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Planetary Magnetism

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1. INTRODUCTION

That the Earth is a magnet has been known with varying degrees of insight for about a millenium, but it was not until slightly over 30 years ago that radio telescopes were able to reveal that any of the other planets had magnetic fields (Burke and Franklin, 1955), and only the Jovian magnetic field was discovered in this manner. The investigation of planetary magnetism has depended by and large on *in situ* observations by space probes. There have been many surprises in these investigations. It was clear from the Luna and Explorer 35 missions that the Moon possessed no significant global magnetic field. However, the returned lunar samples were found to be magnetized, and magnetometers brought to the lunar surface revealed an extensively magnetized crust (cf. Chapter 4). Clearly the Moon once had a significant global field. The next major surprise was at Mercury, not much larger than the Moon, and from all external appearances very Moon-like. Mercury had a global field similar to that of the Earth, albeit much weaker (Ness *et al.*, 1974). This field was sufficiently large that its source must be an active dynamo.

There have been surprises at the other planets, too, but these surprises have been because the magnetic moments were less than expected rather than greater. The two remaining terrestrial planets have no significant planetary magnetic moment. Saturn, even though it has a radius that is only 17% smaller than that of Jupiter, has a magnetic moment that is over 30 times smaller (Smith *et al.*, 1980). Finally, although the exact mechanism behind the generation of a planetary magnetic field is not completely understood, it seemed quite apparent that asymmetry with respect to the rotational axis was important, if not crucial. Thus it was thought that the approximately 11° tilt of the terrestrial and Jovian moments was fundamental to the field generating process. Saturn, however, has no such tilt of its dipole axis. Its moment seems to be nearly exactly aligned with its rotational axis. Uranus, on the other hand, has a moment that is inclined 60° to its rotation axis.

There are four major methods of probing the interior of a planet, and magnetic measurements provide two of them. Measurements of the changes of the orbits of planetary spacecraft allow a determination of the structure of the gravity field and the moment of inertia of a planet. This in turn allows inferences about the radial density structure of the planet. Seismic waves measured by seismometers placed on the surface of a planet are the ideal instrument for probing the interior structure of a planet. However, this technique cannot be used on the giant planets because of their lack of an accessible solid surface and is difficult to implement on Venus because the very high surface temperature mitigates against long-term monitoring. In fact, seismometry has been used successfully on only one other extraterrestrial body, the Moon. Electromagnetic sounding can be used to probe the interior of a planetary body. In terrestrial applications an array of magnetometers are deployed across the surface of the region whose deep electrical conductivity is to be studied. The differences observed in oscillations with periods of many minutes to hours is then used to infer variations in electrical conductivity. On the Moon, two slightly different techniques were used. The first, using two instruments, was used to measure the outer conductivity, one of these instruments was on the surface and one in orbit about the Moon. The second, using one magnetometer orbiting at low altitude, measured the distortions in an otherwise uniform field to deduce the deep interior lunar conductivity. Again, electrical conductivity sounding has been used only on the Earth and on the Moon.

The fourth technique for learning about the interior of a planet is to study its intrinsic magnetic field. If a planet has a strong global field, it must be generated by an internal electrical dynamo. The existence of that dynamo requires both an electrically conducting fluid core and a significant source of energy, for example the growth of a solid inner core. At a very minimum, the existence of a strong intrinsic magnetic field signifies that the planetary core has not frozen completely and in fact is still strongly convecting. Our first indication that Mercury was still internally active came from a measurement of its intrinsic field, a decade before significant quantities of sodium were detected (Potter and Morgan, 1985). The detection of sodium can also be interpreted as signifying an internally active planet.

We should like to learn much more about the interior of a planet from studies of its external magnetic field but we are hampered by our lack of detailed understanding of the dynamo process. For example, we cannot yet accurately infer anything about the interior of a planet from a knowledge of the strength of its magnetic dipole moment. Alternatively, even if we knew all the relevant properties of the interior of a planet, we could not today predict what the magnetic moment would be. However, some progress is being made. A recent proposal by Stevenson (1984) states that the existence and nature of a planetary dynamo depends on the energy flux number, which is directly proportional to the available energy flux and the radius of the core and inversely proportional to its density, angular velocity and the square of the magnetic diffusivity. In this model, for energy flux numbers less than 1 there is no dynamo. For numbers between 1 and about 300 there is an energy-limited dynamo. Mercury may be in this state. Between 300 and 100 000 there is a dynamically determined dynamo whose magnetic field has little sensitivity to the available heat flux. The Earth may be in this state. For energy flux numbers above 10^5 the dynamo becomes turbulent. Jupiter's dynamo is thought to be in this state.

