Planetary Magnetism

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Synopsis:

The chapter on Planetary Magnetism by Connerney describes the magnetic fields of the planets, from Mercury to Neptune, including the large satellites (Moon, Ganymede) that have or once had active dynamos. The chapter describes the spacecraft missions and observations that, along with select remote observations, form the basis of our knowledge of planetary magnetic fields. Connerney describes the methods of analysis used to characterize planetary magnetic fields, and the models used to represent the main field (due to dynamo action in the planet's interior) and/or remanent magnetic fields locked in the planet's crust, where appropriate. These observations provide valuable insights into dynamo generation of magnetic fields, the structure and composition of planetary interiors, and the evolution of planets.

Heading One: Introduction

Planetary magnetism began with the study of the geomagnetic field, and is one of the oldest scientific disciplines. Magnetism was both a curiosity and a practical tool for navigation, the subject of myth and superstition as much as scientific study. With the publication of *De Magnete* in 1600, William Gilbert demonstrated that the Earth's magnetic field was like that of a bar magnet. Gilbert's treatise on magnetism established the global nature and dipole geometry of the field and led to an association between magnetism and the interior of planet Earth. Observations of Earth's magnetic field gathered over time demonstrated that it varies slowly with time. The orientation of the field was observed to execute a long term, or secular drift, called polar wander. Later, when accurate measurements of the magnitude of the field could be made, secular variations in the field magnitude were identified. Given the importance of accurate magnetic field maps in navigation, global surveys of the magnetic field were conducted periodically, at first by ships of discovery and commerce and later by specialized magnetic survey ships built to minimize distortion of the magnetic field they sought to measure. In recent times, polar-orbiting, near-Earth satellites such as MAGSAT, Orsted, CHAMP, SAC-C, and SWARM provided the capability to map the vector magnetic field over the entire surface of Earth in a brief period, to capture a snapshot of the field at each epoch. By studying the magnetic properties of terrestrial rocks, paleomagnetists can extend the history of the geodynamo over time scales of hundreds of millions of years.

Planetary magnetism is also a recent study, born of the space age and a new era of discovery that began in the 1960's with the launch of spacecraft to other bodies. The sun and stars were known to possess magnetic fields, but one could only speculate on whether the planets and satellites in our solar system were magnetized. The planet Jupiter is an exception; its magnetic field was inferred from the detection of nonthermal radio emissions [*Burke and Franklin,* 1955] as early as 1955, before the era of planetary exploration. In the decades that followed, the gross characteristics of Jupiter's magnetic field were deduced from variations of the plane of polarization of the emissions or variations in flux density (see, e.g., the review by *Berge and Gulkis* [1976]). Estimates of the orientation of Jupiter's field (dipole tilt) were more successful than those of its magnitude; the detailed nature of Jupiter's field was established by direct measurement, by spacecraft.

In little more than three decades, spacecraft have visited nearly every large body in the solar system. The first in situ observations of the Jovian magnetosphere were obtained by the Pioneer 10 and 11 spacecraft in December 1973 and December 1974, respectively [*Smith et al.,* 1974, 1975a; *Acuña and Ness,* 1976]. The Pioneer 11 spacecraft continued on to make the first observations of Saturn's magnetic field as well [*Smith et al.,* 1980*a, b; Acuña and Ness,* 1980]. Voyagers 1 and 2 followed, passing through the Jovian magnetosphere in March and July of 1979, respectively [*Ness et at.,* 1979*a, b*]. Voyagers 1 and

2 continued onward to Saturn, obtaining additional observations of Saturn's magnetic field in November 1980 and August 1981 [*Ness et al.,* 1981, 1982]. Voyager 2 continued on its "grand tour" of the solar system, with the discovery of the magnetic field of Uranus in January 1986 [*Ness et al.,* 1986] and that of Neptune in August 1989 [*Ness et al.,* 1989]. Voyager 2's remarkable tour of the outer solar system was made possible by the celestial alignment of the planets, the syzygy, which occurs every 187 years or so.

Jupiter was visited again, in February 1992 by the Ulysses spacecraft en route to the high latitude heliosphere, and again, albeit at great distance, in 2004. Ulysses used a Jupiter gravity assist to escape the eclliptic plane, but in so doing passed to within ~6 R_j [*Balogh et al.*, 1992]. Additional observations were obtained by the Galileo Orbiter beginning in 1995 [*Johnson et al.,* 1992], but, apart from the de-orbit maneuver to send Galileo into Jupiter's atmosphere for planetary protection (of the Galilean satellites), little additional information regarding Jupiter's main field was obtained. Instead, Galileo surprised us with the remarkable discovery of an intrinsic magnetic field of Ganymede [*Kivelson et al*, 1996] and the induced magnetic fields of Europa and Calliso [*Khurana et al*, 1998; *Zimmer et al*, 2000]. Today, Cassini orbits Saturn gathering observations throughout the magnetosphere; observations from high-inclination orbits with periapses near 3 R_s are just now becoming available. Late in the extended ("Solstice") mission, one set of near-polar orbits will send Cassini through the ring plane near the F-ring (2.3 R_s) and another set ("proximal orbits") will send the spacecraft through the ring plane inside of the entire ring system, just above the cloudtops.

The Mariner 10 spacecraft discovered Mercury's magnetic field in brief flybys in 1974 and 1975 [*Ness et al.*, 1974a; *Ness et al.,* 1975, *Ness et al.*, 1976] and provided the first indication that slowly-rotating Venus was not magnetized [*Ness et al.*, 1974b]. As of this writing the MESSENGER spacecraft [*Solomon et al, 2001*; *Gold et al. 2001*] continues to map the Hermean magnetic field from its vantage point, in an elliptical orbit with periapsis at high northern latitudes. The European Space Agency's BepiColombo mission readies two spacecraft that will orbit Mercury in the near future [*Anselmi and Scoon*, 2001].

A great many spacecraft visited Mars [*Connerney et al*., 2005], beginning with the Soviet launch of Mars 1 in 1962, but Mars did not relinquish his secrets until the Mars Global Surveyor spacecraft entered orbit about Mars in September of 1997. Mars Global Surveyor measurements demonstrated that Mars does not have a global magnetic field of internal origin, but its crust is intensely magnetized [*Acuña et al*., 1998]. So Mars once had an internal magnetic field, for at least as long as it took the crust to acquire remanence. The Mars dynamo has long since disappeared, but the crust retains a magnetic imprint, acquired at least 4 billion years ago, that records at least part of the dynamo's history. The only traditional planet (setting aside the debate on the definition of "planet") not yet visited by spacecraft is the Pluto-Charon system, which is scheduled for a visit by the New Horizons spacecraft in July 2015. The New

Horizons spacecraft is not instrumented to measure magnetic fields, so no direct measurement is possible, but the presence of a magnetic field may be inferred from measurements made by other instruments.

Our knowledge of the magnetic fields of the planets is, in large part, based upon the planetary flybys that occurred decades ago. Mars is the most notable exception, having been mapped thoroughly over the course of more than nine years by the polar-orbiting MGS spacecraft, operating at a nominal altitude of 400 km [*Connerney et al., 2001*; *Connerney et al., 2005*]. At present, orbiters are operating at both Mercury and Saturn, dramatically extending observational coverage of those planets; elsewhere the observations are sparse in both space and time, particularly considered in light of the data available for the study of the geomagnetic field. Existing observations are, however, sufficient to describe a diverse group of planetary magnetic fields and provide clues to guide the continued development of dynamo theory. We are also likely to have a much more complete set of observations in the near future, with Cassini's Solstice Mission providing a large number of close-in, polar orbits of Saturn, and BepiColombo promising nearly complete global coverage of Mercury. At this time, the Juno spacecraft is en route to Jupiter, perhaps the best target in the solar system for studies of the dynamo. Beginning with orbit insertion in July, 2016, Juno will map Jupiter's magnetic field both near and far from a near-polar, elliptical orbit that ducks beneath that planet's intense radiation belts [*Bolton et al., 2010*].

Heading One: Tools

A planet with an intrinsic magnetic field stands as an obstacle to the solar wind, the high velocity stream of plasma emanating from the sun. The interaction of the planetary field and the solar wind forms a multi-tiered interaction region, often approximated as a set of conic sections. The supersonic solar wind forms a shock upstream of the obstacle (*bow shock*). The slowed solar wind flows around the obstacle within the *magnetosheath*, bounded by the bow shock and the *magnetopause*, often approximated by a paraboloid of revolution about the planet-sun line. Within the magnetopause is a region dominated by the planetary field, called the *magnetosphere*, a term introduced by Tom Gold in the post war era of rocketry that provided access to space in the early 1950's. The magnetospheric magnetic field extends well downstream in the anti-sunward direction, as if stretched like an archer's bowstring, to form the magnetotail. The observed magnetic field can be regarded as the sum of contributions from several sources, dominated by the planetary dynamo, sometimes referred to as the "main field". Other relatively minor sources include currents flowing on the magnetopause and tail currents arising from the solar wind interaction and distributed currents (*ring currents*) due to the coherent motion of charged particles within the magnetosphere. Field-aligned currents, called *Birkeland currents*, flow between the magnetosphere and the planet's electrically conducting ionosphere, particularly during magnetospheric storms, leading to

intense auroral displays. All of these currents produce magnetic fields, often relatively small when compared to the planetary field, particularly close to the planet's surface. However, if an accurate model is to be obtained from spatially limited or distant observations, these sources can be a significant source of errors if not modeled.

