Magnetic Fields of the Terrestrial Planets

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The four terrestrial planets, together with the Earth's Moon, provide a significant range of conditions under which dynamo action could occur. All five bodies have been visited by spacecraft, and from three of the five bodies (Earth, Moon and Mars) we have samples of planetary material upon which paleomagnetic studies have been undertaken. At the present time, only the Earth and Mercury appear to have a significant dipole magnetic field. However, the Moon, and possibly Mars, appear to have had ancient planetary dynamos. Venus does not now have a significant planetary magnetic field, and the high surface temperatures should have prevented the recording of evidence of any ancient magnetic field. Since the solidification of the solid inner core is thought to be the energy source for the terrestrial magnetic field, and since smaller bodies evolve thermally more rapidly than larger bodies, we conjecture that the terrestrial planets are today in three different phases of magnetic activity. Venus is in a predynamo phase, not having cooled to the point of core solidification. Mercury and the Earth are in the middle of their dynamo phase, with Mercury perhaps near the end of its activity. Mars and the Moon seem to be well past their dynamo phase. Much needs to be done in the study of the magnetism of the terrestrial planets. We need to characterize the multipole harmonic structure of the Mercury magnetic field plus its secular variation, and we need to analyze returned samples to attempt to unfold the long-term history of Mercury's dynamo. We need to more thoroughly map the magnetism of the lunar surface and to analyze samples obtained from a wider area of the lunar surface. We need a more complete survey of the present Martian magnetic field and samples from a range of different ages of Martian surface material. Finally, a better characterization of the secular variation of the terrestrial magnetic field is needed in order to unfold the workings of the terrestrial dynamo.

INTRODUCTION

The solar system provides the dynamo researcher with a broad spectrum of dynamo types. The solar magnetic field is generated in the Sun's outer shell. The inner planets, which for our purposes includes the Earth's moon, provide examples of incipient, present-day and extinct dynamos operating in liquid metal cores. Early magma oceans on these planets may have generated solar-like dynamos. The outer planets provide examples of dynamos in metallic hydrogen cores and in water-ice shells. In this review we will examine part of this spectrum of dynamos, that part provided by the bodies of the inner solar system, leaving the outer planets and the sun to other authors of this special issue.

The inner planets present us with a suite of bodies of different sizes, having different states of thermal evolution, and with similar (but not identical) density and composition. The rotation rate of each of these bodies appears to be sufficient for dynamo action to take place. The Earth's moon should be considered as part of any study of inner solar system dynamos, since there is abundant evidence of ancient lunar fields. While it is possible that these fields may have arisen in a solar-like shell dynamo operating in an early magma ocean, there is also evidence of an electrically conducting core that could have served as the host for an ancient lunar magnetic dynamo. Before proceeding to discuss each of the planets, in turn we present an oversimplified picture of the formation and thermal evolution of rocky, metal-rich planets in order emphasize the importance of planetary size in the generation of planetary magnetic fields and set the stage for the varied magnetic states in which we find these planets today.

Paper number 93JE00981.
0148-0227/93/93JE-00981$05.00

In the beginning as the solar nebula cooled, the standard model tells us that dust condensed and planetesimals formed by gravitational instability. These bodies were small solid objects which accumulated into a smaller number of larger objects through collisions. Near misses led to scattering and some amount of mixing. Thus, the heterogeneity, initially present in the solar nebula because of the radial temperature gradient, was reduced to some extent. On the other hand the present day inhomogeneity in the asteroid belt implies that some of the original inhomogeneity of the nebula should now be exhibited by the inner planets. Thus we expect the bodies of the inner solar system only to be similar, not identical, in composition.

For reasons that still remain veiled, the inner solar system formed with a small iron-rich planet, Mercury, closest to the Sun at 0.39 AU, of radius 2440 km. Further out, near 0.72 and 1 AU two nearly equal sized planets accumulated, Venus, of 6052 km radius, and Earth, of 6371 km radius. Further out at 1.54 AU, a much smaller body, Mars, accumulated. Between Mars and Jupiter, matter was able to condense into planetesimals and accumulate to some degree in larger aggregations but no body greater than 1/4 of 1% of the mass of Mars seems to have evolved.

The precise origin of the Earth's moon is difficult to discern. In the outer solar system some moons clearly accumulated in orbit about their parent bodies, much as the inner planets formed about the sun. Some moons may have been captured due to drag or collision with the gas, dust and particulate rings around the early planet and still other moons may have been captured in three-body gravitational interactions. While the second of these three possibilities could have sufficed for Deimos and Phobos around Mars, none of these scenarios seem to explain the Earth's moon. Instead it is generally believed that a giant impact generated the Moon [*Hartman and Davis*, 1975]. In this process, a Mars-sized object has been conjectured to strike the Earth leaving a Moon's mass of material to later condense in orbit around the Earth. This process is likely to result in a Moon that has a different average composition than the impactor, if the impactor had differentiated prior to impact, since material at different distances from the center of the Earth would follow different trajectories after impact. Thus, we would not be surprised to learn that the Moon was somewhat different in bulk composition from the Earth, perhaps even more so than Venus and Mars.

The varying composition of the bodies of the inner solar system certainly complicates any intercomparison but the varying size of the planets is probably a much more important factor. To appreciate the critical role of size, let us assume for the moment that the inner planets were all uniform in composition but of different size. In this mixture are gases trapped in the rocks, as well as water, silicates, metals and radioactive nucleides. Figure 1 illustrates that homogeneous initial state of these planets. They present us with a significant range of sizes which in turn should provide a significant difference in the rates of thermal evolution. The smallest, the Moon, has a radius about 1/4 of the largest, the Earth. The volume of the Moon is only 2% of that of the Earth and its mass is even a smaller fraction because the Moon is significantly less dense than the Earth.

If all this material were at the same temperature prior to accretion, it would not stay at the same temperature after accretion because of the gravitional potential energy released in collisions. Hence during accretion the larger bodies would become hotter than the smaller bodies. Since the gravitational energy release would become large only once the primary bodies became large, the heating takes place on the outside and not the inside. It is then reasonable to expect that bodies could be molten on the outside at sometime during their thermal history even if not molten on the inside. Since a surface cools between collisions, the radius at which melting and magma ocean formation would occur is a function of the collision rate. Since we have evidence that a magma ocean did form on the moon, we are confident that such a process occurred on the other bodies of the solar system as well.

Bodies should heat up on the inside as well, due to the radioactive decay of the radionucleides. This heating would proceed slowly compared to the release of gravitational potential energy during impacts on the surface. On the smaller bodies it might proceed so slowly that the interior never reached the molten stage. However, even if the interior remained solid, differentiation could occur because metals, especially iron, are very much denser than silicate. Hence, iron pools formed at the bottom of the Moon's magma ocean could have sunk through the underlying silicate and accumulated in the center of the moon without the silicate reaching the melting point [*Stevenson*, 1980]. On planets that more thoroughly melted this differentiation could take place more completely and more rapidly. Figure 2 illustrates this initial differentiated state qualitatively. The relative size of the core and the planet will depend on the initial relative abundance of iron (the most abundant metal) and the completeness of differentiation which will be a function of size as discussed above. Since we have evidence of a conducting core for the Moon, as we will discuss in a later section, and since we know that the Earth has a liquid metal core we feel confident that Mercury, Mars and Venus also

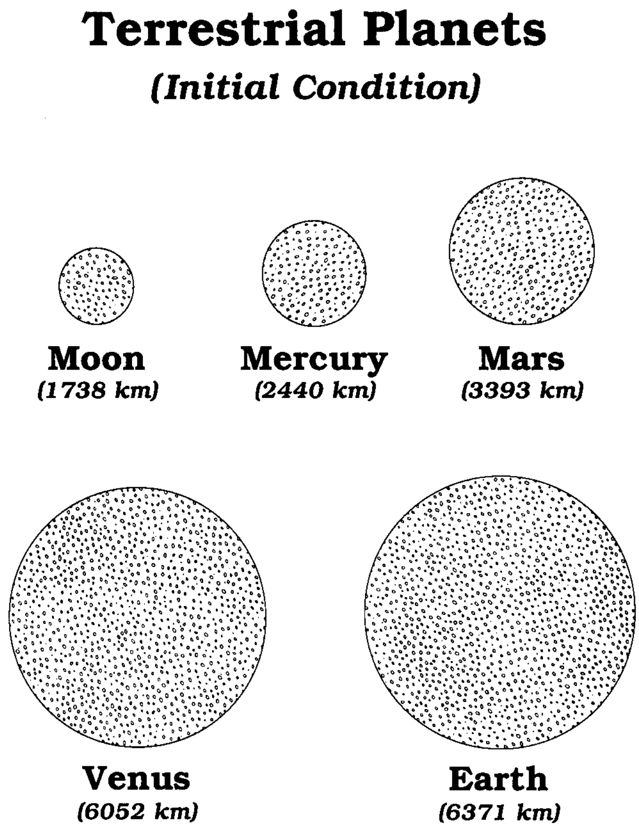

Fig. 1. The initial undifferentiated state of the five bodies of the inner solar system which we consider in this review. In our overly simple picture of planetary thermal evolution we assume that these bodies all began of roughly the same material and accreted homogeneously.

have or had at one time liquid metal cores. Moreover, models of the planetary interiors need such cores, albeit just barely in the case of the Moon, to explain the observed average density and the moment of inertia. We stress that each of these bodies probably reached the differentiated state at a different time with the larger planets differentiating first. The Moon probably differentiated last because its progenitor's impact with Earth is thought to have occurred late in the planetary formation process.

