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MAGNETISM AND THE INTERIOR OF THE MOON

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ABSTRACT

During the time period 1961-1972 eleven magnetometers were sent to the moon. The primary purpose of this paper is to review the results of lunar magnetometer data analysis, with emphasis on the lunar interior. Magnetic fields have been measured on the lunar surface at the Apollo 12, 14, 15, and 16 landing sites. The remanent field values at these sites are respectively 38 γ, 103 γ (maximum), 3 γ, and 327 γ (maximum) Simultaneous magnetic field and solar plasma pressure measurements show that the Apollo 12 and 16 remanent fields are compressed during times of high plasma dynamic pressure. Apollo 15 and 16 subsatellite magnetometers have mapped in detail the fields above portions of the lunar surface and have placed an upper limit of 4.4×10^{13} gauss-cm^3 on the global permanent dipole moment. Satellite and surface measurements show strong evidence that the lunar crust is magnetized over much of the lunar globe. Magnetic fields are stronger in highland regions than in mare regions, and stronger on the lunar far side than on the near side. The largest magnetic anomaly measured to date is between the craters Van de Graaff and Aitken on the lunar far side. The origin of the lunar remanent field is not yet satisfactorily understood; several source models are presented. Simultaneous data from the Apollo 12 lunar surface magnetometer and the Explorer 35 Ames magnetometer are used to construct a whole-moon hysteresis curve, from which the global lunar permeability is

determined to be $\mu = 1.012 \pm 0.006$. The corresponding global induced dipole moment is $\sim 2 \times 10^{18}$ gauss-cm^3 for typical inducing fields of 10^{-4} gauss in the lunar environment. From the permeability measurement, lunar free iron abundance is determined to be 2.5 ± 2.0 wt. %. Total iron abundance (sum of iron in the ferromagnetic and paramagnetic states) is calculated for two assumed compositional models of the lunar interior. For a free iron/orthopyroxene lunar composition the total iron content is 12.8 ± 1.0 wt. %; for a free iron/olivine composition, total iron content is 5.5 ± 1.2 wt. %. Other lunar models with a small iron core and with a shallow iron-rich layer are also discussed in light of the measured global permeability. Global eddy current fields, induced by changes in the magnetic field external to the moon, have been analyzed to calculate lunar electrical conductivity profiles using several different analytical techniques. From nightside transient data with the moon in the solar wind, it has been found that deeper than 170 km into the moon, the conductivity rises from 3×10^{-4} mhos/meter to 10^{-2} mhos/meter at 1000 km depth. Harmonic analyses of dayside data differ from nightside results primarily at the greater lunar depths, where harmonic dayside profiles show lower conductivities than do the nightside results. Recent conductivity analysis of transients in the geomagnetic tail avoids many of the analytical problems posed by asymmetric confinement of induced fields in the solar wind. A conductivity profile calculated from geomagnetic tail transient analysis, increases with depth from 10^{-9} mhos/meter at the lunar surface to 10^{-4} mhos/meter at 340 km depth, then to 2×10^{-2} mhos/meter at 870 km depth. This conductivity profile is converted to temperature and compared with temperature results of other investigators.

CONTENTS

I. INTRODUCTION

Magnetometers placed on the lunar surface and in orbit about the moon have returned a wealth of information about the moon which was not anticipated prior to the Apollo manned lunar missions. Earlier measurements, by USSR and U.S. magnetometers on unmanned spacecraft, indicated that the moon might be electromagnetically inert; during that time investigators often concentrated on the interactions of the moon with the solar wind plasma (Colburn et al., 1967; Ness et al., 1967; Michel, 1968; Spreiter et al., 1970) rather than on magnetic studies of the lunar interior.

The measurement of magnetic fields in the vicinity of the moon began in January 1959, when the USSR spacecraft Luna 1 carried a magnetometer to within several hundred kilometers of the moon. In September 1959, Luna 2, also equipped with a magnetometer, impacted the lunar surface. The instrument aboard Luna 2 set an upper limit of 100 gammas (1 gamma = 10^{-5} gauss) for a possible lunar field at an altitude of about 50 km above the moon's surface (Dolginov, 1961). In April 1966, Luna 10, carrying a magnetometer 10 times more sensitive than that aboard Luna 2, was successfully placed in a lunar orbit that came to within 350 km of the moon (see Figure 1). The Luna 10 magnetometer recorded a time-varying magnetic field in the vicinity of the moon which was at that time interpreted as indicating the existence of a weak lunar magneto- Fig. 1

sphere (Dolginov et al., 1966).

A year later (July 1967) the United States placed the Explorer 35 satellite, with two magnetometers aboard, in orbit around the moon. In its orbit the satellite passes to within 830 km of the moon's surface. Explorer 35 successfully measured magnetic properties of the solar-wind cavity downstream from the moon, but it did not detect the lunar magnetosphere indicated by Luna 10 measurements nor the lunar bow shock and induced-field configuration previously suggested by Gold (1966). In an analysis of the Explorer 35 results Sonett et al. (1967) concluded that if a permanent lunar field exists at all, its magnitude would be less than two gammas at an altitude of 830 km and therefore $\lesssim$ 4 gammas at the surface for a global permanent dipole field. The upper limit on the global permanent dipole moment was set at 10^{20} gauss-cm^3, i.e., less than 10^{-5} that of the earth. In studies of the solar wind interaction with the lunar body (Colburn et al., 1967, Ness et al., 1967), investigators found the solar wind field magnitude to be $\sim$1.5 gammas greater in the diamagnetic cavity on the moon's antisolar side than in the solar wind.

Surveyor spacecraft, used in the first U.S. unmanned lunar landings, carried no magnetometers. Permanent magnets were carried aboard the Surveyor 5 and 6 spacecraft, however, which demonstrated that soils at those two landing sites contain less than

1% (by volume) ferromagnetic iron (de Wys, 1968).

During the early manned Apollo missions it was determined that the moon is much more interesting magnetically than had been expected. Natural remanent magnetization in lunar samples was found to be surprisingly high at all U.S. Apollo sites and at the USSR Luna 16 site (see, for example, Strangway et. al., 1970; Runcorn et al., 1970; Pearce et al., 1971; Nagata et al., 1971; Nagata et al., 1972b; Collinson et al., 1973; Pearce et al., 1973). Such high natural remanent magnetism implies that at some time in the past there existed an ambient surface magnetic field considerably higher than that which now exists on the moon (Runcorn et al., 1970; Gose et al., 1972).

The first lunar surface magnetometer (LSM), deployed at the Apollo 12 site in November 1969, made the first direct measurement of an intrinsic lunar magnetic field (Dyal et al., 1970a). The 38-gamma field measurement showed that not only are individual rocks magnetized, but also that magnetization in the lunar crust can be ordered over much larger regions, of 2 km to 200 km scale sizes (Dyal et al., 1970a; Barnes et al., 1971). The permanent and induced fields measured by the Apollo 12 magnetometer provided the impetus to develop portable surface magnetometers and subsatellite magnetometers for later Apollo missions. Permanent magnetic fields were subsequently measured at four other landing sites: Apollo 14 (103 γ maximum), Apollo 15 (3 γ), Apollo 16 (327 γ maximum), and more recently at several positions along the USSR Lunakhod II traverse. The surface fields were attributed to local magnetized sources ("magcons"); their discovery prompted a reexamination of Explorer 35 magnetometer data by Mihalov et al., (1971), who found indirect evidence that several magnetized regions exist in the lunar crust. Direct field measurements from Apollo 15 and 16 subsatellite magnetometers, activated in August 1971 and April 1972, respectively, yielded maps of some of the larger magnetized regions in the lunar crust (Coleman et al., 1972c; Sharp et al., 1973) which confirmed the existence of magnetized regions over much of the lunar surface. Subsatellite magnetometer measurements have also placed an upper limit of 4.4×10^{13} gauss-cm^3 on the global permanent magnetic dipole moment (Russell et al., 1973).