The planetary field is most often characterized by one of a very few simple models, depending on the application and/or the availability of observations. Simple representations include dipole models of one form or another and a potential field representation as a series of spherical harmonics. External fields, or those arising from currents outside the planet, are often accommodated by explicit models or spherical harmonic methods. The latter approach is useful only in regions free of currents.

Heading Two: The Offset Tilted Dipole

Dipole models have found use in the interpretation of magnetic field observations, dating to Gilbert's analysis of Earth's magnetic field. The dipole is the simplest approximation to a localized source. The vector magnetic field **B(r)** of a dipole at the origin is given by

$$\mathbf{B(r)} = \left[-\frac{\mathbf{m}}{r^3} + \frac{3(\mathbf{m \cdot r})\mathbf{r}}{r^5} \right]$$

where **m** is the magnetic moment in units of $B \text{-} r^3$ and $r = |\mathbf{r}|$. The field of a dipole located near the origin is computed by translation of the origin of the coordinate system. Offset tilted dipole (OTD) models are found using forward modeling techniques to compare in-situ observations to the OTD model field. The dipole offset (x_o, y_o, z_o) is not linearly related to the computed field, so an iterative procedure is often used to find the optimal dipole position [e.g., *Smith et al.,* 1976]. Alternatively one may perform a spherical harmonic analysis (see next section) of degree and order 1 centered upon each of a large number of candidate dipole positions, selecting the offset which minimizes the model residuals. In either case, the result is the OTD which most closely approximates the field measured along the spacecraft trajectory.

The simple OTD representation provides a convenient approximation to the field, particularly useful at larger distances. It is useful in visualizing the gross characteristics of the field and it is often used to chart the motion of charged particles throughout the magnetosphere. However, the simplicity of the OTD is also a limitation, in that complex field geometries cannot be adequately represented with such a limited parameterization. In general, a dipole representation can be expected to closely approximate the field for radial distances $r \gg a$, where a is a characteristic source dimension. At lesser radial distances, an

accurate description of the field requires more flexibility in parameterization, e.g., consideration of higher degree and order moments.

A logical extension of the simple OTD model is one of many dipoles. Earth's field has been approximated by adding one or more smaller dipoles to a "main dipole" and treating the field as a summation over distinct sources; this provides a simple means of introducing an anomaly, or departure from dipole geometry. The Earth's field has been approximated by an ensemble of many dipoles located on the surface of a sphere of radius 0.5 Re, the fluid core radius. In addition to the field generated deep within Earth, induced and remanent magnetization in the crust can be modeled with another set of dipoles, scattered about a sphere of radius 1.0 R_e to approximate sources in the crust [*Mayhew and Estes*, 1983]. A similar approach has been taken to model the crustal magnetic field of Mars [*Purucker et al.,* 2000].

Heading Two: Spherical Harmonic Models

In the absence of local currents ($\nabla \times \mathbf{B} = 0$)**,** the magnetic field may be expressed as the gradient of a scalar potential V (**B** = -∇V). It is particularly advantageous to expand the potential V in a series of functions, called spherical harmonics, which are solutions to Laplace's equation in spherical coordinates. This approach, introduced by Gauss in 1839, has been very popular in studies of the geomagnetic field, and subsequently those of the planets. The traditional spherical harmonic expansion of V is given by [e.g., *Chapman and Bartels,* 1940; *Langel,* 1987]

$$V = a\sum_{n=1}^{\infty}\left\{\left(\frac{r}{a}\right)^{n}T_{n}^{e}+\left(\frac{a}{r}\right)^{n+1}T_{n}^{i}\right\}$$

where a is the planet's equatorial radius. The first series in increasing powers of **r** represents contributions due to external sources, with

$$T_{n}^{e} = \sum_{m=0}^{n}\left\{P_{n}^{m}\left(\cos\theta\right)\left[G_{n}^{m}\cos\left(m\phi\right)+H_{n}^{m}\sin\left(m\phi\right)\right]\right\}$$

The second series in inverse powers of **r** represents contributions due to the planetary field or internal sources, with

$$T_n^i = \sum_{m=0}^{n} \left\{ P_n^m \left(\cos\theta \right) \left[g_n^m \cos\left(m\phi \right) + h_n^m \sin\left(m\phi \right) \right] \right\}$$

The P_n^{m} (cos θ) are Schmidt quasi-normalized associated Legendre functions of degree n and order m, and the g_n^{m}, h_n^{m} and G_n^{m}, H_n^{m}, are the internal and external Schmidt coefficients, respectively. These are most often presented in units of Gauss or nanoteslas ($1G = 10^5$ nT) for a particular choice of equatorial radius a of the planet. Different values of the equatorial radius have been used and should be noted in the comparison of various field models. The angles θ and ϕ are the polar angles of a spherical coordinate system, θ (co-latitude) measured from the axis of rotation and ϕ increasing in the direction of rotation. The three components of the magnetic field (internal field only) are obtained from the expression for V above:

$$B_r = -\frac{\partial V}{\partial r} = \sum_{n=1}^{\infty} \sum_{m=0}^{n} \left\{ (n+1) \left(\frac{a}{r} \right)^{n+2} \left[g_n^m \cos\left(m\phi \right) + h_n^m \sin\left(m\phi \right) \right] P_n^m \left(\cos\theta \right) \right\}$$

$$B_\theta = -\frac{\partial V}{r\partial\theta} = -\sum_{n=1}^{\infty} \sum_{m=0}^{n} \left\{ \left(\frac{a}{r} \right)^{n+2} \left[g_n^m \cos\left(m\phi \right) + h_n^m \sin\left(m\phi \right) \right] \frac{dP_n^m \left(\cos\theta \right)}{d\theta} \right\}$$

$$B_\phi = \frac{-1}{r\sin\theta} \frac{\partial V}{\partial\phi} = \frac{1}{\sin\theta} \sum_{n=1}^{\infty} \sum_{m=0}^{n} \left\{ m \left(\frac{a}{r} \right)^{n+2} \left[g_n^m \sin\left(m\phi \right) - h_n^m \cos\left(m\phi \right) \right] P_n^m \left(\cos\theta \right) \right\}$$

The leading terms in the series (through degree 4) may be computed using the Schmidt quasi-normalized Legendre functions listed in Table 1.

<Table 1 near here>

Observations are often rendered in a west longitude system, in which the longitude of a stationary observer increases in time as the planet rotates. West longitudes are simply related to the angle ϕ by λ = 360 - ϕ. Longitudes for the terrestrial planets may be assigned by reference to surface features. For the outer planets, the rotation period is inferred from observations of episodic radio emissions, which, by assumption, are locked in phase with the rotation of the planet's magnetic field and hence the electrically conducting deep interior of the planet. For these planets, longitudes must be assigned with knowledge of

the rotation rate and time. For Jupiter, observation of high frequency radio emission (penetrating Earth's ionosphere) over several decades led to an accurate determination of the planet's rotation period, although one must be cognizant of the occasional update (e.g., $\lambda_{III}(1957)$ vs $\lambda_{III}(1965)$). These two systems result in significantly different longitudes when observations of the Pioneer era are compared with those of the Voyager era, so one must be careful to transform earlier results to a current coordinate system. A detailed description of Jovian coordinate systems is given by *Dessler* [1983]. For Saturn, Uranus, and Neptune, the rotation periods and longitude system in current use are derived from Voyager observations of radio emissions [*Desch and Kaiser,* 1981; *Desch et al.,* 1986; *Warwick et al.,* 1989]. In practice, the use of a radio-derived rotation period satisfies the need for a planet-fixed coordinate system for fluid planets but exceptions (e.g., Saturn) do exist.

The expansion in increasing powers of *r,* representing external fields, is often truncated at $n = N_{max} = 1$ corresponding to a uniform external field due to distant magnetopause and tail currents, i.e., sources well beyond the region of interest. No potential field can represent the field of local currents, so in practice the external expansion is often truncated at $N_{max} = 1$ and local currents are modeled with the aid of explicit models. The dominant contribution from local currents may be described as a large-scale equatorial current system [*Connerney et al.,* 1981]. Field-aligned, or Birkeland currents, may also play a role in addition to induction effects, but thus far have not been included in such analyses, excepting those of Earth [e.g., *Backus,* 1986; *Sabaka et al.*, 2002]. It is very important to adequately treat the fields of local currents, because even small errors accumulating along the spacecraft trajectory can lead to rather large errors in the spherical harmonic coefficients deduced from observations. Fields produced by external current systems vary over relatively short timescales (e.g., in response to variations in the solar wind ram pressure on the magnetopause), further complicating modeling efforts.