The core formation process also liberates gravitational potential energy as the heavy material migrates to the center of the planet, but eventually, sometime after core formation, the planet can begin to cool. Heat is transported from the interior by convection and conduction and radiated into space. The rate of cooling will depend on how rapidly heat can be transported to the surface and how fast the surface can radiate that heat into space. A small body can generally cool faster than a larger body. Thus, we might expect that, once differentiated, the Moon, Mercury and Mars might cool off most rapidly. Further cooling eventually will lead to the solidification of the core as illustrated in Figure 3. Iron will solidify at the top of the melt and then sink to the center releasing latent heat of fusion and more gravitational energy. This has certainly happened in the core of the Earth where there is a 1200 km radius solid core, but we know nothing of the state of the cores of the other planets except the implications drawn from the presence or absence of a dynamo-driven magnetic field. Again this will happen at a different time for each planet. Of the inner planets, only Mercury and the Earth are thought to have dynamo-driven magnetic fields today.

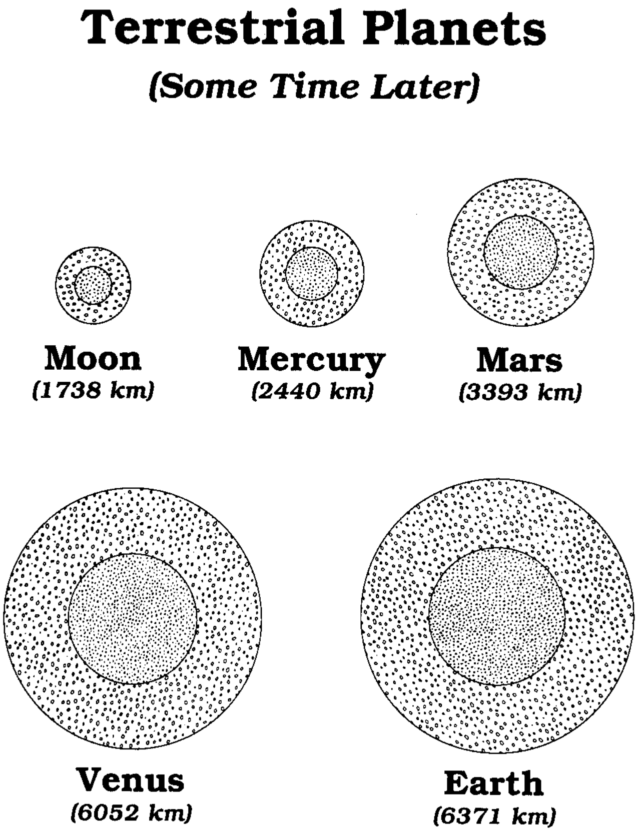

Fig. 2. The state of our five bodies in Figure 1 at some later time that is different for each body. Each of the bodies has differentiated to form an iron or iron alloy core. The largest body will differentiate most quickly. The size of the core will depend on the initial composition of the body and the degree to which differentiation takes place. Since the Moon is believed to have assembled from iron poor material, its core should be relatively smaller than the others.

In this review we will examine what we know about the magnetic fields of the inner planets and test these findings against this simple paradigm. We first examine some of the features of the terrestrial magnetic field to prepare us for what we have and have not learned in our studies of the terrestrial planets.

THE EARTH

The Earth is the most rapid rotator of the inner planets. It rotates once on its axis every 23 hours and 56 minutes (relative to the stars). Its average radius is 6371 km and its interior consists of a mantle 2886 km thick, a liquid metallic core of radius 3485 km and a solid core of radius 1221 km. By volume 0.7% of the planet is solid core, 15% liquid core and 84% mantle. Although the core contains only 16% of the volume, it contains about 36% of the mass of the planet. The energy released by the solidification of the core, through the settling of solid particles to the center of the planet and the heat released by the freezing process, is believed to be the principal sources of power for the dynamo.

The Earth's dynamo varies with time. We do not know when it started or whether it has ever stopped. However, the magnetic field measured on the surface of the Earth is surprisingly variable and it has reversed polarity innumerable times. If one were simply to divide the rate of change of components of the Earth's magnetic field (the secular variation) into the size of each of the coefficients used to describe the field, one would get a characteristic time of about 200 years with the characteristic time for the dipole term being a little longer, about 1000 years. The present rate of reversals (over the last 10 m.y.) is about once every 150,000 years so that the dynamo is much more stable to reversing its polarity than to variations in its amplitude. Even this rate is variable. Sixty million years ago it reversed once every 500,000 years. In terms of the age of the Earth, these rates of evolution are quite rapid.

On the surface of the Earth and above, the dipole component of the magnetic field is dominant, but this is not necessarily true in the core where the energy contained in the field at all orders of complexity (dipolar, quadrupolar, octupolar, etc.) are comparable. The dipole axis is presently tilted at 10.8° and since 1960 has been rapidly decreasing. For over 100 years prior to that time the dipole tilt remained steady at 11.5°. The dipole also drifts westward. In the last 400 years it has drifted over 40°. However, the rate of drift has not been constant. Over the same time the dipole field strength has decreased about 17%. The rate has appeared to quicken over the last few decades.

The Earth's magnetic dipole moment of about 7.8×10^{15} T m^3 is sufficient to stand off the solar wind far above the surface of the planet. The solar wind consists of electrons and protons with a density of about 7 cm^{-3} flowing at a velocity of around

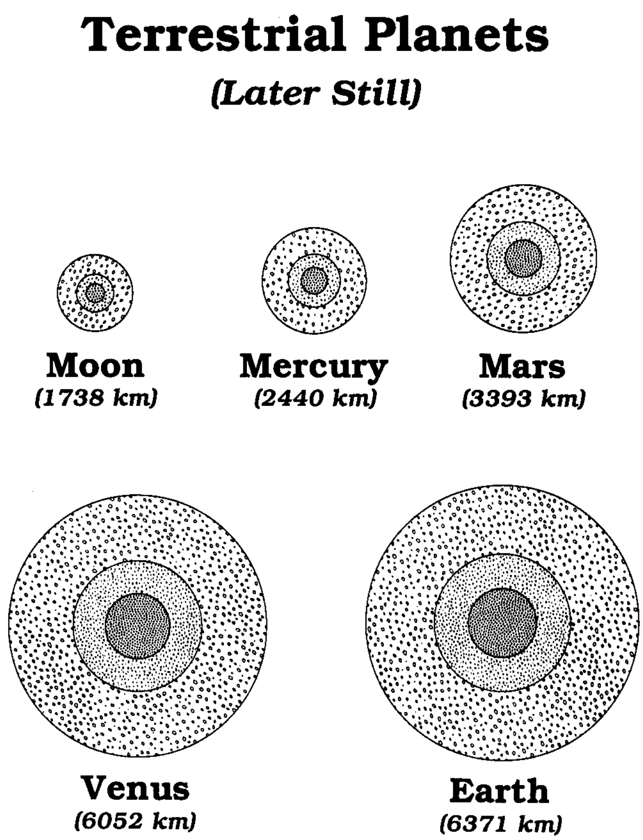

Fig. 3. A later stage of the thermal evolution of the bodies in Figure 2. The bodies now have cooled to the point that their inner cores have begun to solidify. We expect this cooling to occur most rapidly on the smallest bodies while core formation itself should occur most rapidly for larger bodies. Additional heat sources such as tidal forces and radioactivity could slow the cooling process.

440 km s^{-1}. The resulting momentum flux is balanced by the Earth's magnetic field about 10 Earth radii (R_E) above the planet. The magnetic cavity carved out of the solar wind is about 30 R_E across above the dawn-dusk terminators and about 50 R_E across near the orbit of the Moon. A magnetic tail, much like a cometary ion tail, forms in the antisolar direction stretching the order of 1000 R_E or more behind the planet. A sketch of this magnetosphere, as it is called, is shown in Figure 4.

The magnetosphere contains various populations of plasma. There are cold electrons and ions from the ionosphere which extend upward in the inner region of the magnetosphere to form the plasmasphere. There is a hot ion and electron region in the center of the tail, separating the two tail lobes that forms the plasma sheet. There are boundary layer plasmas near the edges of the magnetosphere, called the mantle, representing plasma that has entered the magnetosphere from outside. There is the plasma that enters the polar weak points in the Earth's magnetic configuration, called the polar cusps. This configuration is dynamic on time scales of minutes to hours because the energy transfer to the magnetosphere from the solar wind is strongly modulated by the direction of the interplanetary magnetic field [*Russell and McPherron*, 1973]. When the interplanetary magnetic field is southward, opposite to the direction of the Earth's magnetic field in the subsolar region, the process known as reconnection occurs. The magnetic field lines of the magnetosphere and solar wind become linked and maximal energy transfer ensues. If the rate of energy transfer is sufficiently small, this energy can be stored in the Earth's magnetic tail as a buildup in the magnetic flux. Eventually this storage region becomes unstable and releases the magnetic flux and its associated energy in the process known as a substorm. Through this process the aurora are energized and the ionosphere disturbed.

The solar wind travels relative to the Earth at a velocity greater than the velocity of compressional waves that are necessary to deflect the solar wind around the Earth. Thus to deflect the solar wind a standing shock wave forms, called the bow shock. The bow shock slows, deflects and heats the solar wind. The region between the shock and the solar wind is called the magnetosheath.

In order to measure the intrinsic magnetic field of the terrestrial planets we must make the measurements in such a dynamic environment. These measurements are easiest at Earth, where the intrinsic field is strong, where we make measurements far from the influence of the solar wind and where we have many measurements. However, the measurements at other terrestrial the planets and especially on the Moon are made much more difficult because of the weakness of the intrinsic magnetic fields there.