Investigations of simultaneous surface magnetometer data and solar wind spectrometer data show that the surface remanent magnetic fields interact with the solar wind when on the dayside of the moon (Dyal et al., 1972a; Clay et al., 1972; Dyal and Parkin, 1971b). The interaction is interpreted as a compression of the surface remanent fields by the solar wind; the magnetic pressure at the surface increases in proportion to the dynamic bulk pressure of the solar wind plasma.

In addition to measuring permanent lunar fields, the network of lunar surface and orbiting magnetometers measured fields induced in the lunar interior by extralunar magnetic fields, allowing investigation of deep interior properties of the moon. Behannon (1968) placed an upper limit of 1.8 on the bulk relative magnetic permeability by studying Explorer 35 magnetometer measurements in the geomagnetic tail. Subsequently, simultaneous measurements of Explorer 35 and Apollo 12 magnetometers have been used to yield the more accurate value of 1.012 ± 0.006 (Parkin et al., 1974a,b). This permeability value has been used to calculate free iron and total iron abundance of the moon.

The electrical conductivity of the lunar interior has been investigated by analyzing the induction of global lunar fields by time-varying extralunar (solar or terrestrial) magnetic fields. Since temperature and conductivity of geological materials are related, calculated conductivity profiles have been used to infer temperature of the lunar interior. Early estimates of bulk lunar electrical conductivity were made from lunar-orbiting Explorer 35 data by Colburn et al., (1967) and Ness (1968). For homogeneous-conductivity models of the moon, Colburn et al. placed an upper limit of 10^{-6} mhos/m for whole-moon conductivity, whereas Ness' upper limit was 10^{-5} mhos/m. These investigators also stated that their measurements were consistent with higher conductivity for a lunar core surrounded by an insulating crust.

Theoretical studies of the electrodynamic response of the moon to time-dependent external fields have been undertaken by many authors. Two types of whole-moon magnetic induction fields have been treated; a poloidal field due to eddy currents driven by time-varying external magnetic fields, and a toroidal field due to unipolar currents driven through the moon by the motional solar-wind $\underline{V} \times \underline{B}$ electric field.

The toroidal induction mode, first suggested to be an important process in the moon by Sonett and Colburn (1967), was later developed in detail theoretically for a lunar model totally confined by the highly conducting solar wind (Schwartz and Schubert, 1969; Schubert and Schwartz, 1969; Sill and Blank, 1970). However, analysis of simultaneous Apollo 12 and Explorer 35 magnetometer data later indicated that for the moon, toroidal induction is negligible in comparison to poloidal induction; upper limits on the toroidal field mode were used to calculate an upper limit of 10^{-9} mhos/m for electrical conductivity of the outer 5 km of the lunar crust (Dyal and Parkin, 1971a). In subsequent analysis of lunar electromagnetic induction, toroidal induction has been assumed to be negligible relative to poloidal induction.

The eddy-current response of a homogeneous sphere in a vacuum to time-varying magnetic fields has been described by Smythe (1950) and Wait (1951). Early theoretical application of vacuum poloidal induction to studies of the lunar interior were presented by Gold (1966) and Tozer and Wilson (1967). Poloidal response theory for a lunar sphere totally confined by a highly conducting plasma was developed by Blank and Sill (1969), Schubert and Schwartz (1969), and Sill and Blank (1970).

Since deployment of the Apollo 12 magnetometer in November 1969, electrical conductivity analysis has been developed with two basic approaches: a time-dependent, transient- response technique (Dyal and Parkin, 1971c; Sill, 1972; Dyal et al., 1973) and a frequency-dependent, Fourier-harmonic technique (Sonett et. al., 1971, 1972; Kuckes, 1971; Sill, 1972; Hobbs, 1973). Past analyses have all used magnetometer data recorded at times when global eddy current fields were asymmetrically confined by the solar wind plasma (Reisz et al., 1972; Dyal and Parkin, 1973; Dyal et al., 1973; Schubert et al., 1973c; Smith et al., 1973). The asymmetric confinement of lunar fields is particularly complex to model theoretically (Schubert et al., 1973b); indeed, the general time-dependent asymmetric induction problem has not been solved at the time of this writing. To avoid these complications, recent conductivity analysis has used field data recorded in the geomagnetic tail, which is relatively free of plasma and asymmetric confinement effects. Preliminary results of this analysis will be presented in this paper.

The purpose of this paper is to review the application of lunar magnetic field measurements to the study of properties of the lunar crust and deep interior. Following a brief descriptive section on lunar magnetometers and the lunar magnetic environment, measurements of lunar remanent fields and their interaction with the solar plasma will be discussed. Then the magnetization induction mode will be considered with reference to lunar magnetic permeability and iron abundance calculations. Finally, electrical conductivity and temperature calculations from analyses of poloidal induction, for data taken in both the solar wind and in the geomagnetic tail, will be reviewed.

II LUNAR MAGNETOMETERS AND THE LUNAR MAGNETIC ENVIRONMENT

A. The Lunar Magnetometers

A network of three lunar surface magnetometers has been placed on the moon by astronauts on the Apollo 12, 15, and 16 missions. The vector magnetic field is measured three times per second and transmitted to earth from each of the three Apollo sites shown in Figure 1. Magnetometers in lunar orbit on board the Explorer 35 satellite (Sonett et. al., 1967) and on Apollo 15 and 16 subsatellites (Coleman et al., 1972a,b) measure the field external to the moon and transmit this information to earth. Also, portable magnetometers have been used by the astronauts at the Apollo 14 and 16 sites, to measure remanent fields at different locations along a surface traverse.

The stationary lunar surface magnetometer (LSM) deployed at the Apollo 12 site on the moon is shown in Figure 2. Similar magnetometers have been placed at the Apollo 15 and 16 sites. Fig.
The three orthogonal vector components of the lunar surface magnetic field are measured by three fluxgate sensors (Gordon et al., 1965) located at the ends of three 100-cm-long orthogonal booms. The sensors are separated from each other by 150 cm and are 75 cm above the ground. The analog output of each sensor is internally processed by a low-pass digital filter and a telemetry encoder,

and the output is transmitted to earth via the central station S-band transmitter. The magnetometer has two data samplers, the analog-to-digital converter (26.5 samples/second) and the central station telemetry encoder (3.3 samples/second). The prealias filter following the sensor electronics has attenuations of 3 dB at 1.7 Hz and 58 dB at the Nyquist frequency (13.2 Hz), with an attenuation rate of 22 dB/octave. The four-pole Bessel digital filter has an attenuation of 3 dB at 0.3 Hz and 48 dB at the telemetry sampling Nyquist frequency (1.6 Hz). Instrument resolution is 0.2γ. The instrument is also used as a gradiometer by sending commands to operate three motors in the instrument which rotate the sensors such that all simultaneously align parallel first to one of the boom axes, then to each of the other two boom axes in turn. This sensor alignment permits the vector gradient to be calculated in the plane of the sensors and also permits an independent measurement of the magnetic field vector at each sensor position. A detailed description of the stationary magnetometer is reported by Dyal et al., (1970b).