The maximum degree and order required of the internal field expansion depends on the complexity of the field measured. In usual practice, the series above is truncated at a maximum degree N_{max}, where N_{max} is large enough to follow variations in the field at the orbital altitude of the measurement. The number of free parameters grows rapidly with increasing N_{max}, as $n_p = (N_{max} + 1)^2 - 1$. If the observations are well distributed on a sphere, the spherical harmonics do not covary, and the coefficients obtained are independent of the truncation of the series. However, if the observations are poorly distributed or sparse, as is often the case for a planetary flyby, the usual spherical harmonics covary, so it is advisable to construct new orthogonal basis functions. Furthermore, if an arbitrarily small choice of N_{max} is imposed upon the model in order to obtain a manageable linear system, large errors in low degree and order terms will result from the neglect of higher-order terms. To address these problems, Connerney [1981] introduced a method of analysis based on the singular value decomposition of Lanczos. The method [*Lanczos*, 1971] involves the construction of partial solutions to the generalized linear inverse problem to

obtain estimates of those parameters which are constrained by the observations. The model parameters which are not constrained by the data are readily identified and exploited in characterization of model nonuniqueness. One advantage of this approach is that the physical model of the planetary field (expansion to degree N_{max}) does not depend on the completeness or extent of the available observations. Partial solutions, and estimates of the model parameters that result, may be interpreted within the context of the "resolution matrix" for the particular solution [see *Connerney,* 1981; *Connerney et al.,* 1991]. The close flyby of Neptune in 1989 serves to illustrate this concept well.

Heading One: Terrestrial Planets

Heading Two: Mercury

Mercury, the smallest (R_m = 2439 km) and least massive (M_m = 0.056 M_e) planet in the solar system, occupies the pole position in its race about the sun, with a semimajor axis of 0.39 AU. Due to its large orbital eccentricity (0.206), in one orbit about the sun (88 days) its orbital radius varies from 0.31 to 0.47 AU. It is locked in a spin-orbit resonance (*Pettengil and Dyce*, 1965; *Goldreich and Peale*, 1966) rotating once every 58.65 days or 3 times every 2 orbital periods. It is difficult to observe from Earth, due to its proximity to the sun, and little was known about Mercury prior to observation by spacecraft on 29 March 1974, the first of three encounters with the Mariner 10 spacecraft. Discovery of the intrinsic magnetic field of Mercury was a bit surprising since it was widely assumed that a body as small as Mercury would have completely solidified, ruling out generation of a magnetic field by dynamo action. The Mariner 10 observations of Mercury provided the only in-situ observations obtained in the first four decades of space exploration. As this chapter is being written, the MESSENGER spacecraft [*Solomon et al*., 2001; *Gold et al.,* 2001] is orbiting Mercury, in part to better understand the internal magnetic field and its origin. MESSENGER (MErcury Surface, Space ENvironment, GEochemestry, and Ranging) entered orbit in March 2011, after a series of flybys of Earth (August, 2005), Venus (October, 2006 and June, 2007), and Mercury (three flybys: January and October, 2008, and September, 2009). A more ambitious mission to Mercury, "BepiColombo", is in development by the European Space Agency (ESA) and the Institute of Space and Astronautical Science in Japan (ISAS) for launch in the not to distant future [*Anselmi and Scoon*, 2001].

Heading Three: Observations

Mariner 10, known pre-launch as the Mariner Venus-Mercury spacecraft, was launched on 3

November 1973. The spacecraft used a close flyby of Venus on 5 February 1974 (for gravity assist) to achieve three close encounters with Mercury (29 March 1974, 21 September 1974, 16 March 1975) before it exhausted its expendibles. The orbital period of Mariner 10 after the Venus gravity assist was 176 days, very nearly commensurate with Mercury's orbital (and rotation) period, providing three very similar close encounters. Each time the spacecraft approached the planet from a local time of approximately 1900 hr and departed near 0700 hr. Given the brevity of these flyby encounters, the relatively slow planetary rotation rate, and the orbital commensurabilities, each encounter approximates a shot past a stationary object, the major difference being the altitude and subspacecraft latitude of the point of closest approach. The first and third of these encounters (M10-I and M10-III) provided observations within the Hermean magnetosphere (Figure 1); the second was targeted well above the sunlit hemisphere and beyond the reach of the Hermean magnetosphere in the upstream direction of the solar wind.

< Figure 1 near here >

Mariner 10's first flyby was relatively close, as planetary flybys go, with a periapsis of 1.29 R_m over the darkened hemisphere (anti-sunward direction), passing from dusk to dawn behind the planet as viewed from the sun. Mariner 10 traveled from bow shock (inbound) to bow shock (outbound) in a mere 33 minutes; 17 of which were spent within the confines of a Hermean magnetosphere (Figure 2). These observations are best understood in terms of the field external to the planet, even at such a close distance. The field increased in magnitude from about 45 nT just inside the magnetopause to approximately 100 nT, all the while maintaining the anti-sunward orientation characteristic of a planetary magnetotail. As Mariner 10 crossed the equator plane, the field decreased in magnitude precipitously, as if crossing a current sheet, after which it was much smaller in magnitude and chaotic, lacking the smooth variation and consistent orientation that characterized the inbound pass. These observations indicated that Mariner 10 passed through a small magnetosphere, sized just large enough to contain the solid body of the planet under nominal solar wind conditions [*Ness et al*, 1975]. To form such a magnetosphere, Mercury must have an internal magnetic field, albeit one with a modest dipole magnitude.

< Figure 2 near here >

Mariner 10's second flyby was upstream of Mercury in the solar wind, and too distant to penetrate the small Hermean magnetosphere. However, the third flyby was targeted to pass closely over the northern pole (Figure 3), in an effort to gather measurements optimized for characterization of the Hermean field. During the spacecraft's brief traversal of the magnetosphere (14 minutes) the field increased from about 20 nT to a maximum of 402 nT observed just after closest approach at 1.14 R_m radial distance (1 R_m = 2439 km). The approximate $1/r^3$ dependence of the observed field on spacecraft radial distance establishes the internal origin of the field, and initial analysis [*Ness et al.*, 1976] suggested a dipole moment of

approximately 342 nT-R$_m^3$, aligned with the -Z axis (normal to the Hermean orbital plane).

< Figure 3 near here >

The three MESSENGER flybys (M1, M2, M3) prior to orbit insertion were quite similar to the Mariner 10 first flyby (M10-I), in that all three passed through Mercury's shadow at low latitude, sampling the internal field near the innermost extent of the cross-tail current sheet [*Anderson et al.*, 2010]. The maximum total field measured during the M1 and M2 passes was ~160 nT, a magnitude that is comparable to the ~100 nT of external field produced by tail and magnetopause currents that result from the solar wind interaction [*Korth et al.*, 2004]. The third MESSENGER flyby was interrupted by a spacecraft safe mode entry near closest approach and instruments were 'safed', or powered off, for a brief period of time before nominal operations were restored.

MESSENGER entered orbit about Mercury on March 17, 2011, and became the first spacecraft to orbit the innermost terrestrial planet of our solar system. MESSENGER initially entered a highly elliptical, near polar (82º – 84º orbit inclination) orbit with a period of about 12 hours (Figure 4); these orbits sample all local times as Mercury orbits the Sun, with closest approach altitudes ranging from 200 to 500 km at high (60º – 74º) northern latitudes. After an Earth year in this orbit, the extended mission began with a pair of maneuvers (April 16 & 20, 2012) that reduced the orbital period to 8 hours. As of this writing, MESSENGER has completed over 2,600 such orbits.

< Figure 4 near here >

Heading Three: Models

Two fundamental issues arise in modeling the Hermean magnetic field. The first is a simple consequence of the relative weakness of the field, with a dipole moment only 0.05% that of Earth's; the second is a more subtle consequence of the distribution of observations collected by spacecraft in the immediate vicinity of the planet. With such a weak internal field, the planet occupies much of its magnetosphere, and the contribution of magnetospheric currents – primarily magnetopause currents and cross-tail currents – is relatively large, time variable, and difficult to model with great accuracy. Thus models of Mercury's magnetic field fall largely into two groups, distinguished by their treatment of external fields. The first group of models used a spherical harmonic expansion for both internal and external fields and were limited to the dipole (n=1) terms for internal fields. Since these are potential field models, they are restricted to the region in which no currents flow, so they may be applied for example within only one lobe of the tail. Initial analyses used the first inbound Mariner 10 observations through to the tail current sheet crossing, while later analyses benefited from the Mariner III observations obtained crossing the pole. The latter are better suited to internal field estimation by virtue of improved signal to noise (magnitude relative to external fields). These are listed in Table 2.

