the orbit causes induced field components
that are also normal and constant along the subsatellite orbit, and that no approximation is involved in not using the $B_p$ component.

In our present analysis, the same procedure was adopted for calculating and normalizing $B_n$, $B_r$, and $B_p$. However, rather than averaging data from all orbits as a function of azimuth angle, the $B_n$, $B_r$, and $B_p$ measurements were least squares fit to a model $B_i = B_n \cos \theta + B_r \sin \theta$ for each orbit. Orbits for which more than 25% of the data was missing were not used for this purpose. The induced dipole is obtained from the $B_{nR} - B_{nT}$ estimate. For each orbit an error estimate is computed from the root mean square average of the following quantities: (1) the discrepancy in $G_{\text{obs}}$ as determined from $B_{nR} + B_{nT}$; (2) the apparent dipole at right angles to the average field, $B_{nR} + B_{nT}$; and (3) the orbit periodic variation of the component perpendicular to the field, $B_{nR} + B_{nT}$.

The present study also discards orbits within 3 hr of leaving the plasma sheet or magnetosheath and orbits contaminated by unusually large low energy particle fluxes. The plasma environment is determined from observations by the sensitive SIDE detector (rather than estimated from magnetometer data or the less sensitive SWS detector). Orbits for which the error estimate of $G_{\text{obs}}$ is greater than 0.025 are also discarded (about 10% of the remaining data); such discarded estimates are strongly aliased by temporal variations of the tail field. Finally, the values of $G_{\text{obs}}$ are averaged over all the orbits; the previous work averaged equally north lobe and south lobe estimates. The value obtained by this procedure is $G_{\text{obs}} = -0.0093 \pm 0.0023$ based upon all data from the Apollo 15 and 16 subsatellites that met the stated criteria (27 orbits). This value is almost identical to that of R75 but has smaller statistical errors. In addition, we have computed the average value of the error quantity, $B_{nR} + B_{nT}$, over the same 27 orbits and obtained a value 0.0023. If the external magnetic field had a dipole configuration, this quantity would be zero. In fact, it is equal to the random error we have estimated from scatter in our estimates of $G_{\text{obs}}$. This consistency is useful when assessing the possibility of errors, e.g., biases due to plasma, that would affect differently the magnetic fields observed near the subsolar and anti-solar orbit locations ($B_R$ component) from those observed near the limbs ($B_T$ component).

To illustrate the nature of our observations, estimates of $G_{\text{obs}}$, with associated errors and the moon–sun field component for one lunar passage through the tail are shown at one orbit intervals (2 hr) in Fig. 3. The moon enters the south lobe at orbit 1065; $G_{\text{obs}}$ for this orbit was not included in the overall average because of the immediately preceding plasma environment. The north lobe is entered between orbits 1070 and 1071; thereafter the magnetic field is almost constant until orbit 1087. The time dependence of $G_{\text{obs}}$ can be estimated during this quiet period. The increase at orbit 1070 dominates the conductive response; the prior increases and decreases in the south lobe and magnetosheath regions are somewhat smaller in magnitude and tend to cancel. From orbits 1074 to 1086 $G_{\text{obs}}$ appears to decrease with time, but a partially conducting core should cause an increase of $G_{\text{obs}}$ with time. A constant value of $G_{\text{obs}}$ over a 30 hr period (a highly conducting core) is almost consistent with the indicated error bars.

Averaging all values of $G_{\text{obs}}$ assumes (in effect) the time-independent behavior
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![Graph showing magnetic field data](image)

**Fig. 3.** The Earth-Sun component of magnetic field as observed on a typical tail passage by the Apollo 15 subsatellite; values are shown at two hour intervals. For orbits that meet the data selection criteria, $G_{in}$, and estimated errors are shown.