THE MOON

Our understanding of the Moon benefits greatly from the Apollo program and, at the same time, languishes because of post-Apollo backlash. All our in situ lunar data were obtained at an early stage of planetary exploration when our instrumentation was far less sophisticated. Moreover, the Apollo program explored only a limited part of the lunar surface. After the Apollo project it has proven impossible to return to the moon because its "exploration" value has been diminished. This is truly unfortunate because much could be learned about our nearest neighbor with a modicum of resources.

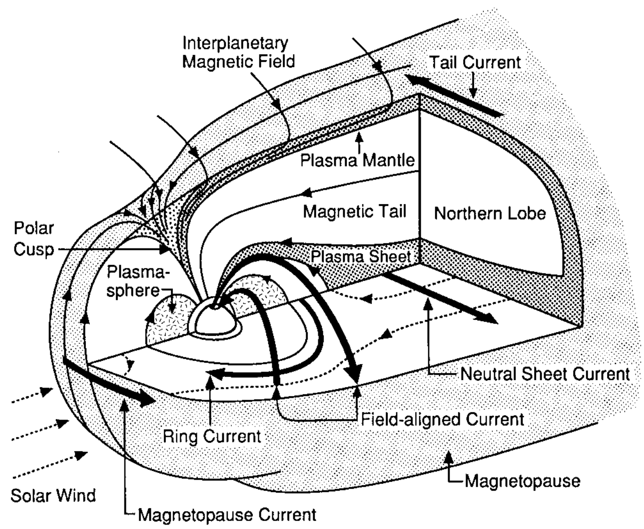

Fig. 4. The solar wind interaction with the Earth's magnetic field. This interaction is typical of those of all the magnetized planets, differing principally in scale size from planet to planet. The bow shock which stands in the solar wind flow in front of the obstacle has been omitted from this diagram.

Despite the placement of seismometers on the Moon by the astronauts, the interior structure and the size of the core remain luna incognita. The average density of the moon is known to be 3.344 $\pm$ 0.002 [*Blackshear and Gapcynski*, 1977] but this is not enough to distinguish models with a 400 km FeS core from core-free models with some extra density in the mantle. The limited seismic data did not place severe enough constraints on mantle density or the location of the core-mantle boundary. The seismic data were limited both because the seismometers were placed only over the limited region on which astronauts were able to land and because the seismometers were turned off to "save money."

Magnetometer data have been used to sound the electrical conductivity of the lunar interior, also with limited success. A metallic core, either liquid or solid, will be highly electrically conducting and will exclude any applied magnetic field for up to several hundred years depending on the size of the core and the precise value of conductivity. Thus any variation in the magnetic field that occurred over time scales as short as those of the Apollo program would be excluded from the core. Two techniques were attempted to measure these induced fields. The first used an orbiting magnetometer on Explorer 35 together with magnetometers on the surface of the moon. Different investigators obtained core radius upper limits of from 350 to 570 km [*Dyal et al.*, 1976; *Hood et al.*, 1982]. The second technique used measurements from low altitude orbit while the moon was in the Earth's tail [*Russell et al.*, 1981]. The amount of exclusion of the magnetotail field then provided an estimate of the size of the conducting core. The resulting core size was 439$\pm$22 km. However, no information was obtained on the possible composition of this conducting core. In short, the body of evidence supports the existence of a small metallic lunar core of radius up to about 500 km but occupying no more than about 2% of the lunar interior.

It was evident from the earliest lunar observations that the lunar magnetic field was weak so that the later discoveries of the magnetism of the lunar rocks [*Runcorn et al.*, 1970; *Strangway et al.*, 1970] and the magnetic fields observed at the Apollo landing sites [*Dyal et al.*, 1974] were surprises to most scientists. Figure 5 shows that the interaction of the solar wind with the lunar surface is much different than the interaction of the solar wind with the Earth's magnetosphere. The solar wind overcomes all magnetic pressure exerted by magnetic fields on the sunward side of the Moon and is absorbed by the surface leaving a wake behind the moon. The solar wind, that flows on either side of the moon and is not absorbed, then attempts to close behind the moon. If the magnetic field is aligned with the flow, this closure is inhibited. It was this interaction which provided the first indication of the magnetization of the lunar surface. When magnetized regions rotated to the terminator region, they were strong enough to deflect the solar wind, causing compressions of the magnetic field and plasma to be observed just upstream of the expansion fan [*Mihalov et al.,* 1971; *Russell and Lichtenstein*, 1975].

The information about the lunar magnetic field from these limb compressions was limited at best. A more quantitative means of probing the lunar magnetic field was direct observations using magnetometers from altitudes of about 100 km and below with the Apollo subsatellites while the Moon was in the Earth's magnetotail. Figure 6 shows a series of magnetic measurements made as the spacecraft flew over the far side of the moon near the crater Aitken. As the spacecraft slowly crosses the region on successive passes the signature evolves. Also seen are some transients that appear on a single orbit. These are easily distinguished with the data from successive orbits.

While it was impossible to map the large scale lunar field from orbit with this technique it was possible to map the fine scale field using the Apollo subsatellite magnetometers as shown in Figure 7 [*Russell et al.,* 1977] and more qualitatively with the electron reflectance technique [*Lin et al.,* 1976]. Hood and colleagues have used these data to estimate the location of possible paleopoles of an ancient lunar dynamo [*Hood et al.,* 1978*a,b*; *Hood,* 1980, 1981]. While the totality of these poles appear random, subsets of the poles appear to group and may indicate that the pole of the ancient lunar dynamo wandered as does the Earth's dipole axis [*Runcorn,* 1982, 1983]. Moreover, if one takes various estimates of the magnetic field strength to which lunar rocks were exposed, as estimated from studies of their paleomagnetism, and then compares this strength with the age of the rocks, one finds that the inferred paleointensities decrease as a function of age in the period from 3.9 to 3.0 Ga before the present. Rocks older than 3.9 Ga and younger than 3.0 Ga seem to have experienced reduced or no magnetizing magnetic fields [*Runcorn*, 1987].

In summary, lunar observations are less constraining than we would desire, in part because of the incompleteness and crudeness of our Apollo and pre-Apollo data. However, there are strong indications in the lunar record that there is a small lunar core, certainly presently inactive. Since the lunar surface is extensively but weakly magnetized, and since paleomagnetic intensities seem to be significant for a discrete interval of time, about 0.9 Ga in duration from 3.9 to 3.0 Ga B.P., we believe that an ancient lunar dynamo likely existed for this period of time. Alternately, or additionally, a dynamo in the outer molten layers of the Moon may have been generated [*Russell,* 1986]. This author remains skeptical of the deductions of ancient paleopoles and ancient polar wandering. The measurement of the surface field from orbit was difficult with the Apollo subsatellites and may not have been accurate enough for such inversions. This author also remains skeptical of impact mechanisms for magnetizing large volumes of lunar crust. The observations of lunar magnetization are tantalizing but much more, and higher quality data need to be obtained before it can provide the constraints on lunar thermal evolution we wish it would provide.

MERCURY

Mercury is the smallest of the inner planets. Its radius of 2440 km places it close to the geometric mean of the radii of the Moon (1738 km) and Mars (3395 km). However, it is much denser than either of these two bodies indicating that it has a larger relative core size than either of them. Mercury has an average density of 5.4 g cm^{-3} compared with the Moon's 3.3 g cm^{-3}. Thus if the mantle were the same density as lunar material and the core the same density as the terrestrial core, the core would occupy about 20% of the volume of the planet and about 60% of the radius.

Mercury rotates more slowly than the Moon and much more slowly than Mars. It rotates with a period of 59 days which is 2/3 of its orbital period of 88 days. Thus every second orbit about the Sun a point on Mercury's surface returns to its same location relative to the Sun. Mercury travels in an elliptical orbit of moderately high eccentricity (e=0.206) and a 7° inclination to the ecliptic. Unlike the Earth its rotation axis is aligned with the normal to its orbital plane. Thus Mercury does not have seasons as the Earth and Mars do except as occasioned by the varying heliocentric distance (0.31 to 0.47 AU) as it moves around its highly eccentric orbit.

Outwardly Mercury looks very much like the Moon because of its heavily cratered terrain. It has no dynamically significant atmosphere that could cause erosion of the surface or even significantly couple to the magnetospheric plasma. Thus the control of the Mercury magnetosphere by the solar wind could differ significantly from that of the Earth. Mercury does have

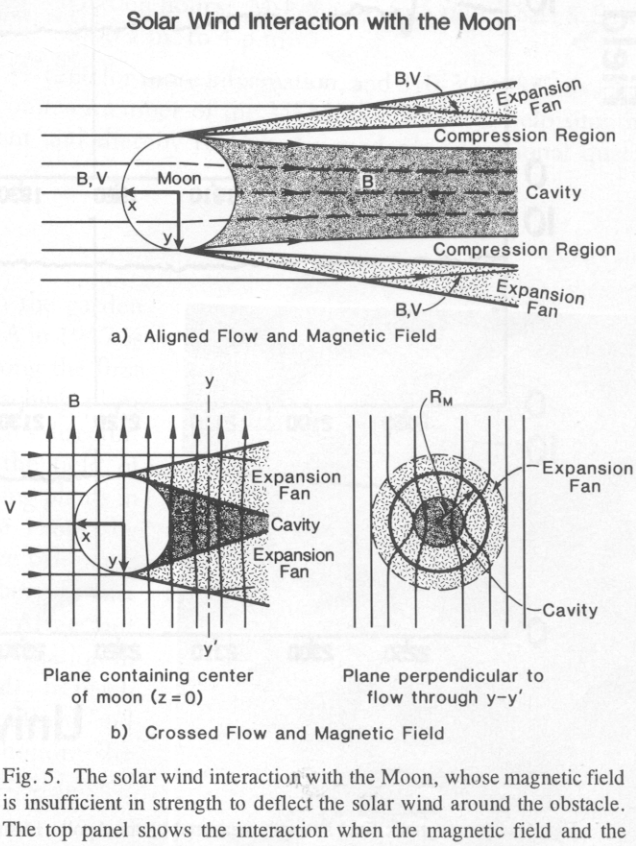

Fig. 5. The solar wind interaction with the Moon, whose magnetic field is insufficient in strength to deflect the solar wind around the obstacle. The top panel shows the interaction when the magnetic field and the flow are aligned. The bottom two panels show orthogonal views of the interaction when the magnetic field is perpendicular to the flow.