Information is also contained in the harmonic content of the field. Since the strength of higher-order magnetic moments falls off faster than the low-order moments, the relative strength of the different orders at the surface of the planets gives us information on the depth of the source region of the field (Elphic and Russell, 1978). A much more accurate technique is to use the fact that in the very highly electrically conducting core of a planet magnetic field lines are 'frozen' to the conducting fluid. The observed secular variation of the field can then be used to deduce the depth of the conducting core. This has been applied to the Earth (Hide, 1978), but not successfully to other planets yet because of insufficient accuracy of the determination of the magnetic moments and length of the temporal baseline. The technique is a promising candidate for future use on missions to Jupiter and Mercury.

Studies of planetary magnetism provide information not only on the state of a planet today but also on how it used to be. If samples can be returned, as they were from the Moon, magnetic analyses can reveal the ancient paleointensities required to magnetize the sample and provide data on the history of the planetary field. Even in the absence of returned samples, measurements of crustal remanent magnetization can be used to infer ancient magnetic poles, and together with geological data, infer the history of polar wandering (cf. Runcorn, 1983).

Finally, of course, the study of planetary magnetism is an intensely interesting subject by itself. How are planetary dynamos powered? Are there several different mechanisms? What determines the harmonic content of the field, and its secular variation? What controls the strength of the dipole and its tilt? We live immersed in a strong magnetic field whose source we do not well understand. Many of our planetary neighbours have similarly strong magnetic fields. We are curious to know why.

Planetary magnetic fields can act to shield a planet and its atmosphere from the solar wind. While this is another important role of planetary magnetism, its principal concern for us is that this interaction distorts the planetary magnetic field, and forms a magnetic cavity called a magnetosphere. When the planet fills a large fraction of its magnetosphere it is difficult to determine the structure of the planetary contribution. This will trouble us at Mercury and Mars. Thus we shall spend some time in this review in describing the solar wind interaction where important.

The planets are treated below in order of their distance from the Sun–in the absence of any good reason to do otherwise. We shall not treat the Earth or meteorites. The magnitude of the terrestrial magnetic moment is 8×10^{15} T m^3. The terrestrial field undergoes secular variation and reversals. Thus, we assume there is nothing fundamental about the present-day polarity of planetary moments. We shall also assume that the present observed strengths of the planetary moments are typical and that none of them are undergoing reversals. As regards meteorites, we note that they contain important clues to the origin of the solar system, but the interpretation of these clues remains even more controversial than the interpretation of lunar magnetism (see Chapter 6). This review concentrates on our present observational knowledge of planetary magnetic fields, and only briefly discusses the complementary topic of the physical and chemical states of the interiors of these planets. Furthermore, we shall minimize discussion of the theory of the dynamo generation of planetary fields, which has been covered in Chapters 1–3. Those interested in a single paper treating all these topics are referred to the excellent review by Stevenson (1983).

2 MERCURY

Mercury is the smallest of the terrestrial planets ($R_m = 2440$ km), intermediate between the Earth's moon and Mars in size. It rotates more slowly than the Moon, rotating with a period of 59 days compared with the Moon's 28-day period. It is heavily cratered like the Moon and apparently has a very ancient crust. Mercury differs from the lunar case in that its rotational period of 59 days is not the same as its orbital period, and hence Mercury does not always keep the same face to the sun. It is much denser than the Moon, 5.4 g cm^{-3}, compared with the lunar 3.3 g cm^{-3}. It travels in an elliptical orbit of moderately high eccentricity ($e = 0.206$) and inclination to the ecliptic ($i = 7°$). Furthermore, Mercury appears to have an active dynamo in its core. The Moon does not. Recently, sodium in Mercury's atmosphere has been observed spectroscopically from Earth (Potter and Morgan, 1985). Although there have been reports of weak lunar outgassing events from surface instrumentation (*c.* 3 ton yr^{-1}) (Hodges, 1977), there have been no terrestrial spectroscopic observations of a lunar atmosphere despite the much greater proximity of the Moon and the greater opportunity for observation. Since sodium is not observed to be sputtered from the lunar surface in significant amounts, it seems unlikely that sputtering could release a significant sodium atmosphere at Mercury. Although the solar-wind flux is greater at Mercury because of its proximity to the Sun, its surface is less exposed to the solar wind owing to shielding by its magnetic field. Thus it seems very likely that the sodium at Mercury is due to outgassing from the interior of Mercury and is another indication that Mercury is an active, dynamic planet despite its outward similarity to the Moon.