< Table 2 near here >

Another set of models were obtained using a spherical harmonic expansion for the internal field, augmented by explicit models of the external field. These included more detailed models of the fields due to magnetopause and tail currents, the latter being of great importance in interpretation of the first Mariner 10 encounter (M10-I) and the MESSENGER M1 and M2 close flybys. An explicit model of the external field allows one to extend analyses through regions of local currents (e.g., tail current sheet), accommodating non-potential fields more readily. A number of these models [*Whang*, 1977; *Jackson and Beard*, 1977; *Ng and Beard*, 1979; *Bergan and Engle*, 1981] were applied to M10 observations and allowed for the possibility of internal field through degree 2 or 3, but only in the zonal harmonic (m=0) coefficient, e.g., g_2^0 and g_3^0. Models allowing for axisymmetric quadrupole (n=2) or octupole (n=3) terms are characterized by a reduced dipole magnitude, relative to models allowing dipole terms only. This applies to all such models, regardless of author or method of analysis or even the details of the explicit models chosen to represent fields of external origin. This curious effect was shown to be related to model non-uniqueness, a consequence of the limited spatial extent of the data [*Connerney and Ness*, 1988]. With spatially limited observations, the internal field parameters and the external field parameters covary, in a predictable way, and can not be uniquely determined, lacking additional information. Thus following the Mariner 10 encounters, there was no evidence either for or against a non-zero axisymmetric quadrupole or octupole. Either model – an offset dipole or a simple dipole [e.g., *Ness et al,* 1979] provides an acceptable, if not entirely satisfactory, description of Mercury's internal magnetic field. This model non-uniqueness could only be resolved with additional close-in observations, preferably distributed more uniformly about the planet, and in particular, above the southern pole. The internal field is just sufficient to keep the dayside magnetopause above the planet's surface under nominal solar wind conditions (Figure 5).

< Figure 5 near here >

Initial analyses of the MESSENGER observations used the M1 and M2 flybys, in combination with Mariner 10 observations, and a variety of explicit models of the external field, to obtain qualitatively similar results. Anderson et al. [2008, 2010] and Uno et al. [2009] found a number of dipole representations with moments ranging from $240 - 270$ nT-R_m^3 and dipole tilts of $5° - 12°$ from the rotation axis, as well as (northward) offset dipole models that fit equally well with reduced dipole moment and negligible tilt. Alexeev et al. [2010] modeled the Mariner 10 and MESSENGER flybys, fitting an elaborate external field model to each, individually, concluding that Mercury's internal field is best represented by an axial dipole, with a moment of 196 nT-R_m^3, offset from the planet's center by 405 km (0.17 R_m) to the north. This result is rather similar to the Mariner 10 models [*Ng and Beard*, 1979; *Bergan and Engle*, 1981] characterized by an aligned dipole (with a moment of $180 - 207$ nT-R_m^3) offset

by 0.17 – 0.19 R_m northward along the z axis.

Heading Three: Discussion

MESSENGER's orbital phase has thus far provided a wealth of additional observations, well distributed in longitude and local time, but close-in observations are still limited to high northern latitudes. Thus the non-uniqueness inherent in the limited spatial distribution of observations – the covariability of the main dipole (g_1^0) term, northward offset or g_2^0 term, and the external field – described by Connerney and Ness [1988] and Korth et al. [2004] remains extremely challenging. Only by constraining, independently, the northward offset of the dipole using relatively distant observations of the location of the magnetic equator [*Anderson et al.,* 2011; 2012; *Johnson et al,* 2012] has it been possible to deduce the magnitude of the dipole (~190 nT/r_m^3, corresponding to a northward dipole offset of ~480 km) from a few Mercury years of MESSENGER orbital observations. This model internal field, along with a prescribed paraboloid magnetospheric model [*Alexeev et al.* 2010], fits selected MESSENGER orbits very well [*Johnson et al.,* 2012]. However, the uniqueness of this particular solution has yet to be demonstrated. The (averaged) magnetic equator locations critical to constraining the dipole offset are available at essentially two radial distances (~1.4 and ~2.5 R_m) determined by MESSENGER's fixed orbit; the innermost magnetic equator observations are available at most local times but in close proximity to the magnetopause. The more distant observations are necessarily confined to a narrow range of local times (near midnight). The magnetotail current sheet geometry may be more complex than assumed, near the magnetopause, and somewhat warped across and down the tail, as is observed at Earth [*Fairfield,* 1980; *Gosling et al.,* 1986; *Zhang et al.* 2006]. Given that the external field contributes up to 40% or 50% of the field measured [*Anderson et al.,* 2010; *Johnson et al.,* 2012] one must be very confident in knowledge of the external field in order to narrow the range of uncertainty in the internal field parameters (especially the *zonal* harmonics). The $g_1^0 – g_2^0$ covariability may only really be resolved satisfactorily after BepiColombo's well distributed orbital measurements (Figure 6) are available [*Scuffham et al.,* 2006]. There remains the intriguing possibility of detecting, and characterizing, a putative crustal magnetic anomaly with the many close passes afforded by either MESSENGER or BepiColombo, but as yet no unambiguous evidence for such has emerged [*Purucker et al,* 2009].

< Figure 6 near here >

Heading Two: Venus

Venus is similar to Earth in size (R_v = 6051 km), mass (M_v = 0.815 M_e), and density, and probably composition as well. Venus orbits the sun once every 225 days at a mean orbital radius of 0.723 AU. Proximity to Earth has invited a relatively large number of missions, particularly in the early decades of space exploration [*Colin,* 1983; *Russell and Vaisberg,* 1983]. However, unlike Earth, Venus has no

measureable internal magnetic field; spacecraft have detected only weak magnetic fields associated with the solar wind interaction [*Russell and Vaisberg*, 1983]. Venus also rotates very slowly, with a period of 243 days. The relatively slow rotation in itself does not explain the lack of a dynamo, as it is easily enough to make Coriolis forces dynamically important [*Stevenson*, 2003].

Heading Three: Observations

The first measurements of the magnetic field near Venus were obtained in December, 1962, by the Mariner 2 spacecraft, on a flyby trajectory with a closest approach of 6.6 planet radii. Magnetic field measurements obtained by the triaxial fluxgate magnetometer aboard Mariner 2 provided no evidence of an intrinsic planetary field [*Smith et al*, 1963; *Smith et al.*, 1965a] at this distance. Venus could not have an internal magnetic field with a moment exceeding 0.05 that of the Earth.

The 1960's and 1970's saw the arrival of the Venera 4 (1967) and Venera 6 (1969) entry probes and the Mariner 5 (1967) and Mariner 10 (1974) spacecraft on flyby trajectories with close approach distances of 1.7 and 1.96 R_v. These were followed by the Venera 9 and 10 orbiters in 1975 and the Pioneer Venus Orbiter (PVO) in 1978. The Veneras were placed in highly elliptical orbits with an orbital period of approximately 48 hours, periapsis at 1545 and 1665 km altitude, respectively, and apoapses of approximately 18.5 R_v. Throughout this period the existence of an intrinsic magnetic field was debated [see *Russell and Vaisberg*, 1983] amid ever decreasing estimates, or upper limits on, the Venus dipole moment. The PVO was placed in a highly elliptical orbit with an orbital period of 24 hours, periapsis at 140-170 km altitude, and apoapsis at approximately 12 R_v. This orbit geometry afforded observation of, and a better understanding of, the interaction of the solar wind with a planetary atmosphere. These observations led to an upper limit on the Venus dipole moment of 4 x 10^{-5} that of Earth's [*Russell et al.*, 1980], which was reduced further (to ~1 x 10^{-5} M_e) by a more exhaustive analysis [*Phillips and Russell*, 1987] that also demonstrated a lack of evidence for magnetic fields associated with crustal remanence. The Venus Express spacecraft, in a highly elliptical polar orbit about Venus since April, 2006, will not likely alter conclusions regarding the internal magnetic field of Venus [*Zhang et al., 2006*].

Heading Three: Discussion

One concludes that Venus does not currently have an active dynamo; at present, little can be said of the possibility that Venus once sustained a dynamo. One would like to understand why this earth-like planet does not have an active dynamo, and there are many possibilities [*Stevenson*, 2003; *Spohn*, 1991]. The favored explanation is that sufficient convection in the fluid outer core is lacking at present, owing perhaps to the inefficiency of heat transport (relative to Earth, where plate tectonics helps cool the core). More specifically, the lack of a present day dynamo may be the result of a cessation of plate tectonics on Venus [*Nimmo, 2002*], as has been proposed for Mars [*Nimmo and Stevenson,*2000].