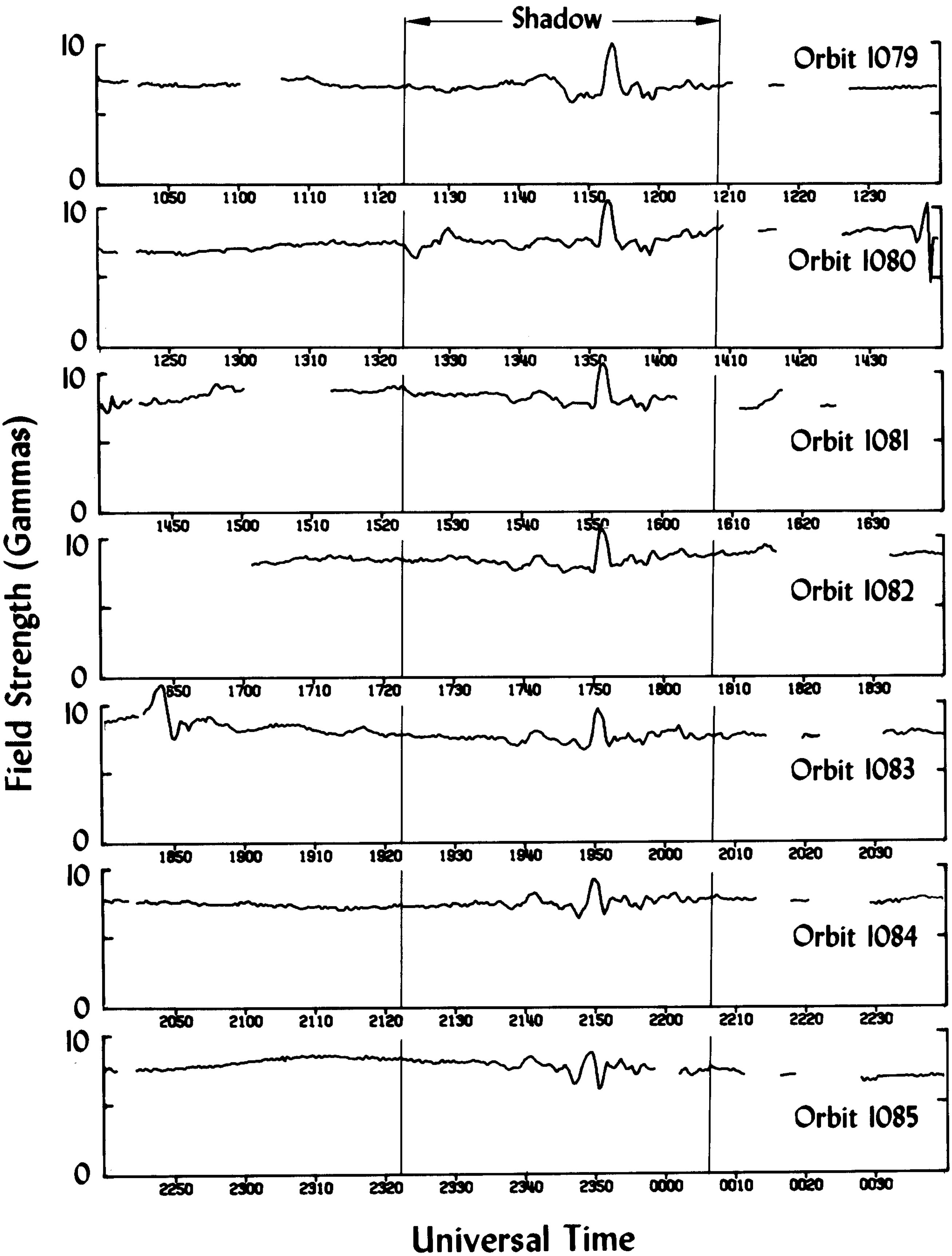

Fig. 6. The magnetic field strength measured by the Apollo 15 subsatellite on seven successive orbits around the Moon, illustrating the repeated signature of lunar surface magnetic fields.

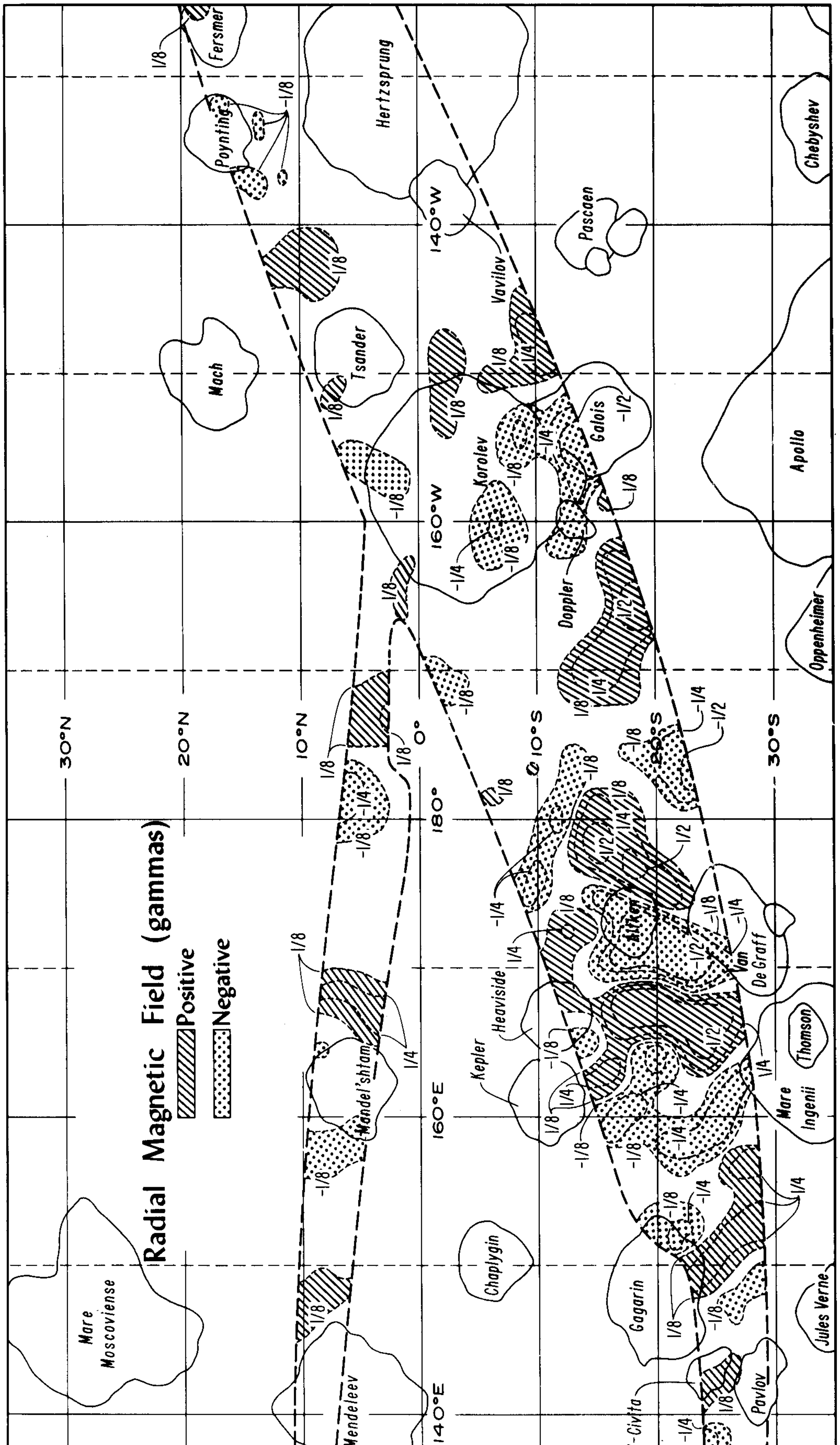

Fig. 7. The radial component of the fine scale lunar magnetic field as mapped at about 100 km by the Apollo subsatellite over the far side of the Moon.

some atmosphere, however weak, and that atmosphere may come from the interior of the planet and therefore may indicate that the planet is tectonically active. Most importantly Mercury has an intrinsic magnetic field, one that is too large to be due to remanent magnetization and this clearly indicates that the interior of Mercury is active.