2.1 Observations

Mercury has been visited by only one spacecraft, Mariner 10, which made three passes by the planet from March 1974 to March 1975, of which only the first and the third passes were suitable for studying planetary magnetism. Figure 1 shows the magnetic-field measurements obtained during the first encounter, commonly called Mercury I (Ness *et al.*, 1975a, b). The spacecraft crossed the dark side of the planet, approaching within 723 km of the surface on this encounter. The maximum field strength observed was nearly 100 nT at closest approach. The characteristics of the data were very similar to those that would be obtained on a pass through the terrestrial magnetosphere, but were on a far smaller scale here. As at the Earth, there was a bow shock behind which the interplanetary magnetic field was compressed while the solar wind flowing out from the Sun was suddenly slowed down and heated as it supersonically reached the planet. Behind the shock another boundary was crossed—the magnetopause. This boundary defines the region that is shielded by the planetary magnetic field from the solar wind.

After the point of closest approach (labelled CA) the regularity observed inbound disappeared. Rather than interpret this as an East–

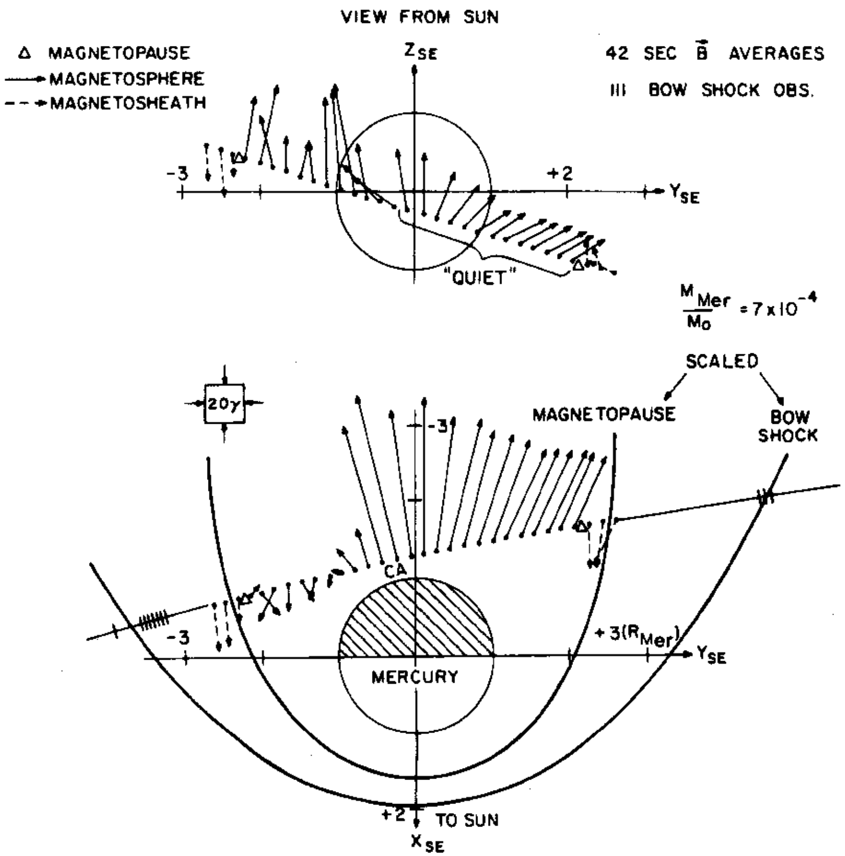

Figure 1. Projections of magnetic field observed along the trajectory of Mariner 10 as it passes behind Mercury on 29 March 1974 (Ness *et al.*, 1975a).

West asymmetry in the planetary magnetosphere, Siscoe *et al.* (1975) considered it to be the result of a temporal change in the magnetosphere similar to a terrestrial substorm. Whatever the cause, the irregular field was accompanied by the appearance of energetic particles, and the data in this region were not suitable for assisting in the determination of the planetary moment.