The lunar portable magnetometer (LPM) was developed for astronaut deployment on the Apollo 14 and 16 missions (the Apollo 16 LPM is shown in Figure 3b). The instrument was designed to be Fig. 3
a totally self-contained, portable experiment package. Three

orthogonally oriented fluxgate sensors are mounted on the top of a tripod, positioned 75 cm above the lunar surface. These sensors are connected by a 15-meter-long cable to an electronics box which contains a battery, electronics, and three field component readouts (meters on Apollo 14 LPM; digital displays on Apollo 16 LPM). The electronics box is mounted on the mobile equipment transporter for Apollo 14 and on the lunar roving vehicle for Apollo 16. Portable magnetometer measurements can be made only by manual activation by an astronaut. Instrument resolution is 1γ or 2γ for the Apollo 14 LPM and 0.2γ for the Apollo 16 LPM. Detailed instrument characteristics are reported by Dyal et al. (1972c).

Magnetic fields of the lunar environment are measured by the Explorer 35 satellite magnetometers. The satellite, launched in July 1967, has an orbital period of 11.5 hours, aposelene of 9390 km, and periselene of 2570 km. Two magnetometers are carried aboard Explorer 35, one provided by NASA-Ames Research Center and the other provided by NASA-Goddard Space Flight Center. Since most of the analysis of lunar internal properties has been carried out using the Ames magnetometer, its characteristics will be considered here. The Explorer 35 Ames magnetometer measures three magnetic field vector components every 6.14 sec at 0.4γ resolution; the instrument has an alias filter with 18 dB attenuation at the Nyquist frequency (0.08 Hz) of the spacecraft data sampling system. A more detailed description of the instrument is reported by Sonett et al. (1967b).

The Apollo 15 (Figure 3a) and 16 subsatellite magnetometers, which orbited approximately 100 km above the lunar surface, also measured fields intrinsic to the moon. The subsatellite period of revolution is 2 hours. The subsatellite realtime sampling rates are 1 per 2 seconds for the field component along the spacecraft spin axis and 1 per second for the field component in the spin plane. Storage rates (for fields measured when the subsatellite is behind the moon relative to earth) are 1 field vector per 12 seconds (high rate) or 1 vector per 24 seconds (low rate). Apollo 15 resolution is 0.4γ or 1.6γ, depending on range; Apollo 16 resolution is 0.2γ or 0.8γ. The subsatellite magnetometer characteristics are described in detail by Coleman et al. (1972 a,b). At the time of this writing, the Apollo 16 lunar surface magnetometer (LSM) is the only lunar surface or orbital magnetometer which is still returning useful scientific data.

B. The Lunar Magnetic Environment

In different regions of a lunar orbit, the magnetic environment of the moon can have distinctly different characteristics (see Figure 4). Average magnetic field conditions vary from relatively steady fields of magnitude $\sim 9\gamma$ in the geomagnetic tail to mildly turbulent fields averaging $\sim 5\gamma$ in the free-streaming solar plasma region to turbulent fields averaging $\sim 8\gamma$ in the magnetosheath. Average solar wind velocity is ~ 400 km/sec in a direction approximately along the sun-earth line. ⟨Fig

The interaction of the solar wind with the earth's permanent dipole field results in formation of the characteristic shape of the earth's magnetosphere; the solar wind in effect sweeps the earth's field back into a cylindrical region (the geomagnetic tail) on the earth's antisolar side. The earth's field magnitude is about $30{,}000\gamma$ at the equator; in the geomagnetic tail the field decreases with distance from the earth with a radial dependence expressible as $R^{-0.736}$ (Mihalov et al., 1968). At the distance where the moon's orbit intersects the tail, the field magnitude is ~ 10 gammas. The moon is in the geomagnetic tail for about four days of each 29.5-day lunation (period between successive full moons). Substructure of the tail consists of two "lobes": the upper or northward lobe has its magnetic field pointing roughly toward the earth, whereas the lower lobe field points away from the earth. The moon can pass through either or both lobes, depending upon the characteristics of the particular orbit, the geomagnetic dipole axis orientation, and perturbations of the geomagnetic field by solar wind pressures.

The total magnetic field at the lunar surface is the vector sum of the following fields: the external (solar or terrestrial) "driving" field, permanent and induced lunar fields, and fields associated with the moon-plasma interaction in the solar wind. The lunar orbiting Explorer 35 magnetometer measures the external driving field. Analysis of these external "input" fields and the corresponding surface "response" fields allows calculation of properties of the lunar interior such as magnetic permeability, iron abundance, electrical conductivity, and temperature. The solar plasma interaction with lunar surface remanent fields is studied by correlating field data with plasma measurements.

Since the external driving magnetic field in the lunar environment can vary considerably with the lunar orbital position (see Figure 4), it is possible to study different lunar properties in different regions of the lunar orbit. A particular property of the moon is analyzed by selecting the lunar environment most favorable for induction of a lunar field response which is functionally dependent upon that particular property. During times when the moon is immersed in the steady geomagnetic tail field, time-dependent induction fields and solar wind interaction fields are minimal. This condition allows investigation of lunar permanent fields and time-independent magnetization ("soft perm") fields; the latter are functions of lunar magnetic permeability. When the moon is immersed in the variable fields of the free streaming solar wind or the magnetosheath, poloidal induction is the dominant lunar response field. In the past, investigators have used data taken in these regions of the orbit to study electrical conductivity. More recent analyses have used magnetic field data measured in the geomagnetic tail, where plasma effects are minimized, in order to avoid complications due to asymmetric plasma confinement.

III. LUNAR REMANENT MAGNETIC FIELDS

The permanent magnetic fields of the moon have been investigated using surface magnetometer measurements at four Apollo sites and one USSR Luna site; orbital measurements from Explorer 35 and two Apollo subsatellite magnetometers, and natural remanent magnetization measurements of returned lunar samples. Lunar remanent field measurements by surface and orbiting magnetometers will be emphasized in this paper. Sample magnetization measurements have been reviewed elsewhere (Hinners, 1971 ; Nagata et al., 1972a; Strangway et al., 1973c; Fuller, 1974).