Mercury has been visited by only one spacecraft, Mariner 10, which passed by the planet three times between March 1974 and March 1975, the first and third of which were on the nightside of the planet. Figure 8 shows the magnetic field observations observed on the first pass which crossed the dark side of the planet, approaching within 723 km of the surface [*Ness et al.*, 1975]. The maximum field strength observed was nearly 100 nT at closest approach. The lack of a significant atmosphere ensures that the observed magnetic field was not induced as a result of the solar wind interaction with an ionosphere but the large size of the magnetosphere is also an indication that the magnetosphere is intrinsic and not induced. As at the Earth, there is a bow shock in front of the magnetic cavity and a magnetotail stretched out behind the planet. After Mariner 10 passed the point of closest approach (CA) it encountered a disturbance in the magnetosphere much like a terrestrial substorm [*Siscoe et al.*, 1975]. No such dynamics were seen on the third pass which also passed closer to the planet (327 km) and measured a field of over 400 nT [*Ness et al.*, 1976].

Many authors have attempted to model the intrinsic and magnetospheric fields from these two passes. Figure 9 shows one of these attempts, based on calculations of *Jackson and Beard* [1977]. As is clearly evident, the magnetic configuration resembles the terrestrial magnetosphere and is strongly affected by the magnetospheric currents over the entire region probed by Mariner. Thus, it is unlikely that one could accurately deduce the planetary magnetic field from only two passes through the night side. This ambiguity was formally addressed by *Connerney and Ness* [1988].

Figure 10 illustrates the problem. This figure shows the

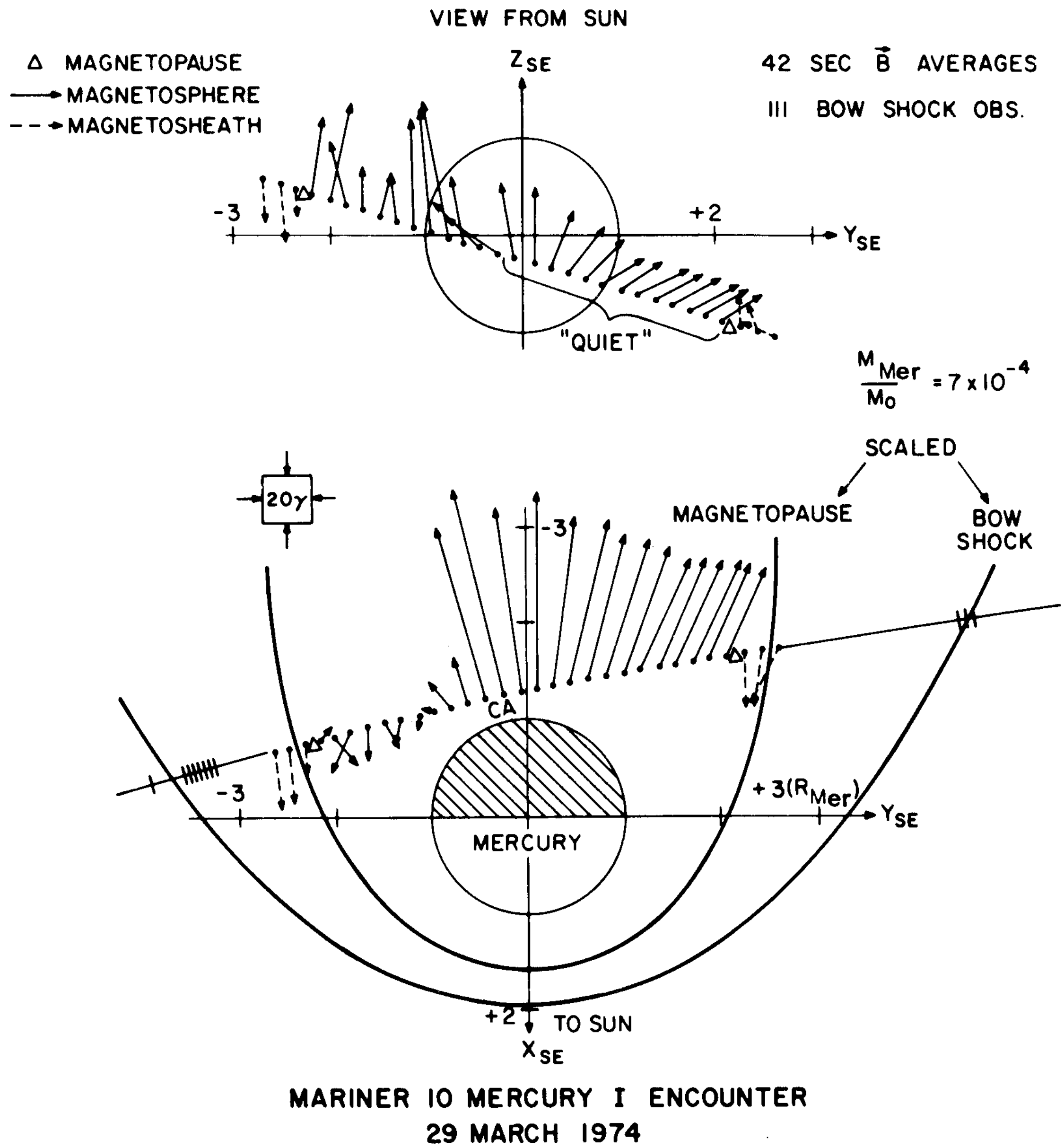

Fig. 8. The initial Mariner 10 measurements of the Mercury magnetic field. The top panel shows the trajectory rotated into the ecliptic plane [*Ness et al.*, 1975].

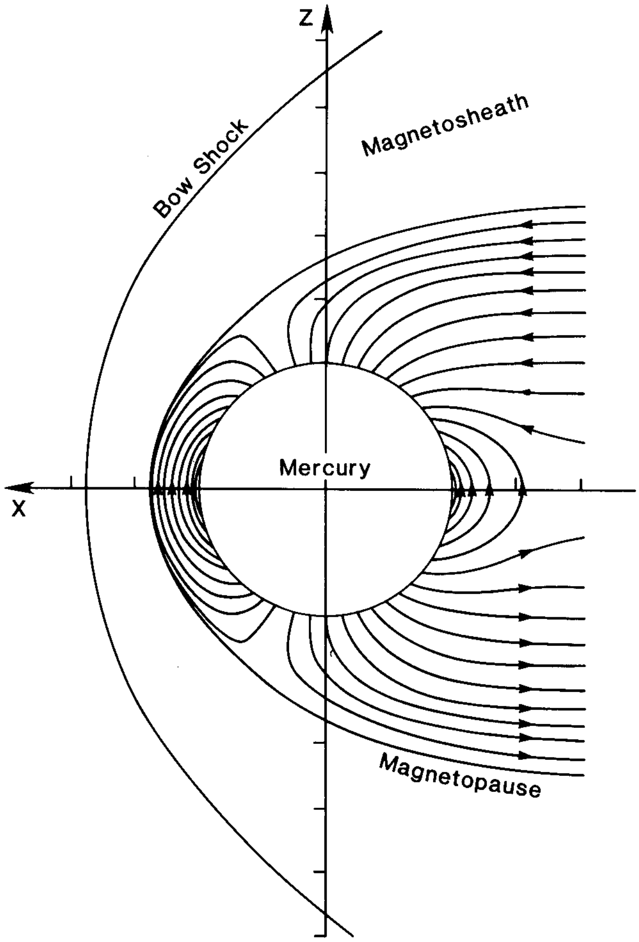

Fig. 9. The noon-midnight meridian of the Mercury solar wind interaction. Magnetic field model of *Jackson and Beard* [1977] has been used.

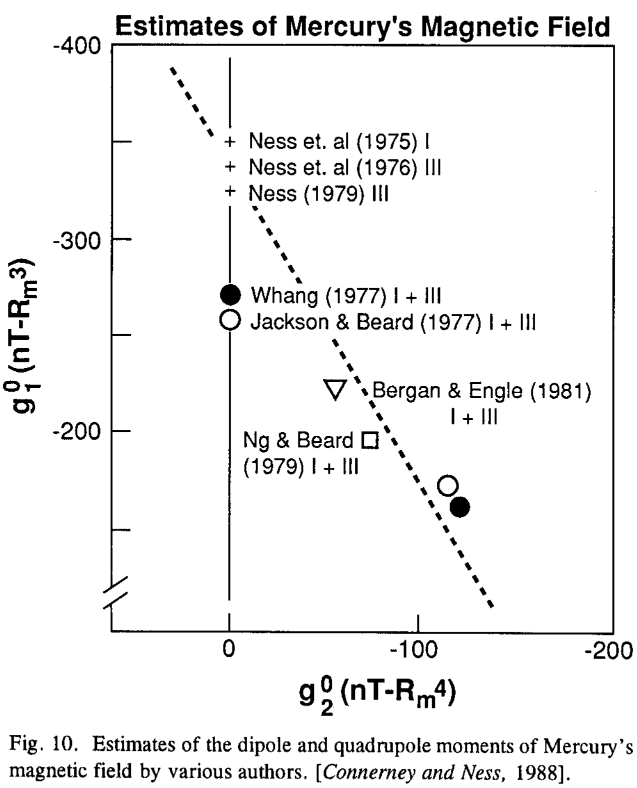

Fig. 10. Estimates of the dipole and quadrupole moments of Mercury's magnetic field by various authors. [*Connerney and Ness,* 1988].

dipole moment deduced by various authors versus the quadrupole moment obtained. Clearly the dipole moment varies over a factor of 2 and is correlated with the quadrupole moment. *Connerney and Ness* [1988] examined possible combinations of optimized dipole and quadrupole moments of varying sizes. The residuals of these models are shown in Figure 11 in the same format as Figure 10. The optimum solution closely follows the fits found by other authors but illustrates that there is no way to distinguish between these fits. Because the data were obtained only over a limited region, it is impossible to deduce the dipole moment of Mercury to within a factor of two. Thus the magnetic dipole moment of Mercury is most probably somewhere between 2 and 5×10^{12} Tm^3 and cannot be determined more accurately without further observations. Moreover, other parameters that are critical to determining the behavior of the internal field and the dynamo such as the higher moments and all information on secular variations are completely absent. To characterize the magnetic field at a level that would be useful to those studying planetary interior or dynamos requires orbital measurements at low altitudes in a polar orbit. Paleomagnetic samples from polar sites that may have avoided intense thermal cycling are also highly desirable.