No such magnetospheric dynamics were observed during the third Mercury encounter, Mercury III, the data for which are shown in Fig. 2. This pass was closer to the planet and at higher latitudes, and thus a magnetic field at closest approach of over 400 nT was measured. Otherwise the features observed on this pass were quite similar to those

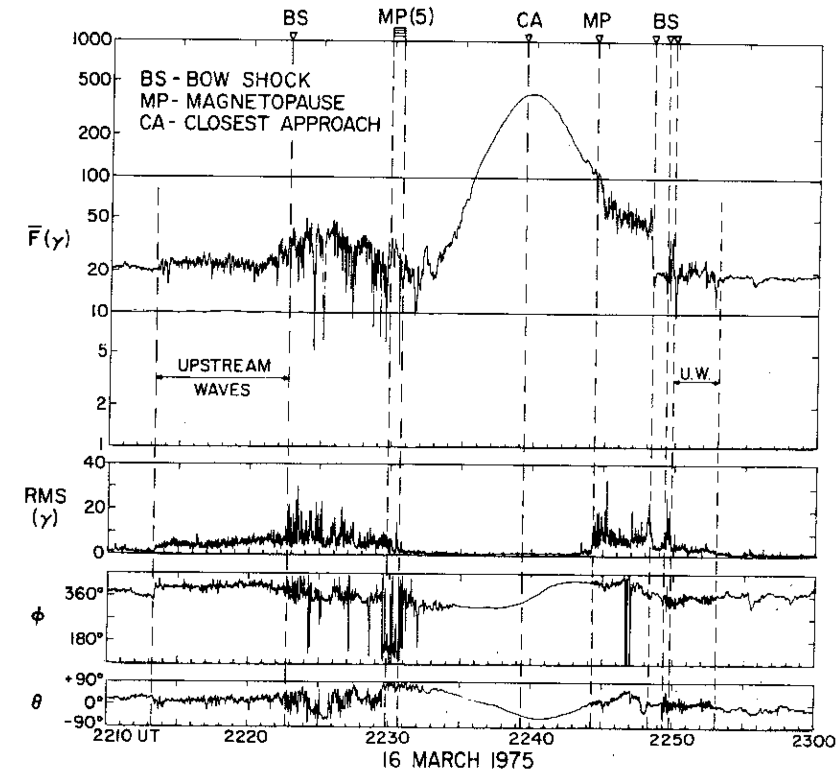

Figure 2. Magnetic field measured by the Mariner 10 spacecraft as it passed behind Mercury on 16 March 1975 (Ness *et al.*, 1976). The data points are 1.2 second averages in a Mercury-centred solar-ecliptic coordinate system. The angle ϕ is the longitude in the ecliptic plane measured toward the East limb of the Sun. θ is the latitude.

observed on Mercury I, and the bow shock and magnetopause crossing were close to their expected positions. The energetic particles did not show evdence for the occurrence of a substorm on this pass.

2.2 Deriving the planetary magnetic moment

The magnetic measurements of Mariner 10 were strongly influenced by the current system flowing on the magnetospheric boundaries. This influence can be seen in Fig. 3, which shows how much the field lines in