A. Surface Site Field Measurements

The permanent magnetic fields of the moon were first measured *in situ* by the Apollo 12 lunar surface magnetometer (LSM) which was deployed on the eastern edge of Oceanus Procellarum. The permanent field magnitude was measured to be 38 ± 3 gammas and the source of this field was determined to be local in extent (Dyal et al., 1970a; Barnes et al., 1971 ; see Figure 5). A remanent field this large was generally unexpected even in light of the NRM discovered in the Apollo 11 samples, and the explanation of its origin yet remains a central problem in lunar magnetism. Subsequent to this measurement of an intrinsic lunar magnetic field, surface magnetometers have measured fields at the Apollo 14, 15 and 16 sites. Fields of 103 ± 5 and 43 ± 6 gammas, at two sites located about a

Fig. 5

kilometer apart, were measured by the Apollo 14 Lunar Portable magnetometer (LPM) at Fra Mauro. (see Figure 6). A steady field Fig. 6
of 3.4±2.9 gammas was measured near Hadley Rille by the Apollo 15 LSM (see Figure 7). At the Apollo 16 landing site both a portable and stationary magnetometer were deployed; magnetic fields Fig. 7
ranging between 112 and 327 gammas were measured at five different locations over a total distance of 7.1 kilometers at the Descartes landing site. These are the largest lunar fields yet measured. A schematic representation of the measured field vectors is shown in Figure 8 . All the vectors have components pointing downward Fig. 8
except the one at Site 5 near Stone Mountain, which points upward. This suggests among other possibilities, that the material underlying Stone Mountain has undergone different geological processes than that underlying the Cayley Plains and North Ray Crater. In fact, Strangway et al. (1973b) proposed the possibility that the light colored, relatively smooth Cayley formation is magnetized roughly vertically; the difference in the vertical component at site 5 was explained as an edge effect at the Cayley Plains-Stone Mountain boundary. A summary of all remanent lunar fields measured by the magnetometers deployed on the surface is given in Table 1. Table 1.

Interaction of the solar wind with the remanent magnetic field has been measured at the Apollo 12 and 16 landing sites. The solar plasma is directly measured at the Apollo 12 and 15 sites (Clay et al., 1972) and simultaneous magnetic field and plasma data show a compression of the steady field as a function of the solar wind pressure at the Apollo 12 and 16 sites (Dyal et al., 1973). The nature of the correlation between magnetic field and plasma bulk flow pressures is shown in Figure 9, which shows data (combined from several lunations) at the Apollo 12 and 16 LSM sites. The plasma bulk flow pressure and the magnetic pressure are related throughout the measurement range, and the magnitudes of magnetic pressure changes are in proportion to the unperturbed steady field magnitudes at each site. Fig. 9

Information on the scale sizes of the permanently magnetized regions near Apollo landing sites is given by gradient measurements of the lunar surface magnetometers, the spacing of vector measurements over the lunar surface, the known interaction properties of these remanent fields with the solar wind plasma, and limits imposed by satellite measurements. The field gradient in a plane parallel to the lunar surface is less than the instrument resolution of 0.13 gamma/meter at the Apollo 12 and 15 sites. At Apollo 14 a field difference of 60γ was measured at two sites located 1.1 km apart. Gradient measurements and the absence of changes in the permanent field at the sites after lunar module ascent demonstrated that the field sources are not magnetized artifacts.

The scale size of the Apollo 12 remanent field has been calculated from local gradient and Explorer 35 measurements to be from 2 km to 200 km (Dyal et al., 1972a). For the Apollo 16 field, portable magnetometer measurements over the lunar roving vehicle traverse showed that the scale size for the field was greater than 5 km; the Apollo 16 subsatellite magnetometer showed no anomalous field attributable to the Descartes area at orbital altitude, implying a surface field scale size upper limit of 100 km. Therefore the Apollo 16 remanent field scale size is between 5 and 100 km.

B. Orbital Field Measurements

Remanent magnetic fields over extended regions of the surface have been studied using lunar orbiting magnetometers. Two techniques have been employed in these studies: measurement of the effects from the solar wind interaction with surface fields near the limb and direct measurement and mapping of the fields from orbit.

Interaction of the solar wind with certain regions near the limb of the moon were used to infer surface field strengths and scale sizes by Barnes et al. (1971) and Mihalov et al. (1971). Their techniques were based on the concept that the highly conducting solar wind plasma may interact with certain high field regions on the moon when the regions are near the limb, to produce a weak local shock or "limb compression". Such shocks could be detected downstream from the moon. Magnetic events in the Explorer 35 magnetometer data were interpreted by Mihalov et al. (1971) as evidence for these limb shocks and were also used to infer locations and relative field strengths of surface field concentrations. They concluded that there may be magnetic concentrations covering much of the lunar surface, with stronger concentrations on the lunar far side than on the near side and generally in highland rather than mare regions. Limb compressions have also been observed by the close orbiting Apollo 15 and 16 subsatellite magnetometers. Using data from these magnetometers Russell et al. (1973) have concluded that the source regions for the limb compressions tend to be concentrated in the lunar highlands and especially in the Tsiolkovsky area on the lunar far side. Limb shocks imply but do not confirm the existence of magnetized regions on the moon, since the exact mechanism for limb shock formation is not known and there are other possible causes for their formation such as crustal conductivity anomalies (Mihalov et al., 197/; Schubert et al., 1974) or solar wind interaction with irregularities in the lunar electron sheath near the limb (Criswell, 1973).

Direct measurements of the lunar surface fields have been made over about 5% of the lunar surface by the Apollo 15 and 16 subsatellite magnetometers (Coleman et. al., 1972a,b). At the subsatellite altitudes of about 100 km the surface field magnitudes are generally less than 0.5 gamma (although sometimes they are as much as 2.5 gammas); by averaging data from several orbits which pass over the same surface area, measurements of the fields can be made along the entire orbit. Then successive orbits can be combined to produce magnetic maps of the regions over which the satellite passes. Using an empirical relation between the height and strength of the field, the orbital values are then referenced to 100 km altitude and contours of equal intensity can be drawn for this altitude. The maps of lunar remanent magnetic field shown in Figures 10 through 13 were constructed using the first four lunations of Apollo 15 subsatellite data (Sharp et al., 1973). (The contour labels are in hundreths of gammas and a positive value indicates an outward directed field) The most obvious characteristic of these maps is the presence of significant lunar fields over most of the mapped regions. Apparently the remanent fields measured at the Apollo surface sites are not anomalous but rather, characteristic of remanent fields over much of the lunar surface. The contour spacing shows another important characteristic of the field: gradients are often relatively small over large distances, implying that some of the field sources are large-scale, i.e., that subsurface materials may be magnetized in a systematic manner over distances of tens to a few hundreds of kilometers. While these magnetized regions may be shallow, it is possible that lunar material is magnetized to a depth of tens or a few hundreds of kilometers, perhaps even down to the depth where the temperature rises to the iron Curie point. Contour lines of the subsatellite maps do not extend over

Fig. 10
11
12
13

the Apollo sites where fields have been measured at the surface. However, at both the Apollo 15 and 16 sites, near the map's perimeter, the field component normal to the surface as measured on the surface and at 100 km seem to be completely unrelated (compare Table 1 and Figure 12).

Comparison of the features in the orbital maps with surface geologic features has been of great interest, and some correlation has been found (Sharp et al., 1973). There is a marked nearside-farside asymmetry. The farside field is generally stronger and more structured than the nearside field. This observation may be due to concentration of source regions in lunar highlands and the asymmetric distribution of the highlands between the far and near side.