VENUS

Venus is sometimes called the Earth's twin because of its proximity (0.72 AU heliocentric distance) and similarity to the Earth. Because it has a similar size (6052 km radius) and density (5.25 g cm^{-3}), it is believed to have a similar interior structure to that of the Earth, with a core whose radius is about 1/2 Venus radius (R_v). However, we have no seismic data with which to check this conjecture. Venus rotates very slowly in a retrograde direction with a sidereal period of 243 days while it orbits the sun in 224.7 days. This is still believed to be sufficient to sustain dynamo action. Nevertheless, despite the presence of an iron core and sufficient rotational velocity for dynamo action Venus most decidedly does not have a planetary magnetic field.

The first limits on the planetary magnetic moment were placed by Mariner 2 in 1962 which passed Venus at a distance of 6.6 R_v and detected no evidence of an Earth-like magnetosphere. Mariner 5 passed much closer to the planet (1.4 R_v) in 1967 and detected an induced magnetosphere but no clear evidence of an intrinsic magnetic field. Thus the magnetic moment had to be less than 10^{-3} of the terrestrial moment [*Bridge et al.,* 1969]. Later measurements on the Venera 4 bus down to 200 km, and on the Venera 9 and 10 orbiters confirmed this small moment. The most definitive measurements of the Venus magnetic environment were made by Pioneer Venus which spent nearly 14 years in orbit around Venus beginning in December 1978. During the first three Venus years of operation sufficient orbital control gas was available to maintain periapsis near 150 km. The periapsis passages in the night ionosphere of Venus should have permitted the detection of magnetic moments of magnitude up to 10^{-5} of the terrestrial moment, yet no such moment was found [*Russell et al.,* 1980; *Phillips and Russell,* 1987].

What Pioneer Venus did observe was a long induced magnetic tail. Figure 12 shows where this tail was encountered at apoapsis. The direction of the magnetic field in the tail was either toward or away from the sun together with a small component across the tail in the direction of the upstream

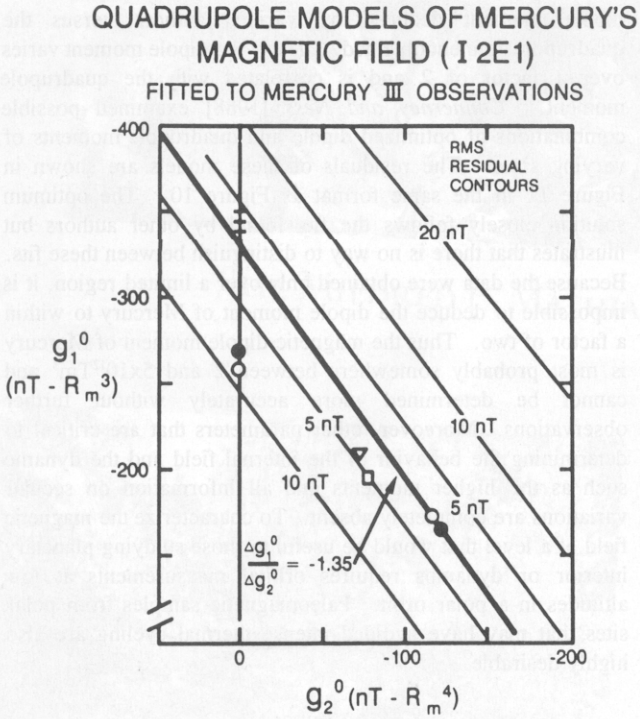

Fig. 11. Contours of the residuals of Mercury magnetic field models when compared with the observed field for varying dipole and quadrupole moments [*Connerney and Ness*, 1988].

magnetic field perpendicular to the flow. As the direction of the interplanetary field varied so did the tail lobe field structure. No evidence for an intrinsic planetary contribution was found at high or at low altitudes. Figure 13 shows how such an induced tail is formed. Upstream of the bow shock the interplanetary magnetic field is convected toward the shock. When it crosses the bow shock the plasma slows down near the planet but the plasma on the same field lines far from the planet does not slow down so the magnetic field lines are bent around the planet. The magnetic field lines that are carried closest to the planet are bent the most and are carried into the center of the tail. Slowing of the flow occurs both because of the planetary obstacle that must be avoided and because ions from the planet are added to the flow. The greater the mass addition, the greater the flow is slowed.

While the highly conducting ionosphere is the ultimate barrier to the flow, the magnetic field carried by the flow and draped across the ionosphere forms a magnetic barrier whose cross section is somewhat larger. The bow shock therefore is found a distance upstream which depends on the size of the ionospheric obstacle, the size of the magnetic barrier on top of it and the Mach number of the solar wind flow which determines how compressed is the solar wind plasma that has to flow around the obstacle. Needless to say as the solar cycle waxes and wanes the ionosphere, magnetic barrier and Mach number all change and the bow shock moves significantly [*Russell et al.*, 1988]. As a result it is difficult to use the location of the bow shock as any indication of the size of the planetary obstacle and to infer the possible intrinsic magnetic moment [*Russell et al.*, 1992*b*] as has been attempted by several authors at Mars.

The weakness of the present field does not mean that Venus has always been bereft of an intrinsic field nor that it always will be. It is possible that, during the years close to the time of core formation when there was sufficient energy available for dynamo generation, Venus had a magnetosphere similar to that of the Earth, but that this field decayed as the available energy in the core decayed. We cannot check this hypothesis through paleomagnetic magnetic studies as we could on the surfaces of the other bodies of the inner solar system because the surface temperature is well above the blocking temperature below which rocks retain any magnetic memory.

A possible reason why Venus has not continued to generate a magnetic field like the Earth has generated has been suggested by *Stevenson* [1983] who hypothesizes that the slightly smaller size of Venus is sufficient to alter the temperature-pressure structure of the interior of Venus as illustrated in Figure 14 so

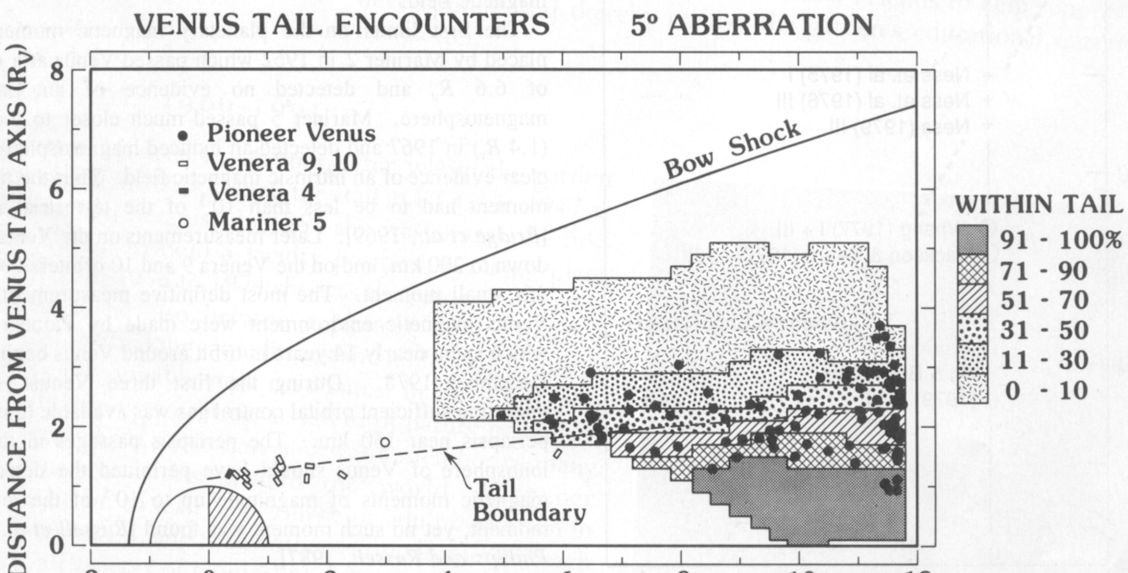

Fig. 12. Occurrence rate of magnetotail encounters as a function of distance along and across the Sun-Venus line [*Saunders and Russell*, 1986].