of a perfectly conducting core surrounded by a perfectly insulating mantle. A time dependent analysis is also possible. Equation 4 can then be approximated in finite difference form using a step size in time of $\delta$,

$$G(t) = \sum_{n=0}^{N} G_{R}((n+1/2)\delta)(H_{c}(t-n\delta) - H_{c}(t-(n+1)\delta))/H_{c}(t),$$

where the sum over past times is extended one day into the solar wind where the value of the external magnetic field is typically low. At times even further in the past, the external field is assumed to have been zero. For each orbital measurement a value of $G(t)$ is obtained which is in effect a two hour average value. Using $H_{c}(t)$ at two hour intervals, predictions of $G(t)$ for each orbit can be made if $G_{R}(\tau)$ is known. $G_{R}(\tau)$ is approximated by the functional form of three different constant values for $\tau = 0-12$ hr, $\tau = 12-30$ hr, and $\tau \geq 30$ hr. The predictions of $G(t)$ are least squares fit to the observations of $G(t)$ as a function of the three free parameters for the value of $G_{R}(\tau)$ in each time interval. The estimates are shown in Fig. 4; the estimate assuming constant $G_{R}$ (average value) is similarly plotted as the response from 0 to 100 hr. Also shown is the calculated response of a 500 km radius lunar core surrounded by an insulating mantle for core conductivities of $10^{-4}$, $10^{-1}$, and $10^{-3}$ mho/m with no correction for magnetic permeability. Magnetic permeability simply adds a constant positive contribution independent of $\tau$ to the response (Eq. (3)). The increase of observed $G_{c} - G_{R}(\tau)$ with time suggests a tendency for the external magnetic field to partially penetrate the conducting core for lag times greater than 30 hr, but a perfectly conducting core is also consistent with the indicated error bars. Thus, the tail crossing shown in Fig. 3 that shows no
tendency for \( G_{\text{obs}} \) to increase at large lag times is not typical of the data as a whole in this regard.

**Possible Error Sources**

The previously reported disagreement between estimates of \( G_{\text{obs}} \), obtained using surface data (PDD74, DPD75) and subsatellite measurements (R74a, R74b, R75) is confirmed by our study. A variety of possible error sources might affect these measurements; a list of some possible explanations is shown in Table 1. None of the proposed explanations reconciles the discrepancy in a satisfactory or convincing fashion. Plasma diamagnetism will, however, create biases that influence both surface and subsatellite observations.

Schubert et al. (1975) demonstrate partial plasma confinement of TE mode waves at frequencies greater than 8 mHz in the tail lobes (no high sensitivity plasma data were available for use in this study). As our measurements are at periods of many hours, plasma inertial effects due to propagating magnetospheric waves (which are significant at periods on the order of the time for an Alfvén wave to travel a lunar radius, i.e., seconds to minutes) are unimportant, but time independent thermal effects and inertial confinement of the induced dipole by anti-solar plasma flows may be significant.

An estimate of plasma diamagnetism is obtained by magnetic pressure balance
Table 1. Possible causes of discrepancy.

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<th>Quality of explanation</th>
<th>Effects</th>
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<td>Drawbacks</td>
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<td>Local magnetic permeability anomalies</td>
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<td>Non-zero contributions from a permanent lunar dipole field</td>
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<td>Statistical fluctuation</td>
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<td>Short period TE mode response</td>
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<td>Observer data selection criteria</td>
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<td>Magnetometer scale factor error</td>
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<tr>
<td>Several of the above</td>
<td>?</td>
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between a plasma and vacuum; thus,