The most promonent magnetic feature measured to date by the subsatellite is located between the craters Aitken and Van de Graaff. This region has been mapped at altitudes of 67 km and 130 km. Most of the main features are qualitatively similar at the altitudes 67 km and 130 km, although more detailed structure in the contours is apparent at the lower altitude. Again there is a wide range of field source scale sizes which are likely in this area. Sharp et al. (1973) have estimated the size and magnetization of the principal source region. If the source is a thin circular disk 160 km across and 10 km thick it would require a magnetization of nearly 7×10^{-5} emu/cm^3 to coincide with measurements. A magnetization typical of lunar samples is about 10^{-5} emu/cm^3. If the source region is a sphere instead of a disk, its center must be 75 km deep to yield the observed radial gradient. With a 10 km radius the magnetization would be 10^{-3} emu/cm^3, producing a 16 gamma surface field. A sphere 75 km in radius would require a magnetization intensity of 1.9×10^{-5} emu/cm^3. Other regions such as the Korolev, Hertsprung and Milne basins were also analyzed in some detail but the results were inconclusive due to unfavorable observational conditions and the smaller sizes of these features.

Some preliminary results from the Apollo 16 subsatellite magnetometer have been published (Sharp et al., 1973); as in the earlier experiment, the data indicate large-scale magnetic features in the lunar crust. The Apollo 16 subsatellite impacted the lunar surface shortly after the two passes of the moon through the geomagnetic tail, making available measurements from very low altitudes near periselene (10 to 0 km) during the last few orbits. Not only are the field magnitudes greatly enhanced (up to 56 gammas), but they change character over much smaller distances than those measured at 100 km. A wide range in field source sizes in the lunar crust is again indicated.

C. Origin of the Lunar Remanent Magnetization

From the beginning of the Apollo missions the origin of the lunar remanent fields has been of great interest. There has been no shortage of mechanism proposed to explain the origin of the lunar fields and remanence. Some of these possible mechanisms are discussed briefly here and displayed schematically in Figures 14 and 15. Fig. 14-15

1. Large Solar or Terrestrial Field. It has been proposed that the near-surface lunar material acquired a thermoremanence from cooling to the Curie point in the presence of a large solar or terrestrial field. One possibility is that the solar field was much stronger than it is now, and was also relatively steady during the rock crystallization (Nagata et al., 1972b). A terrestrial field increase greater than 100 times its present value would probably be necessary to magnetize lunar material at the present-day lunar orbit. Such a large terrestrial field is not indicated by paleomagnetic studies. For an ancient terrestrial field of present-day magnitude, the moon would have to have approached to within 2 to 3 Earth radii, close to the Roche limit (Runcorn et al., 1970; Helsley, 1970) to be subjected to the necessary field strength. All of the alternatives for these hypotheses seem to have shortcomings.

2. Iron Core Dynamo. For this mechanism a whole-moon field results from the self-generating dynamo action of a small iron core (Runcorn

et al., 1971; Strangway et al., 1971). The dynamo is assumed to have been active 3 to 4 billion years ago when surface rocks and breccias were formed. After the thermoremanent magnetization was established in the upper crust material, as it cooled through the Curie temperature, the dynamo turned off. Subsequently meteorite impacts on the magnetized surface randomized the field's sources by a gardening process and destroyed the whole-body magnetization in the crust. The core dynamo hypothesis also has its shortcomings. In the first place it is not clear that even the most efficient dynamo mechanism in a lunar core of limited size would be self sustaining at rotational speeds for which the moon could hold together (Levy, 1972). In addition, it is doubtful that a dynamo, if ever operating, could produce the surface fields to explain the thermal remanent magnetization of some lunar samples (Collinson et al., 1973; Levy, 1972).

Ancient Magnetized Core. Urey and Runcorn (1973) and Strangway et al. (1973a,c) have suggested that near surface material may have been magnetized by the field of a lunar core which had been previously magnetized by one of several possible means: (1) isothermally by a strong transient field, (2) viscous remanent magnetization by a weak field applied over a long period, (3) depositional remanent magnetization during early lunar formation in a weak field, or (4) thermoremanent magnetization of the core by cooling through the Curie point in a weak field. If the moon formed in a cold state, neither accretion nor radioactivity would necessarily have raised the temperature of the deep interior above the Curie point of iron, with accompanying loss of magnetization, until 4.1 to 3.2 billion years ago. In the outer shell, perhaps 200 to 400 km thick, partial melting could easily have been realized during later stages of accretion. During the crystallization of the crustal rocks in the magnetic field of the core, they obtained a thermoremanence. Subsequently, radioactive heating in the interior raised the core temperature above the Curie point, resulting in loss of the magnetization in the core.

4. Shallow Fe-FeS Dynamos. A model related to the lunar core dynamo is one hypothesizing small pockets of iron and iron sulfide (Fe-FeS) melt a few hundred kilometers below the surface (Murthy and Banerjee, 1973) which act as small localized dynamos. The proponents of this mechanism suggest that these "fescons" are about 100 km in diameter. A variation of this local dynamo idea is suggested by Smolychowski (1973) wherein a thin layer of molten basalt generates the field. The existence of such local source regions for magnetic field should be evident once the surface fields have been mapped over more of the lunar surface. However, the recently discovered asymmetry in the electromagnetic field fluctuations at the Apollo 15 landing site (Schubert et al., 1974) could be due to such

a highly conducting subsurface body.

5. Local Induced Unipolar Dynamo. The solar wind transports magnetic fields past the moon at velocities $\underline{V}$ of approximately 400 km/sec; the corresponding $\underline{V} \times \underline{B}$ electric field causes currents to flow along paths of high electrical conductivity (Schwartz et al., 1969; Nagata et al., 1972b) such as molten mare regions, with the highly conducting solar wind plasma completing the circuit back to the lunar interior. The fields associated with these currents magnetize the materials as they cool below the Curie temperature. Because this induction mechanism has the strongest influence while the hot region is sunlit, an average preferred direction is associated with $\underline{V}$. However, the $\underline{V} \times \underline{B}$ induction model requires solar wind magnetic fields or velocities much higher than the present-day sun provides.

6. Local Thermoelectrically Driven Dynamo. Dyal et al. (1973) have proposed a mechanism of thermoelectrically driven currents to account for remanent fields. Thermoelectric potential is a function of the thermal gradient and electrical properties of the geological material. For the mechanism a mare basin is modeled by a disk which has an axial temperature gradient. Thermal gradients in the cooling mare lava could produce a Thomson thermoelectromotive force which would drive currents axially through the mare disk. The solar wind plasma, highly conducting along magnetic field lines, could provide a return path to complete the electrical circuit from

the top surface of the lava to the lunar surface outside the mare and back into the mare through the lunar interior. The upper limit of the fields generated in terrestrial materials by this process is a few thousand gammas. Such fields near a mare disk would produce thermoremanent magnetization in the moon of magnitudes measured in lunar samples. This mechanism awaits experimental verification using materials characteristic of lunar mare composition.

7. Shock Magnetization. Anisotropic compression of rocks by meteorite impacts is suggested by Nagata et al. (1970) as a means of inducing a remanence in certain samples which they studied magnetically. This piezo-remanent magnetization can be significantly large even when the external field is very weak (e.g., the solar wind field) if the uniaxial compression is very large. This mechanism is appealing since it relies on a well-established lunar process and may explain some correlation between craters and magnetic anomalies (Sharp et al., 1973), but the details remain undeveloped.