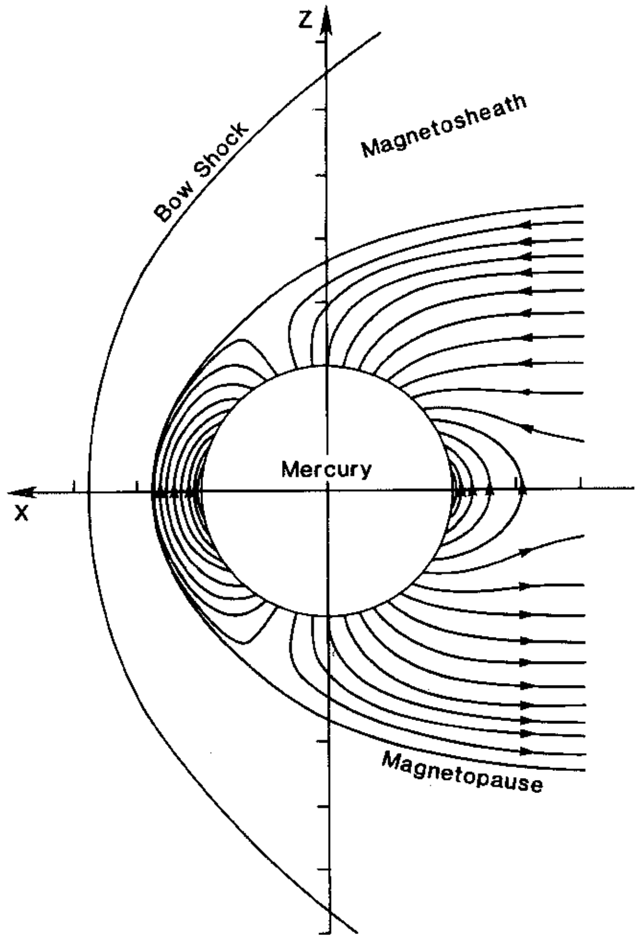

Figure 3. Magnetic-field lines in a model Mercury magnetosphere showing the magnetopause and bow shock to scale. (After Jackson and Beard, 1977.)

the night-side magnetosphere of Mercury differ from a dipolar configuration. External conditions must be taken into account when attempts are made to derive the planetary moment. No solar-wind ion data were obtained on Mariner 10. Solar wind properties were deduced from electron measurements, a less accurate technique (Ogilvie *et al.*, 1977). Furthermore, the two Mariner 10 passes through the Mercury magnetosphere surveyed only a limited portion of the planet. In order to

accurately survey the magnetic moments of the internal field, a low-altitude polar orbit is required.

As a result of this lack of data, the model inversions are nonunique and the derived moments controversial. A recent retrospective examination of the data concluded that even with modern inversion techniques these ambiguities cannot be removed (Connerney and Ness, 1987). Thus we shall adopt the approach of simply providing a historical account of the derivation of the magnetic moment. We shall also include several quite different methods of estimating the moment, because not all techniques have the same ambiguities.

In order to deduce the magnetic moment of a planet, one can use the position of the magnetospheric boundaries together with the solar-wind conditions to determine how large a moment is required to create that magnetospheric volume. This has the disadvantage of returning only an equivalent dipole field strength, but has the advantage of being a global integration of the moment. The location of the boundary of the magnetosphere is not simply determined by local pressure balance, but rather by a global pressure balance. An alternative method for deducing the moments is to invert the magnetic measurements along the trajectory. This technique has the advantage of enabling the computation of higher harmonics, but has the disadvantage of nonuniqueness. Unless the coverage of the surface is sufficiently dense, large moments can remain hidden. These moments would contribute little field themselves on the trajectory along which the measurements were taken or would cancel in combination with other hidden moments along the trajectory. Both of these above methods were used on the Mariner 10 data.

The first reported moment was derived for Mercury I by assuming a standoff distance of 1.6 Mercury radii (R_m) and by using a 'measured' solar-wind velocity of 600 km s^{-1} and a density of 17 cm^{-3} as deduced from the electron measurements (Ness *et al.*, 1974). The formula

$$M = 9.14 \times 10^{-4} R_{sp}^3 (\rho v^2)^{1/2} \quad (1)$$

was used, where R_{sp} is the subsolar radius of the magnetopause (in m), M is the magnetic moment (in T m^3), ρ is the solar-wind mass density (in kg m^{-3}) and v is the solar-wind velocity (in m s^{-1}). The moment thus derived was 5.5×10^{12} T m^3. Later, the solar-wind velocity and density were revised downward to 550 km s^{-1} and 14 cm^{-3} (Ogilvie *et al.*, 1977). Furthermore, these later authors gave a best-fit subsolar radius of $1.4R_m$. Substituting these revised numbers into (1) gaves a moment of 3.1×10^{12} T m^3.

We note that the formula used by Ness *et al.* (1974) tends to overestimate the size of the magnetic moment. If we take a magnetospheric shape factor appropriate for a gasdynamic interaction with a ratio of specific heats of $\frac{5}{3}$ and ignore ring current and tail effects (Schield, 1969) then we obtain (in SI units)

$$M = 6.1 \times 10^{-4} R_{sp}^3 (\rho v^2)^{1/2}, \quad (2)$$

which is one third less than the value used by Ness *et al.* (1974).

Ness *et al.* (1974) also attempted to fit a vacuum dipole model to the data. They obtained a moment of 3.3×10^{12} T m^3 offset $0.47R_m$ from the centre of the planet. In their next treatment of the same data, Ness *et al.* (1975a) added two external sources to their inversion, a uniform field and the first harmonic. The planetary moment thus derived was 5.1×10^{12} T m^3. Ness *et al.* did not give the size of their external terms, but the paper implies that these external sources were very large in this fit. They also stated that a uniform external field together with two internal terms gave a less stable solution. We note, however, that in the latest analyses by these authors only a uniform external field has been used.