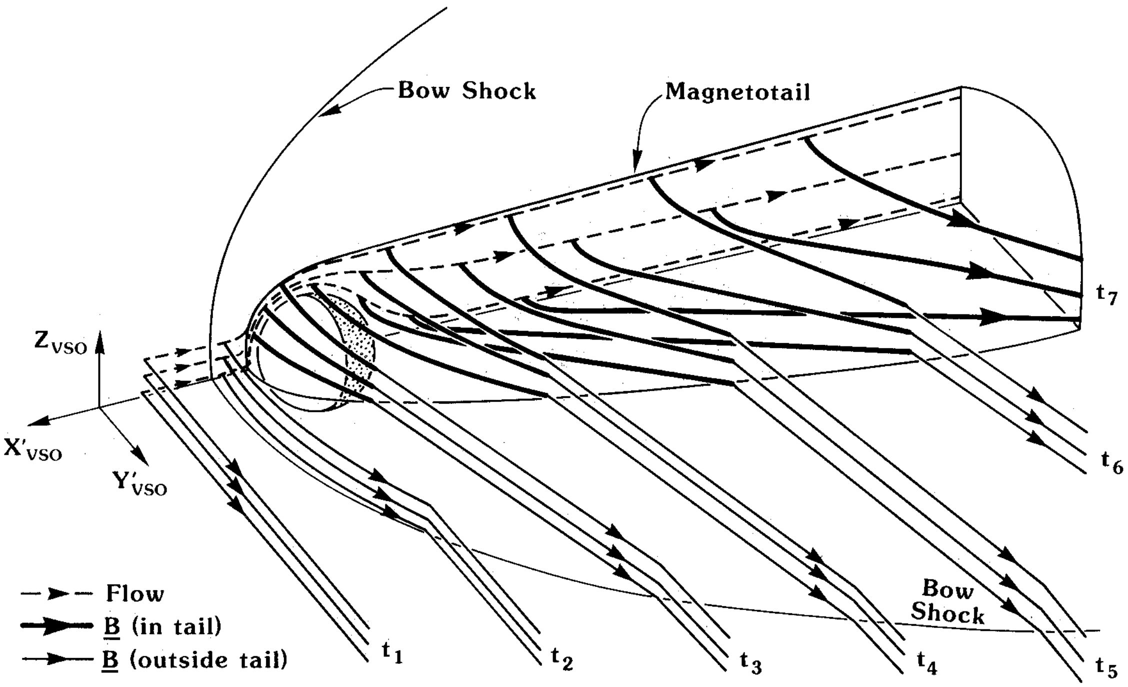

Fig. 13. The process of induced tail formation. Magnetic field lines are carried to Venus by the solar wind plasma. Far from the planet, the field lines are carried along at the undisturbed solar wind velocity, but close to the planet the flow slows down, and the field lines become bent [*Saunders and Russell,* 1986].

that the T-P curve remains above the freezing curve. If there is no solidification of an inner core as on the Earth, the energy available may be just insufficient to drive dynamo action. This of course raises the question of what will happen in the future. Eventually as Venus cools further the interior of Venus should begin to solidify and in the process we expect it to begin to generate a magnetic field.

At the present time the absence of an intrinsic magnetic field at Venus provides only the crudest limits on the structure and dynamics of the interior. In fact to learn more about why Venus has no measureable intrinsic magnetic field we need to learn more about the interior from other techniques. Hence long-lived surface measurements of long period seismic waves are of the highest priority for Venus.

MARS

Despite the many missions to Mars in both the American and Soviet programs, the question of the existence of an intrinsic Martian magnetic field is still controversial. Mars resembles the Earth in its spin period of 24 hours and 37 min and the 25° tilt of its spin axis (obliquity) relative to its orbital plane. However, Mars is much smaller than the Earth, with a radius of only 3394 km and lighter, with a density of only 3.9 g cm^{-3}. This latter value is greater than the lunar density of 3.3 g cm^{-3} and implies a larger iron core than the moon, all else being equal, perhaps 1500 km in radii. Such a core would occupy only about 9% of the volume of the planet. Only two seismometers have been placed on the Martian surface, and unfortunately neither instrument functioned as well as planned. There are no technical problems in performing these measurements. They should be of the highest priority on any future mission to the Martian surface. Mars clearly has been an active planet in the past. The large volcanoes provide evidence of that. It is possible the planet is still quite active.

The so-called SNC meteorites are now widely believed to be samples of the Martian surface and, as such, have provided us with many constraints on the geochemistry and evolution of

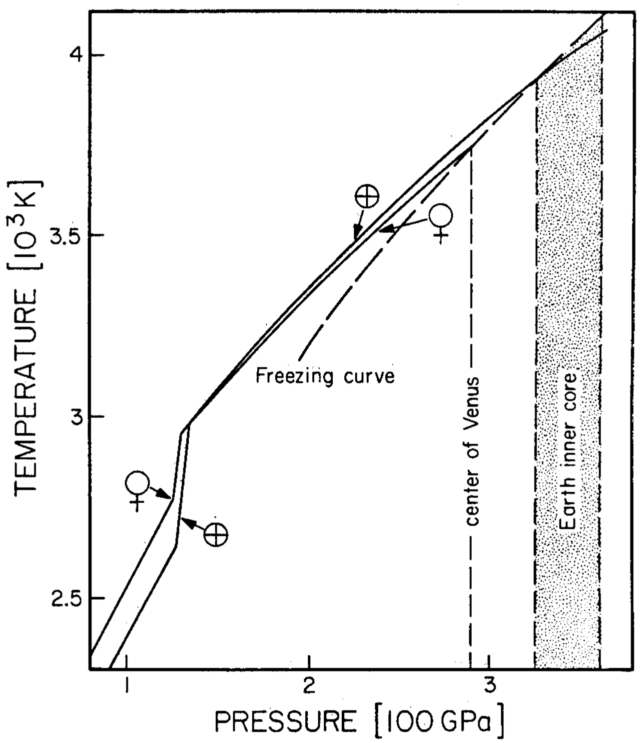

Fig. 14. Temperature pressure curve inside the Earth and Venus illustrating why the core of Venus may not yet be starting to freeze [*Stevenson,* 1983]. The pressure-temperature relationship that defines the liquid-solid phase change intersects the terrestrial curve well above the center of the planet, but at Venus intersects only at the center of the planet.

Mars. Figure 15 shows the process by which the SNC meteorites are believed to have been liberated from the Martian surface to reach the Earth. A large body is believed to have impacted a 1.3 Ga lava flow about 180 Ma ago. Some of the resulting space debris further collided in space and eventually fell to Earth. It therefore would be surprising if these rocks had any clear imprint of any earlier Martian magnetizing field.

Paleomagnetic studies of the SNC meteorites have been carried out by *Collinson* [1986] and *Cisowski* [1986]. These studies suggest that these rocks may have been magnetized by an ancient magnetic field of possibly as high as 10,000 nT but also possibly 1,000 nT or or less. However, they do not provide unambiguous evidence for such a field. In view of the history of these meteorites, these studies are encouraging but should not be viewed as definitive evidence for or against an active dynamo 1.3 Ga B.P.

In situ measurements with the Martian spacecraft, the Mars 2, 3, and 5 spacecraft and the Phobos spacecraft clearly show that the present day magnetic field is weak. If Mars had an active dynamo, we might expect it to have a magnetic moment intermediate between that of the Earth and Mercury since its rotation rate is similar to that of the Earth, and its core is similar in size to that of Mercury. The fact that it does not suggests that, if there ever was an internal dynamo, that dynamo has stopped. Nevertheless, our expectations on the size of the dynamo field could be wrong, or the planet could possess a strong remanent magnetic field. Thus, it is instructive to examine the evidence for and against even weak intrinsic fields.

The first limits were placed on the intrinsic magnetic field of Mars by the Mariner 4 mission by the relatively small size of the obstacle that Mars presented to the solar wind [*Dryer and Heckman*, 1967; *Smith* 1969]. If we assume that the encounter occurred under normal solar wind conditions, we arrive at an upper limit of 2×10^{12} T m^3 from this encounter.

Mars 2 and 3 were injected into Martian orbit on November 27 and December 2, 1971, respectively. The periapsides of the orbits of these spacecraft to Mars were closer than Mariner 4's closest approach. On one orbit of Mars 3, the experimenters interpreted the magnetometer and plasma data as entry into an intrinsic magnetosphere. These data are shown in Figure 16 [*Dolginov et al.*, 1972, *Gringauz et al.*, 1974]. The lower traces show the components of the magnetic field. The upper traces show the electron temperature and current. The experimenters interpreted these data in terms of the passage through an Earth-like solar wind interaction as illustrated in Figure 4. Point 1 was taken to be the bow shock crossing. Points 2 and 3 were taken to be the entry and exit from the intrinsic magnetosphere, and point 4 was taken to be the exit through the bow shock again on the way out. *Dolginov* [1978] inverted these data to obtain a magnetic moment of 2.4×10^{12} T m^3, but the results of the inversion were quite sensitive to the interval chosen. *Wallis* [1975] pointed out that the region identified as the bow shock could instead be the foreshock where electrons stream back into the solar wind along field lines attached to the shock. *Russell* [1978*a*] pointed out that the "intrinsic magnetosphere" might be field lines piled up against the planet by the solar wind. Later, *Russell et al.*, [1984] used the convected magnetic field gas dynamic numerical simulation of the solar wind flow with Mars to show that the field measured by Mars 3 resembled the draped magnetic field of the magnetosheath pressed against the planetary obstacle. The lowest estimate of all was obtained by *Russell* [1978*b*] who suggested that all the tail encounters of Mars 5 were with an

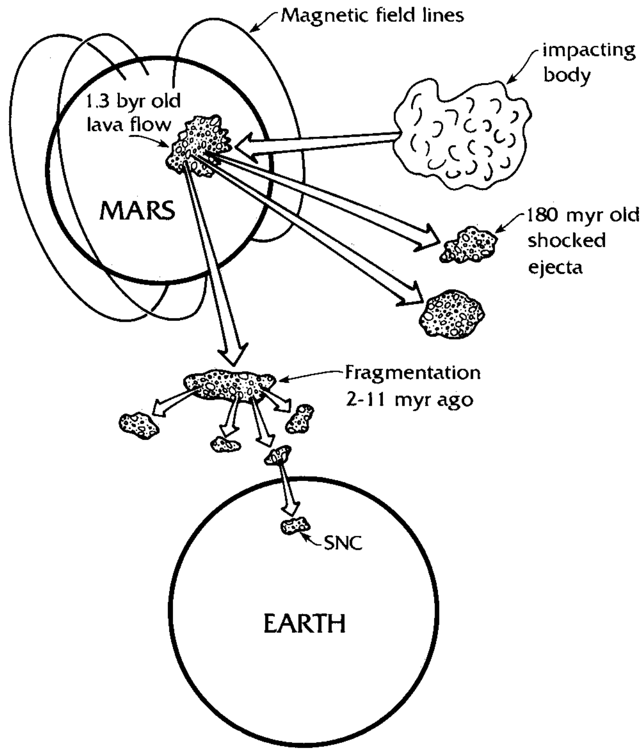

Fig. 15. The inferred history of the SNC meteorites.

induced rather than intrinsic magnetotail, and hence the moment could be as low as 10^{11} T m^3 or a factor of 25 less than Dolginov's estimate.