8. Local Currents from Charged Particle Transport. Any process which results in plasma flow near the lunar surface may generate strong local currents and magnetic fields. Cap (1972), for example, has shown that ionized volcanic ash flows may produce fields up to 10^3 gammas. As another example, Nagata et al. (1970) proposed the idea that lightning may be generated as a result of exploding dust clouds from meteorite impacts. The large currents

associated with an electrical discharge could produce transient magnetic fields up to 10 or 20 gauss, resulting in isothermal remanent magnetization of local material.

IV. GLOBAL MAGNETIZATION INDUCTION: MAGNETIC PERMEABILITY AND IRON ABUNDANCE

Magnetic permeability and iron abundance of the moon are calculated by analysis of magnetization fields induced in the permeable material of the moon. When the moon is immersed in an external field it is magnetized; the induced magnetization is a function of the distribution of permeable material in the interior. Under the assumption that the permeable material in the moon is predominately free iron and iron-bearing minerals, the lunar iron abundance can be calculated from the lunar permeability for assumed compositional models of the interior. Since the amount of iron present in the lunar interior should be consistent with the measured global magnetic permeability, the permeability in effect places a constraint on the physical and chemical composition of the moon's interior.

In this section calculations of global permeability will be reviewed, beginning with earlier measurements which used a single moon-orbiting magnetometer and proceeding chronologically to more recent studies using simultaneous data from surface and orbital magnetometers. More recent values of global permeability have been used to determine lunar iron abundance. This analysis of iron abundance and its implications concerning lunar composition and structure will be discussed.

A. Global Magnetic Permeability

1. Early Results from Explorer 35 Measurements. Behannon (1968) placed an upper limit on lunar bulk permeability using data from the Explorer 35 Goddard magnetometer. In his analysis, Behannon used data measured when the moon was in steady-field regions of the geomagnetic tail, where it was assumed that the external field was uniform over the region of space near the moon and plasma effects were negligible. Under these conditions, for a moon of spherically symmetric permeability distribution, the induced magnetization moment would be dipolar, with the dipole axis oriented along the external field. Behannon compared Explorer 35 field measurements made when the moon was above the neutral sheet (with the induced dipole oriented along the field pointing sunward) with those made when the moon was below the neutral sheet (with the dipole oriented away from the sun); differences in the data pairs, on the average over many Explorer orbits, were considered to represent a measurement of the induced field. At the Explorer 35 periselene the induced field was found to be less than the experimental error. Using the experimental error as the upper limit of the induced field at the surface, Behannon calculated an upper limit of 1.8 for the bulk magnetic permeability of the moon.

2. Results from Simultaneous Apollo and Explorer 35 Measurements. Deployment of Apollo magnetometers on the lunar surface allowed simultaneous measurements of the external inducing field (by Explorer

35) and the total response field at the lunar surface (by an Apollo magnetometer). The total response field measured at the surface by an Apollo magnetometer is the sum of the external and induced fields:

$$\underline{B} = \mu\, \underline{H} = \underline{H} + 4\pi \underline{M} \qquad (1)$$

where $\underline{H}$ is the external magnetizing field and $\underline{M}$ is the magnetization field induced in the permeable lunar material (see Figure 16). The Fig. 16 relative magnetic permeability is $\mu = 1 + 4\pi k$, where k is magnetic susceptibility in emu/cm^3. Since the dipolar magnetization $\underline{M}$ is known to be below the Explorer 35 magnetometer resolution (Behannon, 1968), it is assumed in the dual magnetometer analysis that Explorer 35 measures $\underline{H}$ alone.

For the two-layer lunar permeability model illustrated in Figure 16 (which will be referred to later when iron abundance results are reviewed), the total field at the outer surface of the sphere is expressed

$$\underline{B} = H_x(1 + 2G)\,\hat{x} + H_y(1-G)\,\hat{y} + H_z(1-G)\,\hat{z} \qquad (2)$$

where

$$G = \frac{(2\eta+1)(\mu_1-1) - \lambda^3(\eta-1)(2\mu_1+1)}{(2\eta+1)(\mu_1+2) - 2\lambda^3(\eta-1)(\mu_1-1)} \qquad (3)$$

Here $\eta = \mu_1/\mu_2$; μ_1 and μ_2 are relative permeability of the shell and core, respectively. The permeability exterior to the sphere is $\mu_o = 1$, that of free space; $\lambda = R_c/R_m$; R_c and R_m are radius of the core and the moon, respectively. Equation (2) expresses the total surface field in a coordinate system which has its origin on the lunar surface at an Apollo magnetometer site: $\hat{x}$ is directed radially outward from the lunar surface, and $\hat{y}$ and $\hat{z}$ are tangential to the surface, directed eastward and northward, respectively.

A plot of any component of equation (2) will result in a B - H hysteresis curve. Equation (3) relates the slope of the hysteresis curve to the lunar permeability. The average whole-moon permeability μ is calculated from the hysteresis-curve slope by setting $\mu_1 = \mu_2 = \mu$ in equation (3):

$$G = \frac{\mu-1}{\mu+2} \tag{4}$$

The hysteresis-curve method of permeability analysis was first employed by Dyal and Parkin (1971) to calculate the whole-moon permeability result 1.03 ± 0.13. Since then the error limits have been lowered by processing a larger number of simultaneous data sets and using more rigid data selection criteria (e.g., Parkin et al., 1973).

In the most recent dual-magnetometer results (Parkin et al., 1974a,b), a hysteresis curve was constructed using 2703 data sets (see Figure 17). Since the external magnetizing field is so small ($\sim$ 10 gammas), the familiar "S" shape of the hysteresis curve degenerates to a straight line (Ellwood, 1934). The data were fit by a leastsquares technique which yields the slope best-estimate of 1.008 ± 0.004. Using this value with the radial (x) component of equation (2) and equation (4), the whole-moon permeability was calculated to be $\mu = 1.012 \pm 0.006$ (2 σ error limits). Both extrema are greater than 1.0, implying that the moon, as a whole, acts as a paramagnetic or weakly ferromagnetic sphere. This result has been used to calculate the iron abundance of the moon as discussed in the next section.

C. T. Russell (private communication) has recently made permeability calculations using data from a single magnetometer, the Apollo 15 subsatellite magnetometer orbiting at an altitude about 100 km above the moon. The results to date indicate that the relative permeability of the entire spherical volume enclosed by the satellite orbit is below 1.0, implying that the layer between the moon and the satellite orbit is diamagnetic. Whether such a layer exists is uncertain at this time; further investigation is required using both magnetic and plasma data.

B. <u>Lunar Iron Abundance</u>

Iron abundance calculations have been presented by various authors, in theoretical treatments based on geochemical and geophysical properties calculated for bodies of planetary size (Urey, 1962; Reynolds and Summers, 1969; Urey and MacDonald, 1971) or on measured compositions of meteorites (Wänke et al., 1973). Recently Parkin et al. (1973, 1974 a, b) used a global lunar permeability measurement, determined from magnetic field measurements, to calculate lunar iron abundance for the moon. In their calculations the moon was modeled by a homogeneous paramagnetic rock matrix (olivine and orthopyroxene models were used), in which free metallic iron is uniformly distributed. Pyroxenes and olivines have been reported to be major mineral components of the lunar surface fines and rock samples (Nagata et al., 1971; Zussman, 1972; Weeks, 1972), with combined iron present as the paramagnetic Fe^{2+} ion. The ferromagnetic component of lunar samples is primarily metallic iron which is sometimes alloyed with small amounts of nickel and cobalt (Nagata et al., 1972b; Pierce et al., 1971). This free iron is thought to be native to the moon (because of its low nickel content) rather than meteoritic in origin (Strangway et al., 1973a). Orthopyroxene and olivine models are consistent with geochemical studies (Urey et al., 1971; Wood et al., 1970; Ringwood and Essene, 1970; Green et al., 1971) and geophysical studies (Töksoz, 1973).