Heading Two: Mars

Mars is about half the size (R_m = 3394 km) of Earth and much less massive (M_m = 0.108 M_e). Mars resides at a mean orbital distance of 1.52 AU, orbiting the sun once in 687 days. It has also been an attractive target for planetary exploration, playing host to more than 30 attempted missions since 1960 [*Connerney et al*, 2004]. At present, one of the indefagitable Mars Exploration Rovers (MERs) continues to explore the surface after more than ten years of service [*Crisp et al*, 2003]. The Mars Global Surveyor spacecraft [*Albee et al*, 1998] fell silent in 2006 after more than nine years in orbit, and Mars Odyssey [*Saunders et al*, 2004], Mars Express [*Chicarro et al.,* 2004] and the Mars Reconnaissance Orbiter [*Graf et al.,* 2005] all continue to acquire observations from orbit. The Mars Science Laboratory (MSL) was delivered to the surface in August, 2012 to begin its search for environmental conditions suitable for past or present microbial life, and the MAVEN spacecraft is en route to Mars (arrival September 21, 2014) to begin a year-long study of the Mars atmosphere. It is the most impressive armada of spacecraft ever assembled for planetary exploration, although only one currently active vehicle – the MAVEN spacecraft - is instrumented to measure magnetic fields.

Heading Three: Observations

After five unsuccessful attempts (four by the USSR; one by the USA), Mariner 4 became the first mission to return data from Mars on July 15, 1965. Mariner 4's flyby trajectory brought the spacecraft to within 4 R_m of the planet, from which vantage point it observed the near-Mars environment. The magnetic field was indistinguishable from the interplanetary environment; no bow shock, no magnetopause, no magnetosphere was observed. If Mars had an intrinsic magnetic field, it had to be very small in magnitude, less than 3 x 10^{-4} M_e (M_e, the magnetic dipole moment of Earth, is 8 x 10^{25} G cm^3), to remain unseen [*Smith et al*, 1965b]. This upper limit is equivalent to an equatorial surface field of 100 nT, well less than Earth's (30,000 nT) but not so much less than the as yet undiscovered magnetic field of Mercury (330 nT).

In the following three decades, numerous spacecraft visited Mars, but no spacecraft, probe, or lander instrumented to measure magnetic fields would pass close enough to the surface to establish the presence of an intrinsic magnetic field. The US missions following Mariner 4, Mariner 9 (1971) and Viking 1 and 2 (1975), were not instrumented to measure magnetic fields. Additional magnetic field observations were obtained by the Soviet Mars 2,3 and 5 spacecraft [*Russell*, 1978a; 1978b; 1979; *Ness*, 1979] in the early 1970's and the Soviet Phobos 2 mission in 1988. These observations fueled a great deal of debate regarding the existence of a putative Mars dipole (with equatorial surface field in the range 20-65 nT) but no convincing evidence of an intrinsic field was found [*Riedler et al*, 1989].

Thus it was the Mars Global Surveyor (MGS) spacecraft [*Albee et al*, 1998], launched in 1996 as a

(partial) replacement for the unsuccessful Mars Observer, which provided the first unambiguous detection of the intrinsic magnetic field of Mars. On September 11, 1997, the Mars Global Surveyor spacecraft was inserted into a highly elliptical polar orbit from which it would transition to a circular polar mapping orbit (nominal 400 km altitude) via a series of aerobraking maneuvers [*Albee et al*, 1998]. Aerobraking uses atmospheric drag to slow the spacecraft, gradually reducing apoapsis by repeated passage through the atmosphere; trim maneuvers are used to raise or lower the altitude at closest approach (periapsis) to obtain the desired drag. Each aerobraking pass brought the spacecraft over the surface of Mars, to altitudes < 100km, about as close to the surface as any orbiting platform can achieve. During each aerobraking pass, the MGS magnetometer/electron reflectometer (MAG-ER) investigation (*Acuna et al*, 1992) acquired measurements of the vector magnetic field along the orbit track at varying altitude above the surface.

Periapsis pass 6 (day 264, 1997) provided a wealth of scientific insight in a few minutes time: lacking a significant field of global scale, Mars has no active dynamo, but must have had one in the past, when the crust acquired intense remanent magnetization [*Acuna et al*, 1998]. These observations are illustrated in Figure 7, showing the spacecraft trajectory and vector magnetic field measured at 3 second intervals along the trajectory. From this image one can see that the intrinsic field is not global in scale, as it would necessarily be if it were generated deep within the planet by dynamo action. Instead, the field is localized to a small part of the trajectory, with a characteristic length scale comparable to the altitude of observation. This is the signature of a strong and localized magnetic source in the crust of Mars, just beneath the spacecraft orbit track. This particular source produces a field of nearly 400 nT at ~110 km altitude, requiring a very large volume of intensely magnetized material, with a moment of ~1.6 x 10^{16} A-m^2 [*Acuna et al*, 1998].

< Figure 7 near here >

The MAG-ER investigation on MGS acquired magnetic field observations on over 1000 aerobraking passes during the transition to mapping orbit. On each such pass the crustal field was sampled at low altitude along a limited ground track centered on the latitude and longitude of periapsis and extending approximately +/- 25 degrees in latitude north and south of closest approach. Periapsis passes were distributed more or less randomly in longitude, as the orbit period was shortened by aerobraking, and over the course of aerobraking, the latitude of periapsis slowly evolved from ~30 degrees north latitude to ~87 degrees south. Thus MGS was able to sample crustal fields over nearly the entire planet surface, which led to a map of the global distribution of magnetic sources in the Mars crust [*Acuna et al*, 1999]. This map, reproduced here (Figure 8), is compiled by color coding the (largely north-south) subspacecraft trajectory using the radial magnetic field component measured at that latitude and longitude. No correction for spacecraft altitude, which varies along the track from periapsis (~100 km) to 200 km, was attempted in this presentation. The most significant magnetic sources are distributed non-randomly,

largely confined to the older, heavily cratered terrain south of the Dichotomy boundary; vast regions of the smooth northern plains are relatively non-magnetic, as are the Argyre and Hellas impact craters, and prominent volcanic edifices [*Acuna et al*, 1999]. This led Acuna et al. to conclude that the Mars dynamo likely operated for a brief (few 100My) period following accretion (~4.5 Ga) and had ceased to operate by the time of the late large impacts, about 3.9 Ga. More comprehensive studies of the magnetic fields above impact basins arrived at essentially the same conclusion [*Lillis et al., 2008*; *Lillis et al, 2013*].

< Figure 8 near here >

In March of 1999, MGS completed the transition to circular, polar mapping orbit with nominal altitude of 400 km, from which it would begin its mapping mission. The orbit is fixed in local time (2 am – 2 pm), and subsequent orbits cross the equator 28.6 deg westward of the previous orbit. Each "mapping cycle" of 28 days duration provides uniform sampling of the globe with track-to-track separation of about 1 degree of longitude (59 km) at the equator. Subsequent mapping cycles increase the density of ground tracks, providing complete global coverage with track-to-track spacing of 3 km at the end of 1 Mars year (687 days). MGS passes over the same location on Mars many times; thus the vector magnetic field at satellite altitude can be mapped with extraordinary signal-to-noise [*Connerney et al*, 2001]. This particular map was compiled with observations obtained over the nightside, where time variable fields due to the solar wind interaction are minimized, to improve the signal fidelity of the crustal field. The crustal field at mapping altitude reaches a maximum of ~220 nT over the intensely magnetized southern highlands. This and other compilations of MGS observations are available online at http://mgs-mager.gsfc.nasa.gov.

A much improved map of the field produced by crustal sources was obtained by using a more effective means to remove residual external fields [*Connerney et al*, 2005]. This map (Figure 9) is a 360 x 180 pixel "image" of the filtered radial magnetic field, each pixel representing the median value in a 1 deg x 1 deg latitude-longitude bin. This map represents the change in radial field, or ΔBr, in nT per degree latitude, as the spacecraft moves north to south over the darkened hemisphere. The map is superposed upon a Mars Orbiter Laser Altimeter (MOLA) shaded topography map [*Smith et al*, 1999], which appears in any pixel for which ΔBr falls below a minimum threshold of +/- 0.3 nT per degree latitude. The threshold is chosen well above the noise background, so little noise appears in the figure to obscure the context image. The extraordinary signal fidelity of this map can be appreciated by recognizing that pixels adjacent in longitude are statistically independent, by virtue of the way MGS orbits accumulate. The map spans over two orders of magnitude of signal dynamic range.