$$\mu_p/\mu_0 = \sqrt{1 - 2\mu_0 P/B_0^2},$$

where $P$ is the plasma pressure, $B_0$ is the vacuum field value, and $\mu_p$ is the "magnetic permeability" of the plasma. For a plasma density of 0.05 cm$^{-3}$ (assumed if SIDE detects no plasma) or of 0.3 cm$^{-3}$ (a typical tail value) and a temperature corresponding to 15 eV, we obtain $\Delta G = 1/3(\mu_p/\mu_0 - 1) = 0.0005$ for no observable plasma and $\Delta G = 0.003$ otherwise, where $\Delta G$ is the apparent increase in lunar magnetic permeability as observed by the subsatellite (this is a factor of two overestimate as no currents flow over the lunar nightside surface for anti-solar plasma fluxes). Naive estimates of plasma diamagnetism increase (rather than explain) the discrepancy between subsatellite and surface observations. In a previous work (Goldstein and Russell, 1975) we estimate that an equally important contribution to subsatellite observations resulted from depletion of plasma on magnetic field lines connected to the moon, and the consequent increase in magnetic field required for pressure balance. In fact, these computations used a subsatellite altitude of 200 km rather than the correct value of 100-120 km; in addition, finite ion gyroradius effects cause the subsatellite to be outside the region of depleted plasma, an even smaller fraction of the time. The effect upon subsatellite observations is therefore only about 0-20% of other plasma diamagnetic effects. Finally, there is the consideration that photoelectrons
from the lunar surface will tend to cancel ordinary plasma diamagnetic effects. For example, if the velocity distribution of photoelectrons emitted from the lunar surface were identical to that of the incoming electrons associated with low energy particle events, isotropy of the electron distribution would imply no diamagnetic effects at the lunar surface. Goldstein and Russell (1975) show that for very low plasma densities the escaping photoelectrons will be hotter than the incoming electrons, whereas for typical tail densities (0.1–0.5 cm⁻³) the temperatures will be approximately equal. In view of these uncertainties, errors in $G$ of as much as ±0.0015 for densities of 0.3 cm⁻³ to errors of ±0.0005 for densities of 0.05 cm⁻³ due to thermal effects upon subsatellite observations might be expected. The effect upon $G$ determined by comparison of Apollo 12 and Explorer 35 data would be larger by a factor of 5/2 as the Explorer 35 magnetometer measures smaller magnetic fields that are unaffected by plasma depletion on field lines passing through the moon.

Compression of the induced dipole by the anti-solar flow of low energy particle events (Hardy et al., 1975) may also occur. For an extremely crude model of an MHD flow past the moon (cylindrical moon, particles emitted from the night side of the moon so that the flow equations have upstream-downstream symmetry, $\vec{V}$ and $\vec{B}$ parallel far upstream, small perturbations) an equation for $\vec{B}$ in the plasma flow can be derived:

$$\left(1 - \frac{V_0^2}{C_A^2}\right) \frac{\partial^2 \vec{B}}{\partial x^2} + \frac{\partial^2 \vec{B}}{\partial y^2} = 0.$$  \hspace{1cm} (8)

The far upstream values of $\vec{B}$ and $\vec{V}$ are in the $x$ direction, the axis of the cylindrical moon is in the $z$ direction, and only velocity and magnetic components in the $x$, $y$ plane are allowed. The derivation of the above equation is not described as we intend to investigate a more realistic model at a later date. Qualitatively, when the upstream flow speed, $V_0$, is greater than the upstream Alfvén speed, $C_A$, the induced dipole is excluded from the plasma and compressed to the surface. For flow velocities on the order of 140 km/sec, $n = 0.3$ cm⁻³, and $B = 10\gamma$, the ratio of length scales,

$$\sqrt{1 - \frac{V_0^2}{C_A^2}},$$

is 0.936. If $n = 0.005$ cm⁻³ the ratio of lengths is 0.989. Assuming that magnetic field scales as the cube of length, these numbers suggest magnetic compressions on the order of 3–18% for the suggested range of density. Plasma inertial effects are therefore a possible source of bias, but are insufficient to greatly modify our conclusions or explain the discrepancy between the observation techniques.

The above discussion suggests that no significant error should occur when SIDE detects no plasma. Despite the narrow view angles of the SIDE detectors, both the Apollo 14 and Apollo 15 detectors usually agree as to whether plasma is present; also, a high correlation exists between detection of plasma and the East–West component of the interplanetary magnetic field (Hardy and Freeman, personal communication). We conclude that SIDE is an effective monitor of the plasma environment except during periods of solar directed plasma sheet flow.
These conclusions would not apply if a substantial high energy particle population was present during periods the moon was on field lines passing through the plasma sheet, but both SIDE and PFS electron data indicate this occurs only about 10% of the time. We have used SIDE data to determine these periods; future studies should also use PFS electron data.

Plasma diamagnetic effects have also been analyzed by separately averaging only those 12 orbits during which no plasma was observed. For the data reported here we obtain a value of $G_{\text{obs}} = -0.0062 \pm 0.0045$. The apparent diamagnetic effect is more than two times larger than estimated above but is not statistically significant. Other observational evidence (Goldstein and Russell, 1975) shows no plasma effects such as decreased magnetic fields on field lines not connected to the moon or frontside-backside asymmetries that would be expected from anti-solar plasma flow. In general, plasma effects such as inertial compression plasma depletion, or thermal effects can be expected to have a non-dipole configuration. The consistency between estimates obtained from $B_r \cos \theta$ and $B_r \sin \theta$ (see observations) is an important argument for the reliability of the subsatellite data.