Since the susceptibility of free iron changes several orders of magnitude at the iron Curie temperature (T_c), Parkin et. al. have used a two-layer model with the core-shell boundary R_c at the Curie isotherm (see Figure 16). For $R > R_c$, $T < T_c$, and for $R \leq R_c$, $T > T_c$. Therefore, for $R > R_c$ any free iron is ferromagnetic while at greater depths where $T > T_c$, the free iron is paramagnetic. The Curie isotherm location is determined from the thermal profile used for a particular model. Three thermal models have been used in the calculations. For model profile T_1 the Curie isotherm is spherically symmetric and located at $R_c/R_m = 0.9$. Shell and core temperatures are $600^\circ C$ and $1400^\circ C$, respectively. For the model profile T_2 the shell is $500^\circ C$ and the core is $1300^\circ C$, while the Curie isotherm boundary is at $R_c/R_m = 0.85$. Temperatures are $300^\circ C$ and $700^\circ C$ for shell and core of model profile T_3, which has $R_c/R_m = 0.7$. In the outer shell there are both ferromagnetic and paramagnetic contributions to the total magnetic permeability $\mu_1 = 1 + 4\pi k_1$. The susceptibility of the shell is $k_1 = k_{1c} + k_{1a}$, where k_{1a} is "apparent" ferromagnetic susceptibility and k_{1c} is paramagnetic susceptibility. The ferromagnetic component is metallic free iron, assumed to be composed of multidomain, noninteracting grains; the paramagnetic component is Fe^{2+} combined in the orthopyroxene or olivine rock matrix. The measured ferromagnetic susceptibility of the shell material is an apparent value which differs from the intrinsic ferromagnetic susceptibility of the iron because of self-demagnetization of the iron grains and the volume fraction of iron in the shell. For $R < R_c$ the lunar material is paramagnetic only, with susceptibility $k_2 = k_{2c} + k_{2a}$; k_{2c} is the contribution of paramagnetic chemically combined iron and k_{2a} is the apparent susceptibility of free paramagnetic iron above the Curie temperature.

From the magnetic properties of lunar compositional and thermal models, Parkin et al., (1974a) calculated iron abundances for the moon which were consistent with measured global permeability. Their results are summarized in Table 2. The <u>minimum</u> total iron abundance consistent with the hysteresis curve can be calculated assuming the whole-moon permeability corresponds entirely to ferromagnetic iron in the outer shell where the temperature is below the Curie point. For this case the bulk iron abundance is 0.9 ± 0.5 wt. %. It is noted that the susceptibilities of both olivine and pyroxene are about an order of magnitude too small to account for the measured permeability without some ferromagnetic material present.

C. <u>Considerations of an Iron Core and Iron-Rich Layer</u>

The whole-moon permeability was also used by Parkin et al. (1974a) to investigate the magnetic effects of a hypothetical iron core in the moon. Density and moment of inertia measurements for the moon limit the size of such a core to less than 500 km in radius (Toksöz, 1973). If this hypothetical iron core were entirely paramagnetic and the surrounding core were orthopyroxene of average temperature 1100°C the global permeability would be 1.0003. This value is small compared to the measured permeability of 1.012 ± 0.006, implying that if such a small paramagnetic iron core exists, its magnetization is masked by magnetic material lying nearer to the surface. Therefore the hysteresis measurements can neither confirm nor rule out the existence of a small iron core in the moon.

An iron-rich layer in the moon has been considered by several investigators (Wood et al., 1970; Urey et al., 1971; Gast and Giuli, 1972). It is possible that early melting and subsequent differention of the outer several hundred kilometers of the moon may have resulted in the formation of a high-density, iron-rich layer beneath a low-density, iron-depleted crust. Constraints have been placed on an iron-rich layer by Gast and Giuli (1972) using geochemical and geophysical data (for example, measurements of lunar moments of inertia). One set of their models consists of high-density layers between depths of 100 km and 300 km. At a depth of 100 km the allowed layer thickness is 12 km; the thickness increases with increasing depth, to 50 km at 300 km depth. Also presented are a set of layers at 500 km depth. Using exactly the same considerations as were used in the iron abundance calculations Parkin et. al. have calculated whole-moon permeabilities which would be expected from lunar models with these iron-rich layers. The calculations indicate that all iron rich layers allowed by geophysical constraints as outlined by Gast and Giuli, if wholly above the iron Curie temperature, would yield global permeabilities of about 1.00006. As for the case of a small lunar iron core, the magnetization field of such paramagnetic layers would be masked by ferromagnetic materials elsewhere in the moon, and the hysteresis curve measurements can neither confirm nor rule out these layers. This conclusion would particularly apply to the Gast-Giuli layers at 500 km depths, which are almost certainly paramagnetic. If the iron-rich layers are below the Curie temperature and therefore ferromagnetic, they yield global permeabilities of about 3.5. This is above the upper limit for the measured permeability of 1.012 ± 0.006 and the Gast-Giuli layers can be ruled out if they are cool enough to be ferromagnetic. It is important to realize that the high density layers discussed by Gast and Giuli (1972) can be thought of as limiting cases and that there are innumerable less dense and thinner layers which are allowed by geophysical, geochemical and magnetic constraints.

V. GLOBAL EDDY CURRENT INDUCTION: ELECTRICAL CONDUCTIVITY AND TEMPERATURE

Electrical conductivity and temperature of the moon have been calculated from global eddy current response to changes in the magnetic field external to the moon. When the moon is subjected to a change in the external field, an eddy current field is induced in the moon which opposes the change (see Figures 18 and 19). The induced field responds with a time dependence which is a function of the electrical conductivity distribution in the lunar interior. Simultaneous measurements of the transient driving field (by Explorer 35) and the lunar response field (by an Apollo surface magnetometer) allow calculation of the lunar conductivity. Since conductivity is related to temperature, a temperature profile can be calculated for an assumed compositional model of the lunar interior. Fig. 18 19

When the moon is in the solar wind, lunar eddy current fields form an induced lunar magnetosphere which is distorted in a complex manner due to flow of solar wind plasma past the moon, as illustrated in Figure 16. The eddy current field is compressed on the dayside of the moon and is swept downstream and confined to the "cavity" on the lunar nightside. Because of the complexity, early analysis included a theory for transient response of a sphere in a vacuum to model lunar response as measured on the lunar nightside (Dyal et al., 1970c; Dyal and Parkin, 1971c; Sill, 1972) and a harmonic theory of a sphere totally confined by the solar wind to model response as measured on the lunar dayside (Sonett et al., 1971, 1972; Kuckes, 1971; Sill, 1972). Both the transient and the harmonic techniques have subsequently been further developed. Transient analysis has evolved to include effects of cavity confinement on nightside tangential data and to introduce analysis of magnetic step transients measured on the lunar dayside (Dyal and Parkin, 1973; Dyal et al., 1973). Harmonic analysis has been developed with the purposes of eventually developing a dynamic response theory for the case of asymmetric confinement (Schwartz and Schubert, 1973) and of extending the data analysis to lower frequencies and correcting for diamagnetic effects in the solar wind (Kuckes et al., 1974).