< Figure 9 near here >

The Electron Reflectometer (ER) instrument of the MAG/ER investigation [*Acuna et al,* 1992; *Acuna*

et al. 1998] provides another means of sensing – remotely – the magnitude of the magnetic field between the spacecraft and atmosphere [*Lillis et al*, 2004; *Mitchell et al,* 2007]. The technique is similar to that used in lunar orbit to infer surface magnetic field magnitudes [*Anderson et al,* 1975; *Lin et al,* 1998; *Mitchell et al,* 2008] from the properties of the (reflected) electron distribution; however, at Mars, the presence of an absorbing atmosphere must be taken into account [*Lillis et al,* 2004]. Application of this technique at Mars is particularly effective in sensing weaker fields at an altitude of about 170 km above the surface. The ER technique is, however, an indirect method of sensing the magnetic field magnitude at some distance from the spacecraft, and application of the technique requires attention to detail. The method is based on the magnetic mirror effect, and the identification of a population of electrons, sorted by pitch angle, that have reversed their motion along the magnetic field, in response to an increased field magnitude encountered en route to the surface. In the ideal case (adiabatic approximation), an electron follows a helical path along B preserving its energy and magnetic moment, $\mu = mv^2 \sin^2 \alpha/2|B|$, where α is the angle ("pitch angle") between the electron velocity v and the magnetic field. As the electron moves towards an increasing field magnitude, the electron pitch angle rotates towards 90º, at which point the electron reflects (reverses course along B) and is returned to the spacecraft. Electrons with sufficiently small pitch angle (more field-aligned) will not be reflected before reaching the surface (or being absorbed by the atmosphere) and their absence in the electron distribution allows one to identify the maximum magnetic field magnitude encountered between spacecraft and surface. The vector magnetic field measured at the spacecraft is used to infer where (beneath the orbit) the increased field magnitude was encountered, and to correct the field magnitude due to crustal sources for externally applied fields. The adiabatic approximation requires that the ratio of the electron gyro radius to the characteristic dimension over which the field changes to be small, and that the field does not change appreciably during a gyroperiod. In practice, one must assume that the field does not change direction appreciably between the spacecraft (where it is measured) and the electron's reflection point, and that drift motion due to electric fields and magnetic gradients are negligible. Parallel electric fields, which can also reflect electrons, may be recognized by examination of multiple electron energies [*Halekas et al.,* 2002] and corrected for, where necessary.

A global map of field magnitudes at ~170 km altitude [*Mitchell et al,* 2007], with a spatial resolution of ~150 km (and using a minimum B threshold of 10 nT at 170 km altitude), compares very well with the map (Figure 9) compiled from magnetic field measurements. The broad regions that appear in shaded topography in Figure 9, in which the magnetic field falls below threshold, are nearly identical to the weak or zero field regions (to an accuracy of a few nT) mapped by the ER [*Mitchell et al, 2007*; *Lillis et al, 2008*]. The ER method works best where field lines are mostly radial at satellite altitude, as over a crustal field well approximated by a radial dipole. The method will return either no result or a less well

constrained result above field lines that are essentially horizontal beneath the spacecraft. The ER method was originally developed to sense magnetic fields above the lunar crust, populated by weak sources of relatively small spatial scale or coherence. In that environment, the field measured at satellite altitude may be straight-line continued nearly all the way to the surface, due to the limited reach of the weak crustal sources. At Mars, crustal fields are much stronger and of large spatial scale, limiting the accuracy with which the field may be continued (via the straight-line approximation, i.e., non-self consistently) down toward the reflection point.

< Figure 9 near here >

Heading Three: Models

The representations routinely employed throughout the solar system – dipoles and spherical harmonic expansions – lose much of their appeal when applied to the crustal field on Mars. Such global models require an awkward number of model parameters or coefficients to represent the crustal field. For example, Cain's [*Cain et al*, 2003] spherical harmonic model extends to Nmax = 90, requiring (Nmax + 1)^2 – 1 = 8280 model coefficients to calculate the field at one point above the surface. Likewise, Purucker's many dipole model [*Purucker et al*, 2000] uses 11,500 dipoles distributed on a spherical surface (1 Rm) to represent the field by superposition. There are a variety of spherical harmonic [*Cain et al.*, 2003; *Arkani-Hamed*, 2001a; *Arkani-Hamed*, 2002] and equivalent source models [*Purucker et al*, 2000; *Langlais et al.*, 2004] to chose from. These models can be obtained from the authors in electronic form. They offer a means of extrapolating the field from the measurement surface (400 km) and volume spanned by aerobraking passes to nearby points; however, none can be expected to accurately predict the field near the source region.

Mars crustal magnetism is amenable to study using many of the methods and tools developed for interpretation of magnetic surveys on Earth. In *source modeling*, one attempts to fit the observed vector field using one or more magnetized sources, e. g., a collection of thin plates, prisms, etc., often guided by constraints suggested by local geology and material properties. General techniques for *continuation* of a potential field [e.g., *Blakely*, 1995] may also be applied to extrapolate the field away from a surface upon which the vector field is known. Downward continuation (towards the source) must be done with caution, as it is essentially a differencing operation that amplifies short wavelength random noise (exponentially in z). Source modeling must also be done with caution, as it is not possible to uniquely determine crustal magnetizations using measurements external to the source.

Connerney et al. [1999] used a source model to fit observations obtained during several aerobraking passes over the intensely magnetized southern highlands (Terra Cimmeria and Terra Sirenum). These quasi parallel features appear lineated in the east-west (cross-track) direction; many can be traced over

1000 km in length. The model consisted of multiple uniformly magnetized parallel thin plates, aligned (and extending to infinity) in the east-west direction. The volume magnetization of each strip was determined using a linear inverse methodology to best fit the vector observations along one or more aerobraking passes (Figure 10). The observations could be fit well with a model characterized by strips of alternating polarity and volume magnetizations of as much as +/- 20 A/m (assuming a 30 km thick plate). This figure is likely a lower bound, since only the product of thickness and volume magnetization is constrained by the observations [*Connerney et al*., 1999]. If the magnetization is borne in a thin layer, say 3 km, then volume magnetizations of order 100 A/m would be required. This is a rather astonishing intensity of magnetization required throughout large volumes of crust. At the other extreme, *Parker* [2003] developed a theory to establish the minimum possible intensity of magnetization consistent with the data modeled by Connerney et al. Interpolating Parker's results to a crustal thickness of 30 km, Parker finds a minimum magnetization intensity of ~10 A/m is required to satisfy MGS observations. *Connerney et al* [1999] found the magnetic lineations on Mars sufficiently reminiscent of the pattern of magnetization associated with sea floor spreading on earth that they proposed a similar mechanism – crustal spreading in the presence of a reversing dynamo – for the evolution of the Mars crust.

< Figure 10 near here >

Sprenke and Baker [2000] modeled crustal magnetization in the southern highlands using the same aerobraking observations and inverse methods used by *Connerney et al* [1999]. Their models imposed constraints on the direction of magnetization ("normal" and "reversed") and allowed for a variation in the intensity of magnetization in the cross-track direction. They found similar magnetization intensities and obtained a good fit to the observations with a sea floor spreading model, allowing for variation of magnetization intensity (but not direction) along the stripes. The more uniformly distributed observations obtained at mapping altitude over the same region were downward continued using the Fourier transform method [*Jurdy and Stefanick*, 2004]. Continuation to a 100 km altitude surface allowed a favorable comparison with the aerobraking observations, and revealed abrupt lateral changes in field direction and magnitude suggestive of, but not requiring, reversed lineations.

There is great interest in identifying a Mars paleopole, and a number of authors have sought to determine ancient pole position(s) on Mars by fitting various magnetized sources to MGS magnetic field observations [*Hood and Zakharain*, 2001; *Arkani-Hamed*, 2001b; *Sprenke and Baker*, 2000; *Arkani-Hamed and Boutin*, 2004; *Frawley and Taylor*, 2004; *Hood et al.*, 2010; *Langlais and Purucker*, 2007; *Boutin and Arkani-Hamed*, 2006]. A large number of putative paleopoles have been proposed but as yet no consensus has emerged (for a recent compilation, see *Milbury et al.,* 2012). A general consideration of non-uniqueness in the interpretation of potential fields would suggest that without additional constraints the problem remains intractable [*Connerney et al*, 2004]. In one case where *a priori* information (a

gravity high) may help constrain the location of a magnetic source (one associated with the Apollinaris volcanic edifice), a paleopole location (65º S, 59º E) near the current Mars rotational pole was found by two sets of authors [*Langlais and Purucker*, 2007; *Hood et al.,* 2010]. However, Apollinaris was active a few hundred million years *after* the inferred demise of the dynamo, based on analysis of the giant impacts [*Lillis et al.*, 2013], which ought to produce a more dependable record of dynamo history. Where the association between volcanism and magnetic field is greatest, volcanism has either completely or partially demagnetized a pre-existing magnetic imprint [*Connerney et al.*, 2005]; long-lived volcanoes, in bulk and sensed from orbit, are unlikely to be effective recorders of paleofield directions [e.g., *Lillis et al.,* 2013]. An alternative approach to locating paleopoles, based on a statistical analysis of many sources [*Sprenke*, 2005] suggests a preference for paleopoles near 17º N, 230º E, and 17º S, 50º E, near those proposed earlier based on geomorphology and analysis of grazing impacts [*Schultz and Lutz,* 1988]. In another statistical approach, *Milbury and Schubert* [2010] searched among a distribution of paleopoles, assigning a magnetization intensity and direction to portions of the crust, in such a way as to best match the observed magnetic field, taken, for convenience, from several spherical harmonic model representations. They found a weak but significant correlation for spherical harmonic degrees 2 and 3 and a set of paleopoles, and none for higher degrees. This may indicate that only a global scale correspondence exists, but it may also reflect a sampling bias, in that large portions of the crust (northern lowlands, Tharsis complex, and major impact basins) were removed from the model. The paleopoles are found at low to mid-latitudes and are suggestive of polar wander and polarity reversal [*Milbury and Schubert,* 2010]. Co-existence of near-equatorial and near-polar paleopoles can be reconciled with the expectation of a moderate angular separation of dipole (magnetic) and rotational axes if one invokes significant polar wander during the active dynamo era [e.g., *Sprenke,* 2005; *Boutin and Arkani-Hamed,* 2006; *Milbury et al.*, 2012].