The conductive response of the lunar mantle might also affect our results, the 0–12 hr response shown in Fig. 4 is more negative than the average response at periods greater than 12 hr (but the difference is not statistically significant). Variations in the magnetic field shortly before observation are typically random and of either sign, thus TE mode effects due to changes immediately before observation periods will also be random and contribute no bias. We have chosen a 3 hr waiting period after increases in the magnetic field before obtaining measurements; thus, non-random effects due to mantle conductivity for times greater than 3 hr would be expected to contribute a core-like signal to the measurements. If a small fraction melt of the lunar interior is assumed on the basis of seismic data (Nakamura et al., 1974), then conductivities on the order of 0.01 to 0.1 mho/m are expected (Waff, 1974). The Cowling diffusion time for magnetic field scales as conductivity multiplied by the square of radius. Examination of Fig. 4 showing the response of a 500 km lunar center of conductivity 0.1 mho/m at a period of 3 hr shows effectively no contribution. The Cowling time of a 700 km asthenosphere with conductivity 0.05 mho/m would be the same but the response would be larger by a factor of 7.5 due to the larger radius. However, only a small portion of our data is obtained at lag times as short as 3 hr (Fig. 1), so that bias from this source should be negligible. In fact, as discussed earlier, there was a slight tendency for $H$ to decrease within 1 hr before the observation periods used; due to this effect we estimate a negligible mantle bias which would give the appearance of a smaller core (increase the observed value of $G$).

Finally, we note that instrument temperature dependent effects in the subsatellite data have been calibrated on the ground, are very small, and have been corrected in data processing. Neither will a lunar ionosphere affect magnetospheric measurements (Goldstein and Russell, 1975). The subsatellite data are not subject to magnetometer scale factor error that might bias results obtained by comparing measurements from different magnetometers. Summing up, the effect
of permanent magnetism, plasma diamagnetism, and mantle conductivity upon our data has been investigated observationally. The data are of insufficient accuracy and quantity to investigate all sources of bias simultaneously; instead, each source of error has been separately investigated. Two techniques have been used to investigate plasma diamagnetism; one gives a null result and the other shows a moderate effect that is of debatable statistical significance and is somewhat larger than is expected on theoretical grounds. We conclude that there may be some biases in our measurements and estimate a maximum plausible error in the value of $G_{\text{obs}}$ of $\pm 0.003$ due to mantle conductivity and plasma diamagnet-

![Graph showing core radius in kilometers and $G_\mu$ values]

Fig. 5. Values of $G_{\text{obs}}$ obtained by several studies. Sources of the measurements are: (a) Dyal et al., 1975; (b) Parkin et al., 1973. (c1 and c2) This study using Apollo 12 and Explorer 35 data, and (d) this study using subsatellite data. Solid brackets are statistical errors. Dashed brackets show possible effects of bias on subsatellite observations. The curves determine the possible radius of a perfectly conducting lunar core surrounded by a perfectly insulating mantle as a function of $G_{\text{obs}}$ for specified values of lunar magnetic permeability.
ism. This is insufficient to bring our data into agreement with other work (PDD74 and DPD75). After correcting for estimated biases and errors, a negative value of $G_{obs}$ for subsatellite data results. A value of $G_{obs}$ less than zero requires the existence of a lunar core.

Our estimate of induced moment using comparisons of Apollo 12 and Explorer 35 data suggest substantial influences of plasma diamagnetism and/or spatial gradients between the two magnetometers. However, the Apollo 15 and 16 comparison (DPD75) used SIDE data to eliminate periods when plasma was observed; also, plasma depletion on field lines connected to the moon would not affect these measurements. Several other sources of error which might affect surface magnetometer observations are: local permeability anomalies, compression of local remanent fields by ambient plasma (Dyal et al., 1972; Siscoe and Goldstein, 1973), and magnetometer drifts (scales, not just offsets). However, studies using different data sets (Explorer 35 and Apollo 12 LSM data, PDD74, Apollo 15 and 16 LSM data, DPD75) are in good agreement (see Fig. 5). We conclude that the cause of the discrepancy between the surface and subsatellite measurements is unknown and that no definitive reason exists for choosing between the measurements.