Recently time-dependent poloidal response of a sphere in a vacuum has been applied to data measured in the geomagnetic tail where plasma confinement effects are minimized. In this section lunar electrical conductivity from transient-response and harmonic analysis results are reviewed with emphasis on the former technique, and preliminary results of transient response analysis in the geomagnetic tail are presented. Also, representative temperature profiles determined from electrical conductivity analyses are presented and compared with temperatures derived from other analytical techniques.

A. <u>Electrical Conductivity Analysis: Moon in Solar Wind Plasma</u>

1. Lunar Nightside Data Analysis: Transient Response

The lunar electrical conductivity has been investigated by analysis of the lunar response to transients in the solar wind magnetic field. The response, measured by an Apollo magnetometer on the nightside of the moon, is theoretically approximated by the response of a conducting sphere in a vacuum. The theory was developed by extending the work of Smythe (1950) and Wait (1951) for a radially varying lunar conductivity profile (Dyal et al., 1972b). The measured response (illustrated in Figure 20a) is the Fig. 2 average poloidal field response (for the radial surface field component) to a normalized fast-ramp decrease in the external field. Error bars are standard deviations of the measured responses. Chosen for the analysis is a ramp input function which falls from unity to zero in 15 seconds, a time characterizing convection of a solar wind discontinuity past the Moon. (For a 400 km/sec solar wind, this time is 10-20 sec, depending on the thickness of the discontinuity and the inclination of its normal to the solar wind velocity.)

For a family of conductivity profiles, all of which monotonically increase with depth in the moon, the theoretical response to a fast ramp is calculated and compared to the measured response. A particular set of these conductivity profiles yield response functions which pass within all data error bars of Figure 20a. These profiles define the shaded region of Figure 23 and are all consistent with the nightside response data.

2. Lunar Dayside Data Analysis: Transient Response.

Theoretical solutions for an eddy current field totally confined to a sphere of homogeneous conductivity are derived from Maxwell's equations in Dyal et al. (1973). Figure 20b shows averages of normalized rising step transients measured on the lunar dayside in response to increasing step transients in the free-streaming solar wind (error bars are standard deviations). The overshoot maximum is amplified by a factor of 5 over the external input field step change, by solar wind dayside confinement of the surface tangential field components. The data are fit by a lunar conductivity model with a homogeneous core of radius $R_c = 0.9R_m$ and conductivity $\sigma \sim 10^{-3}$ mhos/m. This result is consistent with the nightside conductivity profile illustrated in Figure 23 to depths allowed by the duration of the response data which is shown in Figure 20b.

3. Lunar Dayside Data Analysis: Harmonic Response

The harmonic analysis of the lunar conductivity has been applied to data taken on the lunar dayside. The theoretical modeling is based on the assumption that any global induced field is excluded from the oncoming solar wind by currents induced in the highly conducting solar plasma; it is assumed that in effect the solar wind completely confines the induced field in the lunar interior and in a thin region above the lunar surface. The confinement current is considered to be a surface current and provides a boundary condition of total confinement by currents around the whole moon, permitting solution of Maxwell's equations at the lunar surface. This spherical confinement case is then applied to magnetometer data measured on the lunar sunlit side. Basic theoretical development of harmonic solutions can be found in several references (Backus and Gilbert, 1967; Parker, 1970; Schubert and Schwartz, 1969; Sill and Blank, 1970; Kuckes, 1971).

Once the poloidal fields have been derived, the analytical technique requires calculating frequency-dependent transfer functions which are defined as follows:

$$A_i(f) = \frac{b_{Ei}(f) + b_{Pi}(f)}{b_{Ei}(f)} \tag{5}$$

$i = x,y,z$, where the A_i are transfer functions of components of frequency-dependent magnetic fields, expressed in the orthogonal coordinate system with origin on the surface of the moon ($\hat{x}$ is radial and $\hat{y}$ and $\hat{z}$ are tangent to the surface); $b_{Ei}(f)$ is the Fourier transform of the external driving magnetic field; and $b_{Pi}(f)$ is the Fourier transform of the induced global poloidal field.

The harmonic data analysis involves Fourier-analyzing simultaneous data from the Apollo 12 lunar surface magnetometer, taken during lunar daytime, and the lunar orbiting Explorer 35 magnetometer.

Then ratios of the surface data to orbital data are used to calculate a transfer function given by equation (5).

The form of the transfer function is determined by the internal conductivity distribution in the moon; therefore, a "best fit" conductivity profile can be obtained by numerically fitting the measured and theoretical transfer functions. Figure 23 includes the "best fit" conductivity profile of Sonett et al. (1971) obtained in this manner. The conductivity profile is characterized by a large "spike" of maximum conductivity about 1500 km from the lunar center. Other conductivity profiles have been calculated using the Sonett et al. (1971) transfer function, showing that the spike profile is not unique but that the frontside transfer function can be fitted by simpler two- and three-layer models (Kuckes, 1971; Sill, 1972; Sonett et al., 1972; Reisz et al., 1972; Phillips, 1972).

B. Electrical Conductivity Analysis: Moon in the Geomagnetic Tail

The theoretical models outlined so far have all assumed spherical symmetry to describe lunar eddy current response to changes in the external field. However, analyses to date have used data taken when the moon is immersed in the solar wind plasma with asymmetric confinement of the induced fields. The shortcomings of using spherically symmetric approximations to describe the induced lunar magnetosphere, which is actually asymmetrically confined, have been pointed out in the literature for both the nightside vacuum approximation (see, e.g., Schubert et al., 1973a) and the dayside totally-confined approximation (see, e.g., Dyal and Parkin, 1973). Three dimensional, dynamic asymmetric confinement presents a difficult theoretical problem which has not been solved at the time of this writing. Previous theoretical approximations of the asymmetric problem have included a two-dimensional approximation (Reisz et al., 1972); three-dimensional static theory for a point-dipole source, with substantiating laboratory data (Dyal and Parkin, 1973); a three-dimensional "quasi-static" approach (Schubert et al., 1973b); and a three-dimensional dynamic theory for one particular orientation of variations in the external field (Schwartz and Schubert, 197 In order to circumvent this problem, recent analysis has considered lunar eddy current response during times when the moon is in the geomagnetic tail (see Figure 19), where plasma interaction effects encountered in the solar wind (asymmetric confinement, remanent field compression, plasma diamagnetism, etc.) are minimal.

1. <u>Poloidal Response of a Sphere in a Vacuum: Theory</u>

To describe the response of the lunar sphere to an arbitrary input field in the geomagnetic tail, we define the magnetic vector potential $\underline{A}$ such that $\underline{\nabla} \times \underline{A} = \underline{B}$ and $\underline{\nabla} \cdot \underline{A} = 0$. We seek the response to an input $\Delta\underline{B}_E b(t)$, where $b(t) = 0$ for $t < 0$ and $b(t)$ approaches unity as $t \to \infty$. (Since the governing equations are linear, the response to a more general input is readily found by superposition.)