When the Mercury III data became available, the two-external–one-internal term inversion was used to derive a moment of 4.8×10^{12} T m^3 (Ness *et al.*, 1975b). Then the data for Mercury I and III were reanalysed with a one-external (uniform) and one-internal term model. The result was a moment of 5×10^{12} T m^3 (Ness *et al.*, 1976). Here the authors state that the magnitude of the uniform external field during Mercury III was from 35 to 60 γ compared to the 400 γ field seen at closest approach.

A completely different approach to modelling was employed by Y. C. Whang and N. F. Ness (unpublished manuscript, 1975), using an image dipole and a two-dimensional current sheet to fit the Mercury I data. Here they used the preliminary solar-wind parameters given by Ogilvie *et al.* (1974) to derive a stagnation field strength of 160 γ. As a further constraint, they used the location of the observed magnetopause crossings. The moment so derived was the first of a series of 'low' moments for Mercury, 2.3×10^{12} T m^3.

Later, Whang (1977) added the data from Mercury III to this analysis and computed a quadrupole and an octupole moment. The quadrupole and octupole moments in this model are not general; rather, they are just the axial components, but they are useful in suggesting the possible magnitude of the internal higher-order moments. Again the moment is low, 2.4×10^{12} T m^3. The ratio of the dipole to quadrupole to octupole fields at the planetary surface is 1 : 0.45 : 0.29.

A different approach to the same problem was used by Jackson and Beard (1977) and Ng and Beard (1979), who fitted the data to a magnetospheric model field. They used $1.4R_m$ as the best average stagnation point for Mercury I and III instead of letting this be different for the two encounters. They then calculated the ratio of Mercury's compressed dipole field to that of the Earth for each encounter separately (and all together), getting a different dipole moment for each encounter rather than different external conditions. Nevertheless, the solutions obtained are instructive. For Mercury I, Jackson and Beard found that the dipole moment derived is not very sensitive to the addition of a quadrupole moment, probably because Mariner 10 did not come as close to the planet on the pass. However, for Mercury III, which came within 330 km of the surface, the derived dipole moment was very sensitive to the addition of a quadrupole moment. Overall, their best dipole moment (for the first half of Mercury I and all of Mercury III) was 2.5×10^{12} T m^3, the axial quadrupole moment being about two thirds of this value. We note that in solutions without an octupole moment, Whang (1977) obtained a similar quadrupole moment. Ng and Beard (1979) repeated the analysis of Jackson and Beard (1977), but used an offset dipole rather than a dipole plus quadrupole. They found a dipole moment essentially unchanged in magnitude, 2.75×10^{12} T m^3, offset slightly away from the Sun and toward dawn with a significant ($0.19R_m$) northward displacement.

These later estimates of the moment by Whang, Whang and Ness, Jackson and Beard, and Ng and Beard seem surprisingly low in comparison with the preliminary estimates. However, if we take the approach of simply fitting conic sections to the observed magnetopause and shock positions, we derive a stagnation radius of $1.35R_m$ (Russell, 1977, 1979a), which for normal solar-wind conditions at Mercury's aphelion (the location of Mercury I and III) would correspond to a moment of only 1.8×10^{12} T m^3.

Recently, Beard (personal communication, 1985) has advocated the use of the formula

$$\frac{R}{R_{sp}} = 1 + 0.0851\phi^2 + 0.0251\phi^4 \quad (3)$$

to extrapolate to the subsolar point, where ϕ is the angle of the observation from the subsolar point measured in radians. Doing this, we obtain subsolar radii for the magnetopause of only 1.23 and $1.22R_m$ respectively for the two encounters. Using (2) above and the revised solar-wind parameters of Ogilvie *et al.* (1977), this is equivalent to a moment of 1.3×10^{12} T m^3.

Slavin and Holzer (1979a,b) have criticized the above analyses because they did not take into account the tangential stress on Mercury's magnetopause, which they feel would be more important on Mercury than on Earth. Erosion of the terrestrial magnetosphere is caused by the inhibition of return flow to the magnetopause to replace flux eroded by merging with the interplanetary field (Coroniti and Kennel, 1973). Since Mercury has either no ionosphere or a very weak one, line tying is weak, and it is not clear whether erosion is as important at Mercury as at Earth. Slavin and Holzer (1979a,b) derived a dipole moment of $6 \pm 2 \times 10^{12}$ T m^3, presently the largest of the estimates of the Mercury moment.