**DISCUSSION**

Contributions to the very long period lunar response would be expected from both lunar magnetic permeability and a highly conducting core. PDD74 show that only ferromagnetism due to free iron below the Curie temperature is important; they also plot the Curie temperature as a function of depth (pressure) beneath the surface. The lunar interior temperature profile that determines the Curie temperature depth can be established from the electrical conductivity of the lunar interior (Sonett and Duba, 1975; Dyal et al., 1974, 1975). Using different analysis methods for data obtained in different plasma environments, the two groups report interior conductivities that are in reasonably good agreement. The corresponding temperature estimates suggest a possible depth of the Curie point of about 250 km; however, compositional changes at depth may affect these estimates. Thermal history models suggest a depth of 350 km (DPD75, Hubbard and Minear, 1975).

The ferromagnetic contribution from free iron at depths less than 60 km is severely reduced by the drastic depletion (Fuller, 1974) of free iron in the differentiated lunar crust. If the upper portions of the lunar interior at depths less than the present Curie point underwent total or large fraction melting, then the amount of free iron remaining in the upper mantle might be so small as to have no measurable effect upon lunar magnetic permeability. For the purpose of estimating the maximum size of a lunar core from the induced magnetic moment, a maximum plausible value of lunar magnetic permeability must be assumed. If the amount of free iron in the upper mantle below 60 km is assumed to be typical of whole moon values before differentiation, then plausible free iron concentrations are from 4 to 6% by weight (Ganapathy and Anders, 1974), about 1.9% (Wänke et al., 1973), or even lower if troilite accreted with iron. (Ganapathy and Anders
determine lunar Fe from magnetic permeability measurements of Dyal and co-workers, so circular reasoning is involved. However, the PDD73 constraint on total lunar Fe is determined by lunar density with uncertainties due primarily to assumptions of lunar mineral composition, not free iron content.) We adopt a lunar model with no free iron at depths less than 60 km, and with a constant percentage by weight, \( q \), from 60 km to the Curie depth. The magnetic permeability of the upper mantle can then be approximated by \( \mu = 1 + 0.0153 \ q \) (PDD74). Using Eq. (2), \( q = 6.0 \), and a Curie depth of 250 km, the maximum likely value of \( G_\mu \) is 0.008. For the rest of this discussion, it is assumed that \( G_\mu \) lies in the range 0.000 to 0.008; large estimates would result in somewhat larger estimates of core size. A Curie depth of 350 km and as much free iron as 6\% would increase \( G_\mu \) to 0.012.

An estimate of lunar core size is now possible. For a model with an insulating mantle and a perfectly conducting core, values of \( G \) for fixed values of \( G_\mu \) are shown in Fig. 5 as a function of core radius; also shown are measurements from this and other studies. (Preliminary results of Wiskerchen et al. (1976) establish a larger upper limit on core radius than either of the magnetospheric studies if similar allowances are made for lunar magnetic permeability.) Subsatellite estimates are shown as (d) with the solid brackets indicating statistical error and the dashed brackets including possible estimated biases. These results establish an upper (lower) limit for core radius of 580 (350) km for \( G_\mu = 0.008 \) (0.000). A possible 350–170 km radius core has been estimated from seismic data (Nakamura et al., 1974), but a core as large as 500 km could be compatible with the seismic data (Nakamura, personal communication). The moment-of-inertia constraint (Solomon, 1974) derivable from the recent measurement of the lunar moment of inertia (Gapcynski et al., 1975) allows a maximum FeS core radius of 725 km and a maximum Fe core radius of 480 km.