Heading Three: Discussion

Mars has no global field, therefore no dynamo at present, but must have had one in the past when the crust acquired intense remanent magnetization. It is likely that a molten iron core formed early, after or during hot accretion 4.5-4.6 Ga, and for at least a few hundred million years, a substantial global field was generated by dynamo action in the core. The chronology proposed by *Acuna et al*. [1999] attributes the global distribution of magnetization to an early demise of the dynamo, prior to the last great impacts (~4 Ga) that left large unmagnetized basins in the crust. This view has been supported by more complete analyses of the large impact basins [*Lillis et al., 2008*; *Lillis et al, 2013*], leading to more precise estimates of the dynamo's demise (activity extinguished before formation of Hellas and Utopia basins 4.0-4.1 Ga). *Schubert et al*. [2000] argued in favor of a late onset of the dynamo, and propose that the

southern highlands crust acquired magnetization from local heating and cooling events that post-date the era of large impacts and basin formation. Early onset and cessation of the dynamo is difficult to reconcile with the notion of a dynamo driven by solidification of an inner core [*Schubert et al,* 1992], the preferred energy source for the Earth's dynamo. Alternatively, an early dynamo can be driven by thermal convection, with or without plate tectonics, for the first 0.5 – 1 Gyr [*Breuer and Spohn*, 2003; *Schubert and Spohn*, 1990; *Stevenson et al.,* 1983; *Connerney et al*., 2004], persisting as long as the core heat flow remains above a critical threshold for thermal convection [*Nimmo and Stevenson*, 2000].

After more than 2 full years of mapping operations, MGS produced an unprecedented global map of magnetic fields produced by remanent magnetism in the crust [*Connerney et al*, 2005]. This map (Figure 9) reveals contrasts in magnetization that appear in association with known faults; variations in magnetization that are clearly associated with volcanic provinces; and magnetic field patterns that appear shifted along small circles in the manner of transform faults at spreading centers [*Connerney et al,* 2005]. *Connerney et al* proposed that the entire crust acquired a magnetic imprint via spreading in the presence of a reversing dynamo and that erasure of this imprint occurred where the crust was buried (thermal demagnetization) by flood basalts to depths of a few km. In the remaining magnetic imprint one can identify the tell tale signature of transform faults, in Meridiani, and the signature of craters that formed both before and after the demise of the dynamo. Of course, transform faults are unique to plate tectonics, so if these features are indeed transform faults then the Mars crust formed via sea floor spreading as on Earth [*Connerney et al,* 1999; *Sleep,* 1990]. The interpretation of this map is not without controversy, and many regard the sea floor spreading analogy championed by Connerney and colleagues as speculative, in large part due to the paucity of correlative observations. The alignment of the great volcanic edifices on Mars is consistent with plate motion over a mantle plume [*Connerney et al., 2005*] or, conversely, volcanic chains formed above subducting slabs [*Sleep, 1990*; Sleep, 1994; *Yin, personal communication*, 2012]. The topographical relief along much of the dichotomy boundary has been interpreted as a series of ridge/transform fault segments [*Sleep, 1994*]. A recent structural analysis of the Valles Marineris fault zone [*Yin,* 2012] likens this trough system to the left-slip, transtensional Dead Sea fault zone on Earth joining the Red Sea spreading center to the Zagros suture zone in the north: an undisputed plate boundary. It is difficult to understand how such a structure evolved on Mars in the absence of plate tectonics.

Fairen et al. [2002] proposed that the magnetic lineations might instead be the signature of accretionary terrains formed during an early era of plate tectonics, while others have proposed the formation of a series of quasi-parallel dike intrusions as alternatives to the crustal spreading hypothesis [*Nimmo, 2000; Hood et al., 2007; Ravat, 2011*]. Using an automated cluster analysis of lineated features in Connerney et al.'s [2005] map, Kobayashi and Sprenke [2010] identified two groups, or clusters, of features that they attribute to thermoremanent magnetization of chains of intrusions formed by

lithospheric drift over mantle hotspots. However, none of these proposed mechanisms manufacture magnetic material with the uniformity, efficiency, and spatial scale of crustal spreading in the presence of a global magnetic field. To account for the enormous magnitude of the field above the Mars crust, one requires a great deal of material coherently magnetized with a rather startling intensity of magnetization. It is difficult to imagine any process that can compete with sea floor spreading in this regard.

Magnetic mapping is an even more powerful tool on Mars than an Earth, where it has been indispensible in understanding the evolution of the Earth, especially in the development of the unifying theory of plate tectonics. On Earth, magnetic surveys are complicated by the presence of the main field, which makes measurement of crustal anomalies challenging: the Earth's magnetic anomalies are very weak, compared to the main field, resulting in limited signal to noise in the measurement. Magnetic anomalies are also more difficult to interpret on Earth because they may result from either induced or remanent magnetization, or both. Mars, lacking a main field, is free of such complications; the global map of its crustal field is quite remarkable, affording an opportunity to apply the methods of exploration geophysics on a global scale. It is clear that the crust of Mars retains a magnetic imprint of its formation and subsequent evolution. That history may be read with increasing resolution as magnetic surveys are conducted at lower altitudes. One may anticipate great insights from such surveys as well as from Mars paleomagnetism when that field develops.

Heading One: Gas Giants

Heading Two: Jupiter

Jupiter, the largest (R_j = 71,372 km) and most massive (M_j = 318 M_e) planet in the solar system, resides at a mean orbital distance of 5.2 AU from the sun. Jupiter, capturing much of the gas component of the solar nebula, is largely a solar mix of H and He, possibly with a rock and ice core of some tens of Earth masses. The Jovian system has been visited often by spacecraft, both as a destination and as a waypoint for vessels destined to travel onward. A passing spacecraft can use Jupiter's considerable mass and orbital velocity to advantage, acquiring a boost to greater radial distances (e.g., Cassini, New Horizons to Pluto) or to achieve higher solar latitudes (e.g., Ulysses), than would otherwise be possible. Thus Jupiter's orbit serves as a crossroads for travelers to other worlds as well.

Jupiter was known to have a magnetic field long before the first in situ observations were obtained. *Burke and Franklin* [1955] had detected non-thermal decameter wavelength emissions (22 MHz) from Jupiter, leading to speculation that Jupiter possessed a magnetic field of internal origin. Observations at decimetric (~ 1 GHz) wavelengths subsequently revealed synchrotron radiation, emitted by high-energy

electrons trapped in a Jovian Van Allen belt. This emission, as observed, is modulated by the rotation period of the planet (9.925 hrs). Since synchrotron radiation is narrowly beamed in a direction perpendicular to the magnetic field guiding the electron motion, the modulation can be used to infer the geometrical properties of the field. Thus the tilt (9.5°) and phase (200° λ_{III}) of the Jovidipole were deduced [*Morris and Berge,* 1962], and estimates of the dipole offset (from the origin) were made [*Warwick,* l963*b*], well before the first spacecraft arrived at Jupiter. The magnitude of the field, however, remained in doubt.

Continued observations of Jovian radio emission at decameter wavelength revealed yet another periodicity in the occurrence of radio storms, one that matched the orbital period (42.46 hrs) of the innermost Galilean satellite, Io [*Bigg,* 1964]. The intensity and occurrence rate of radio storms varies dramatically with Io's orbital phase, as well as Jovian central meridian longitude (CML). Activity peaks when Io's orbital phase is near 90° and 240° (measured in the direction of Io's orbital motion from geocentric superior conjunction). Io's influence on Jovian radio emissions was attributed to an electrodynamic interaction [*Goldreich and Lynden-Bell*, 1969] that drives currents along magnetic field lines to and from Jupiter's magnetosphere.

Heading Three: Observations

Pioneer 10, the first spacecraft to enter the Jovian magnetosphere, passed within 2.8 R_j of the planet at closest approach in December 1973. Pioneer 10's flyby trajectory carried it through the Jovian system at relatively low magnetic latitudes (Figure 11). This region, near the equator, is home to a large-scale system of azimuthal currents (ring currents) extending from about 5 R_j to the outer reaches of the magnetosphere (~ 100 R_j). This current system imparts a disklike geometry to the magnetic field, with field lines at great distances stretched nearly parallel to the equator (leading to the descriptive "Jovian magnetodisc"). Precisely one year later, Pioneer 11 approached to within 1.6 R_j on a retrograde λ_{III} trajectory. The Pioneer 11 trajectory provided measurements over a wide range of magnetic latitudes (Figure 11) and longitudes.