Ness (1979) has criticized the quadrupole and octupole modelling studies as not improving the dipole-moment calculation. He attributes these apparent moments to spatial-harmonic aliasing. In other words, there is not sufficient planetary coverage to calculate higher moments. On the other hand, spatial-harmonic aliasing works both ways, and these studies at a minimum show how uncertain the dipole moment is. Furthermore, these studies derive dipole moments that are quite similar to those derived from the observed boundary crossings and the solar-wind pressure. This latter calculation is not affected by spatial aliasing. From the size of the core, we should expect sizable higher-order moments at Mercury (Elphic and Russell, 1978). Connerney and Ness (1987) have put some order into the confusing array of models of the Mercurian field by noting that because of the limited planetary coverage there is a basic uncertainty in the derived fields. This uncertainty takes the form of a functional relationship between the quadrupole and dipole terms such that there is little quadrupole moment when the dipole is large and less of a dipole component when the quadrupole is large. Along this line of theoretical uncertainty lie all the derived models. The resolution of this ambiguity cannot come from further inversions of the same data. Either new data or different constraints are needed.

We note that since the core of Mercury is a significant fraction of the planatary radius, the planetary magnetic field is stiffened against compression (Hood and Schubert, 1979; Suess and Goldstein, 1979). Accordingly, a solar-wind dynamic pressure of 25–150 times the normal solar-wind pressure would be necessary to push the magnetopause down to the planetary surface. Thus the percent of time that the solar wind directly impacted the surface of Mercury may be less than that calculated by Siscoe and Christopher (1975), even though they used a much larger planetary moment than the later authors.

2.3 The source of the intrinsic field

The magnetic field at Mercury is clearly of internal origin. It is too large to be caused by any external induction process (Herbert *et al.*, 1976). There are two possible sources of an intrinsic field: permanent or

remanent magnetization of the crust, and regenerative dynamo action in a liquid core. The effectiveness of these sources depends on the present thermal state of the planet, and in the case of remanence also on the thermal history of the planet. For the crust of Mercury to be magnetized, it must be below the Curie point of some carrier of magnetic remanence. For a global moment to have arisen, the crust must have cooled through the Curie point in some applied field. If the applied field were external to the planet, what was its source that remained so long and unidirectional as the crust cooled? If the source were an internal dipole, an overall dipole moment would not arise as the crust cooled unless if it were due to second-order effects such as crustal asymmetries and finite-cooling-time effects, since a dipolar magnetized shell produces no external magnetic field (Runcorn, 1975a,b). It is possible to produce a field as large as Mercury's present field by invoking these second-order effects (Stephenson, 1976), but the values of iron content, ancient magnetizing field and crustal thickness are all rather large. The iron content would have to be greater than allowed by thermal-evolution models (Stevenson *et al.*, 1983). If the internal magnetizing field reversed repeatedly as the crust was cooling and becoming magnetized, much of the remanent magnetism would be cancelled by layers of opposite magnetization, and the resulting planetary moment would be much less (Srnka and Mendenhall, 1979). Thus it appears quite certain that Mercury has an intrinsic field generated by currents in a fluid core.

The problem with the hypothesis of a Mercurian dynamo is that the field strength should be larger than observed if the dynamo is similar to that of the Earth. Stevenson (1984, 1986) has proposed two ways out of this. First, as stated in §1, the Mercurian dynamo may be energy-limited. There is simply not enough energy to power the dynamo to the level at which the Earth's dynamo works. Alternatively, there is no self-sustained dynamo at all. Rather, according to Stevenson (1986), vertical structure on the core–mantle interface of Mercury could lead to temperature differences along the core–mantle interface that could drive thermoelectric currents. The poloidal thermoelectric currents would generate toroidal magnetic fields. The fluid motions in the convective core would then, by the so-called α-effect, generate poloidal fields. In this model the observed field is generated by fluid motions, but the dynamo is not self-sustaining. Verification of such ideas for the Mercurian field must wait for many years, as no new missions to Mercury are scheduled.

3 VENUS

Of the terrestrial planets, Venus is the closest to the Earth, and in size is almost a twin to the latter ($R_v = 6051$ km). However, Venus rotates much