The lower limit of core radius will not be substantially reduced from the previous simple estimate by the effects of finite conductivity in the mantle. To study the effects of mantle conductivity we adopt a three layer lunar model with constant conductivity within each layer. The core conductivity is taken as 10\(^3\) mho/m for an FeS or Fe composition (Johnston and Strens, 1973). Other materials of sufficient conductivity to exclude magnetic field would be a complete melt of mantle material with conductivity on the order of 10 mho/m (Khitarov et al. 1970), but convective cooling makes this an unlikely model for the lunar interior. Another high conductivity possibility is a sodium-containing mineral which might act as a semi-conductor at high temperatures. Piwinskii and Duba (1974) have found a conductivity for albite of at least 4 mho/m at 1384°K (within 10°K of melting). Such conductivities would be almost sufficient to account for the subsatellite measurements (e.g., see the 10 mho/m curve in Fig. 4). Above the core we assume a partial melt zone with conductivity of 0.1 mho/m (a probable overestimate to exaggerate the contribution of mantle conductivity); melts of from 3 to 5\% might cause such conductivities (Waif. 1974). The outer shell consists of non-molten or small-fraction melt material with an assumed conductivity of zero. A fit of individual measurements of \( G_\mu \) to the response of the
different models was computed and the effects of mantle conductivity were found to be negligible at the long time periods. However, if we assume that the 0–12 hr estimates are somehow affected by mantle conductivity anyway, and use only the response at periods greater than 12 hr and apply the estimated maximum plasma correction, we obtain a minimum core radius of 320 km. This suggested core size (0.6–2.5% by volume) is not in conflict with a moon formed predominantly of refractory, high temperature, materials. In fact, seismic data tentatively suggest that the moon might have a small differentiated core. The subsatellite measurements are in disagreement with other magnetometer results (PDD74, DPD75, this work) that allow, but do not require, a highly conducting core. The only really safe procedure at present is to take an upper limit from the subsatellite data and conclude that the maximum possible radius of a conducting lunar core is 580 km; this conclusion is consistent with all sets of data.

OTHER ARGUMENTS CONCERNING A LUNAR CORE

Hodges and Hoffman (1975) have recently reported observations of atmospheric 40Ar. The only possible source is decay of radioactive potassium in a partially molten asthenosphere. It is argued that a lunar core is incompatible with retention of K in the asthenosphere (Hodges and Hoffman, 1975; Taylor and Bence, 1975), and that a lunar core is therefore an unlikely possibility. If the whole moon underwent differentiation, then the above arguments require K in a lunar core. However, alternative scenarios can provide for core formation while retaining K in an undifferentiated asthenosphere.

Recent cool thermal models (Hubbard and Minear, 1975; Solomon and Chaiken, 1976) suggest that the upper 200–350 km of the moon (30–50% of the volume) underwent differentiation. Consequent volcanism led to eruption of mare basalts whose source regions were saturated in sulfur (Brett, 1975); this strongly suggests coexistence with an immiscible FeS melt of unknown quantity. The question that then arises is whether or not this melt somehow got to the center of the moon.

One model for core formation operates via a “zone refining” mechanism. In the Hubbard-Minear model the zone of partial melt did not progress continuously into the lunar interior; rather, the outer portions of the moon solidified followed by deep partial melting. However, slight perturbations in the thermal model would allow a small fraction partial melt zone to descend continuously into the lunar interior; in this scenario the core forms late in lunar history by trickling down of the FeS melt. Cool thermal models (Solomon and Chaiken, 1976) require that convection brings down the melt. As the deep lunar interior is composed of undifferentiated material (although convection will cause some mixing with the differentiated material above), K is retained in the asthenosphere.

An alternative method of core formation (the one that is favored for the earth) was suggested by Elsasser (1963). In this model a dense layer of metallic melt overlying less dense mantle material develops a gravitational instability. The FeS melt drains into depressions and eventually iron drops (perhaps 50 km in radius,
formation of large drops is favored) form. The iron drops then sink through the asthenosphere to the center of the moon. However, the drops cannot sink through rigid material; Stoke's law for the rate of descent of a drop in a viscous fluid suggests that mantle viscosities on the order of $10^{24}$ poise or less are required to allow core formation. Such low viscosities require that the mantle material be within 70% of solidus temperature (Stocker and Ashby, 1973).

In the thermal model of Solomon and Chaiken (1976) the deepest portions of the lunar interior reach a temperature within 70% of solidus about 2 b.y. after formation of the moon. As a result, core formation cannot occur early in lunar history. However, the core may have formed even later than 2 b.y. into lunar history; smaller drop sizes would delay core formation and little is known about the actual process of drop formation. Murthy et al. (1971) suggest that the FeS melt is trapped in subsurface pockets and does not descend into the interior. The difference between our model and theirs is due to the higher mantle viscosity ($10^{26}$ poise) that they use.