The controversy about the size of the Martian magnetic moment was not settled by any of these arguments. Clearly, more data were needed, especially data closer to the planet. In 1989, the Phobos spacecraft did provide such data [*Riedler et al.*, 1989] with the closest approaches to within 800 km of Mars, and then over a month of data in an 8-hour circular orbit that carried the spacecraft through the tail repeatedly. No clear evidence for an intrinsic field was seen on any of these close passes. Furthermore, the direction of the tail magnetic field which was sampled repeatedly in circular orbit as illustrated in Figure 17 was found to be controlled by the direction of the interplanetary magnetic field in the same manner as the tail of the Venus [*Yeroshenko et al.*, 1990].

Despite the fact that the magnetic moment was clearly much less than previously estimated, many authors continued to interpret the observation in terms of a planetary magnetosphere. *Dolginov and Zhuzgov* [1991] simply inverted measurements close to the planet to get a moment of 1.4×10^{12} T m^3. *Mohlmann et al.* [1991] claimed that spectral peaks at 8, 12 and 24 hour peaks proved that Mars had an intrinsic field. *Verigin et al.* [1991] found that the radius of the Martian tail depends on the solar wind dynamic pressure in much the same way as the size of the Earth's magnetosphere depends on the solar wind dynamic pressure. Finally, *Slavin et al.* [1991] pointed out that the bow shock is found on occasion far in front of Mars, perhaps caused by an expanding magnetosphere at a time of low solar wind dynamic pressure. None of these arguments provides conclusive evidence for an intrinsic magnetic field, however. *Russell et al.* [1992*a*] has shown that 8, 12 and 24 hour period spectral harmonics can occur in the interaction with a nonmagnetized planet. *Gringauz et al.* [1993] found that Venus

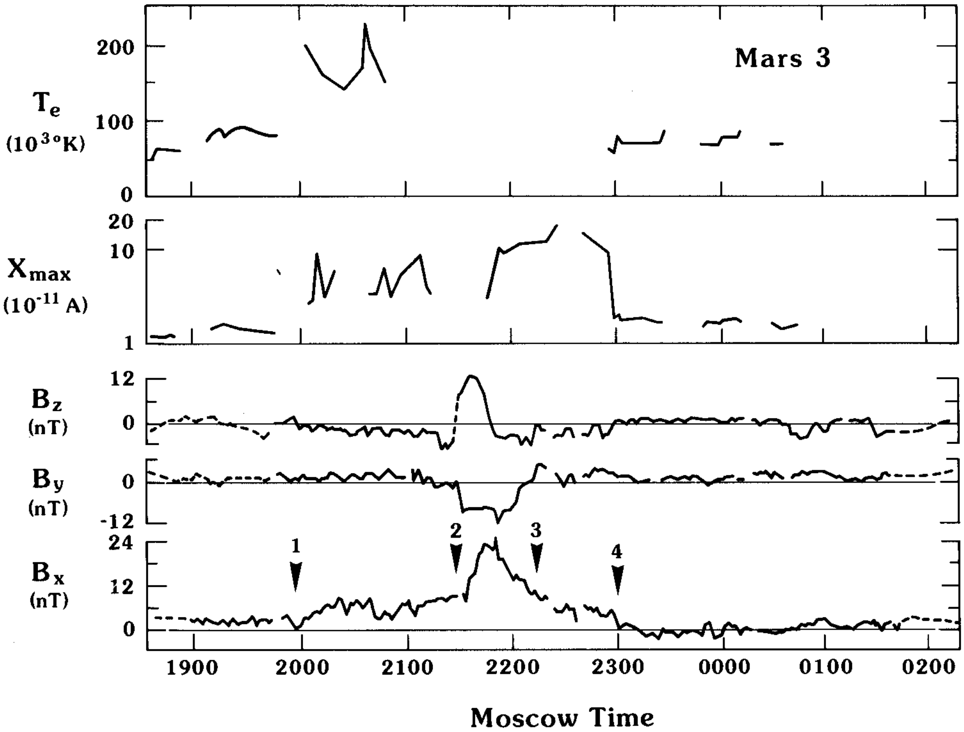

Fig. 16. (Top) Electron temperature and current measurements and (bottom) magnetic field components for an orbit of Mars 3 during which it was supposed to have entered the Martian magnetosphere. The Mars 3 electron spectrometer faces away from the sun, and hence is sensitive to backstreaming electrons [after *Gringauz et al.*, 1974].

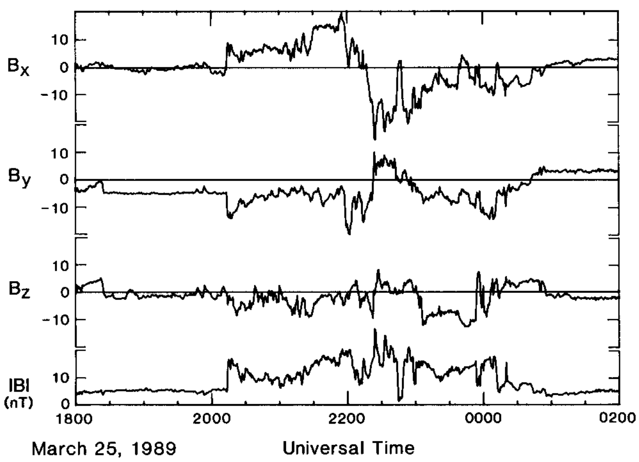

Fig. 17. The vector magnetic field and its magnitude measured by the Phobos spacecraft on one of the circular orbits about Mars.

has the same tail radius as Mars for the same solar wind pressure, and *Russell et al.*, [1992*b*] showed that the bow shock can move anywhere upstream of a planet when the Mach number approaches unity; and also, the ionosphere expands in much the same way as a magnetosphere when the solar wind pressure drops.

Again, the study of the Martian magnetic field would have benefitted from a lower periapsis and much greater coverage. However, the mission did lower the estimate of the magnetic moment of Mars by even the strongest intrinsic field advocates by a factor of 2 to 4. In the judgement of this observer, the true moment must be 10^{11} T m^3 or less, since the tail is so responsive to the solar wind direction. It now remains for the Mars Observer spacecraft and the Mars 94 missions to try to lower these estimates and to try to determine what is the level of magnetization that our intuition tells us must be in the Martian rocks.

SUMMARY AND CONCLUSIONS

Of the five major bodies of the inner solar systems, we have sample materials for paleomagnetic studies from three of them: the Earth, the Moon, and Mars (through the SNC meteorites). These samples reveal or at least suggest the presence of significant magnetic fields at each of these bodies during some time in the past. Even though our space probes reveal no evidence for such significant planetary fields today, we believe that it is extremely likely that both the Moon and Mars had a significant dynamo driven magnetic field in the past. We have no information on the possible existence of ancient magnetic fields on Venus. The surface of Venus is too hot to have retained an imprint of ancient magnetic fields, and we have no known samples of Venus surface material. Venus may have had an early dynamo, perhaps driven by the heat of accretion, but it has no significant dynamo at present. As Venus continues to cool, its core may become active as an inner core starts to form, but this activity seems not to have yet begun. Finally, Mercury, despite its small size, seems not to have yet cooled sufficiently that its dynamo has ceased. Unfortunately for Mercury, we do not have any retrieved samples for paleomagnetic analysis, we do not know the dipole moment within a factor of 2, have no idea of the harmonic structure of the magnetic field, and no constraints on the secular variation. Eventually, we will need all these data in order to understand the processes underlying the Mercury dynamo.

In summary, it appears that the inner solar system has provided us with all the material we need to study the temporal evolution of dynamos. We have two bodies, the Moon and Mars, which seem to have once had active dynamos, but which have now ceased. We have two bodies, Mercury and the Earth, with presently active dynamos, perhaps of different relative ages, and we have one body, Venus, whose dynamo has not yet started up. To achieve this potential, we need more data, however. At Mercury, we need to make a detailed low altitude magnetic survey, to do this several times to determine the secular variation, and to return paleomagnetic samples. At Mars, we need to make surface measurements and return unshocked paleomagnetic samples, and we need a more complete survey of lunar magnetism. For Venus, all we can do is wait.

Acknowledgment. **This work was supported by the National Aeronautics and Space Administration under research grant NAGW-2573. This work was presented at the "Planetary and Solar Dynamos" meeting at AGU's All-Union Session in Montreal, spring 1992.**

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(Received December 3, 1992;
revised April 7, 1993;
accepted April 13, 1993.)