Returning to the issue of K in the lunar interior, in the sinking drop model of core formation potassium is retained in an undifferentiated asthenosphere as the lunar interior need never reach solidus. In our zone refining model K is also retained in the asthenosphere if the melt is brought down by convection; however, K might be differentiated upwards in warm models with a continuously descending partial melt zone. Finally, the atmospheric argon observations are also compatible with a core and whole moon differentiation if K is retained in the core (Hodges and Hoffman, 1975). In summary, the atmospheric argon observations do not imply any serious constraints upon core formation, but do imply a partially molten lunar interior containing K.

Quantitative estimates of possible core size in these models depend primarily upon the amount of S in the moon. If we assume that $f_D$ is the fraction of the moon that has undergone differentiation, and letting $q_s$ be the fraction by weight of S in the moon, all S enters the sulfide phase, letting $\rho$ denote density, and taking a eutectic ratio of Fe$_{0.75}$S$_{0.25}$ by weight, then the volume fraction of the FeS melt is

$$V_{\text{melt}}/V_{\text{moon}} = (\rho_{\text{moon}}/\rho_{\text{FeS}}) f_D 4q_s.$$  (9)

Wänke et al. (1975) have suggested $q_s = 0.002$; Ganapathy and Anders (1974) and Vizgirda and Anders (1976) suggest $q_s = 0.003$. If S was present in the moon in the same proportion as lithophile volatiles ($8 \times 10^{-2}$ of cosmic abundances, Vizgirda and Anders, 1976) then $q_s$ could be as large as 0.006 to 0.008; depletion of lunar siderophiles and sulfur would then be attributed to formation of a sulfur phase rather than preaccretion loss. Taking $f_D = 0.3$ and $q_s = 0.003$, the fractional volume of a lunar core would be 0.22% with a fractional radius $\lambda_s = 0.14$ (230 km). Assuming that the original weight of metallic plus troilite associated Fe in the lunar mantle was 1.5% (Wänke et al., 1975), formation of the sulfide phase would leave only 0.6% in the upper mantle. On this basis, $G_s = 0.0015$, $G_c = 0.0012$, and $G_{\text{obs}} = +0.0003$. This model value is just at the lower limits of our surface measurements. If the best fit surface values are correct, more free iron and/or less sulfur in the moon or less differentiation will be required. If $q_s = 0.008$ and
$f_D = 0.500$, then the fractional volume of the core is 0.010 with a fractional radius of 0.21 (370 km). A core of this size would be suggested by subsatellite observations and would not occupy such a substantial volume of the asthenosphere as to raise problems for potassium content. Resolution of the discrepancy in the magnetometer observations would allow distinguishing between these two alternative models of the lunar interior.

We also mention that in any of the models we have discussed, dynamo action in a lunar core cannot provide a magnetizing field to account for lunar remanent magnetism as the core forms too late in lunar history. Rather, dynamo action in FeS melts in the upper crust (Murthy and Banerjee, 1973) could provide the source of magnetization with or without consequent formation of a lunar core. Russell and Goldstein (1976) discuss scaling laws for planetary dynamos; a lunar core as large as 400 km would be expected to result in a permanent dipole field large enough to have been detected by the subsatellite magnetometer.

**Conclusions and Future Prospects**

The subsatellite observations reported here require the existence of a conducting lunar core. However, plasma diamagnetic effects in the tail lobes are a source of uncertainty in magnetospheric observations. Disagreement between our observations and those of other workers (PDD74) require that caution be exercised when interpreting magnetospheric observations of lunar induced magnetic fields. Until the source of the discrepancy is determined the only safe conclusion allowed is a maximum lunar core radius of 580 km. Further studies are required using overdetermined measurements (more magnetometers than necessary) and a variety of plasma data. However, present data are both too few and too noisy. The Lunar Polar Orbiter mission provides an opportunity to obtain data from two orbital magnetometers simultaneously, thus avoiding problems of plasma interactions with local remanent fields that might influence surface observations. A plasma detector (such as proposed electron mirroring experiments) would also be necessary to eliminate periods of plasma diamagnetism. Such instrumentation would resolve the current discrepancy in estimates of induced lunar magnetic moment and conclusively test present lunar core results.

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