<FIGURE 11 Near Here>

The Pioneer 10 and 11 spacecraft were instrumented to measure magnetic fields with a Vector Helium Magnetometer (VHM) provided by the Jet Propulsion Laboratory [*Smith et al.,* 1975b]. This instrument could measure a maximum of +/- 1.4 G in the highest of its 8 dynamic ranges. Pioneer 11 also carried a small high field tri-axial fluxgate magnetometer (FGM) provided by the Goddard Space Flight Center [Acuna and Ness, 1976]. The FGM was added as a precaution, to extend measurement capability to a field of 10 Gauss along each axis (Earth's field magnitude ~ 0.6 G at the pole). The magnitude of Jupiter's field was unknown at the time of launch, and the concern was that without a high field

instrument Pioneer 11 might not have the dynamic range needed to make observations throughout the encounter. Fortunately, the maximum field experienced by Pioneer 11 was ~ 1.12 G, observed just as the spacecraft went into occultation on its way to closest approach at 1.6 R_j. There are small but significant differences between the two datasets that have not been fully resolved, so most analyses have tended to use data from one or the other but not both.

The two Voyager encounters followed in March and July of 1979. Like Pioneer 10, the Voyagers remained near the Jovigraphic equator, confined to relatively low magnetic latitudes. Voyager 1 was targeted for a close flyby of Io and the Io flux tube; its closest approach to Jupiter was 4.9 R_j, just inside the orbit of Io (5.95 R_j). The maximum field measured by Voyager 1 was just ~3330 nT (*Ness et al.*, 1981). The Voyager 2 flyby was relatively distant, with a closest approach distance of 10 R_j (*Ness et al.,* 1982). In this region of the magnetosphere, the measured magnetic field has a significant contribution due to the magnetodisc currents. These will tend to reduce the θ component of the planetary field near the equator and introduce a component of the field in the radial direction which changes sign in crossing the magnetic equator. Local currents can not be modeled with potential fields and if present, may lead to large errors in estimates of the planetary field if not accounted for [*Connerney,* 1981].

Jupiter was not visited again until the Ulysses flyby on February 8, 1992, more than two decades after the Voyager flybys (*Balogh et al.,* 1992). Ulysses was designed to study the heliosphere at high solar latitudes; the Jupiter flyby was used to kick the spacecraft into a high inclination orbit about the sun. The spacecraft approached Jupiter at relatively high northern latitude, measuring a maximum field of 2372 nT just after closest approach at 6.3 R_j. This encounter was much like that of Voyager 1, in that for much of the time nearest closest approach the spacecraft was immersed in the local currents of the Jovian magnetodisc.

The recently completed Galileo mission to the Jovian system was designed to deliver an atmospheric probe, explore the distant magnetosphere and provide opportunities for many satellite encounters [*Johnson et al*, 1992]. Galileo's mission plan was not well suited to provide appreciable additional information on the planetary field. Likewise, the Cassini spacecraft encounter was a distant flyby, engineered to speed the spacecraft onward to its destination (Saturn) in the outer solar system. The spacecraft that passed relatively close to Jupiter did so once, providing in-situ observations along but one path; the spacecraft (Galileo) that enjoyed multiple orbits remained at a distance and near the equatorial plane. In order to characterize Jupiter's magnetic field, and the dynamo that generates it, observations must be obtained at close-in radial distances (where higher degree and order harmonics of the field are relatively large) and well distributed globally about the planet. The Juno spacecraft, launched in August 2011 and now en route to Jupiter for a July 2016 arrival [*Bolton et al., 2010*] will do just that. Juno's

mission plan wraps the planet in a net of observations (Figure 12) provided by 32 near-polar orbits with periapsis just above the cloud tops, uniformly distributed in longitude (separation of 12º). The Juno magnetic field investigation is designed to map Jupiter's magnetic field with very high accuracy, sufficient to characterize the field to degree and order 12 – 15, depending on the dynamo core radius (estimated 0.75 -0.85 R_j) and other assumptions [*Connerney et al., 2014*].

< Figure 12 near here >

Heading Three: Models

The large number of Jovian magnetic field models available is a result of several factors, interest in the Jovian system undoubtedly foremost among them. There is a wealth of observations of Jovian phenomena, many of which (e.g., radio emission, aurorae) depend critically on details of the magnetic field near the planet where few direct observations exist. The need is great. In addition, techniques for modeling the field have evolved, and improved, in part in response to the challenges of separating the planetary field from that produced by external currents. Techniques have also evolved to take advantage of new observations (e.g., emission from the foot of the IFT) that serve as valuable constraints on the field near the surface. Here we describe a subset of the available models, intended to illustrate the development of new approaches.

The first models of Jupiter's planetary field were the OTD approximations (D2 and D4) obtained from the Pioneer 10 and 11 VHM observations [*Smith et al.,* 1976]. Analyses of the Pioneer 11 FGM observations suggested that a more capable parameterization was needed, leading to development of the O_4 model [*Acuña and Ness,* 1976], the first of many spherical harmonic models of Jupiter's internal field. This model and the SHA model of *Smith et al*, [1976] allowed spherical harmonics of degree and order 3 (octupole) for the internal field and approximated external fields with external harmonics of degree 1 or 2. Table 3 gathers together the spherical harmonic coefficients of several models, fit to various subsets of data and using a variety of methods. Estimated errors are given for some of these models, but not reproduced here, for brevity. The original publication should be consulted.

< Table 3 near here >

The Pioneer 11 VHM and FGM observations have proven most useful for studies of the planetary field, in large part due to the relatively good spatial distribution of observations afforded by that spacecraft's trajectory. The problem of inverting flyby observations to obtain estimates of the spherical harmonic coefficients has been posed as an eigenvalue problem and solved using the singular value decomposition (SVD) method [*Connerney,* 1981]. This technique allows partial solutions to underconstrained linear inverse problems and can be used to show how even small measurement errors – such as noise or unmodeled external fields – can lead to large errors among the internal field coefficients.

Pioneer 11 also remained largely outside of the region of current flow (equatorial magnetodisc) near closest approach so external fields are less of a problem.

The Voyager 1 observations yielded a considerably reduced dipole moment, upon preliminary analysis, which was attributed to the presence of magnetodisc currents that were not adequately modeled [*Ness et al.,* 1979*a*]. Subsequent analyses of the Voyager 1 observations used an explicit model of the magnetodisc [*Connerney et al*, 1981] in combination with a spherical harmonic model for the planetary field. An inversion scheme that allowed for variation of the parameters of the magnetodisc along with those of the internal field resulted in a partial octupole model (V1 17ev) of the planetary magnetic field [*Connerney et al.*, 1982*a*]. The Voyager 1 model of the field (epoch 1979) could be compared with the Pioneer 11 model (GSFC O_4) of the field (epoch 1973) to limit Jovimagnetic secular variation ($g^0{}_1$) to no more than 0.2%/yr [*Connerney and Acuña,* 1982]. By comparison, the present-day decrease in the terrestrial $g_1{}^0$ term is about 0.075%/yr.

Preliminary analysis of the Ulysses encounter observations [*Balogh et al.,* 1992] suggested good agreement with the magnetic field calculated using the O_6 planetary field augmented with a magnetodisc current sheet model [*Connerney et al.,* 1981] if one allowed for variations of the external field during the encounter. The Ulysses close approach observations were obtained in the immediate vicinity of the inner edge of the current sheet, so some attention to external fields is required. *Dougherty et al.* [1996] found the data consistent with Connerney's O_6 model, allowing for a reduction of the $g_1{}^0$ coefficient (main dipole term), from 4.242 to 4.059 G. *Connerney et al.* [1996] derived an octupole model of the field (Ulysses 17ev), also with a reduced dipole $g_1{}^0$ coefficient (4.109 G), and demonstrated that the internal zonal harmonic coefficients covary with the parameters of the current disc, complicating attempts to separate internal and external fields. The magnetodisc current system was considerably weaker during the Ulysses epoch, relative to that observed during the Pioneer and Voyager epochs. Connerney et al. [1996] quote a figure of 20% less total integrated magnetodisc current at the time of Ulysses encounter, Dougherty et al. [1996] quoting a 28% reduction. Differences among the Ulysses era (Ulysses 17ev) model and those of Voyager (Voyager 1 17ev) and Pioneer (O_4) appear within, or comparable, to estimated parameter errors [*Connerney et al,* 1996]; so there is little evidence of secular variation of the main field from these data [*Connerney and Acuna, 1982; Dougherty et al., 1996*] or from the more distant observations acquired throughout Galileo's tour [*Yu et al., 2009; Russell and Dougherty, 2010*].

The O_6 model of Jupiter's magnetic field used observations from both Pioneer 11 (1973) and Voyager 1 (1979), assuming that the magnetic field has not changed appreciably between encounters, and adds a constraint on Jupiter's harmonic spectrum [*Connerney,* 1992]. This model was designed to benefit from the better spatial distribution of observations afforded by combining trajectories. It was also designed to allow for the possibility that unmodeled fields of higher degree and order than octupole might be present

in the observations (particularly those of Pioneer 11, passing to within 1.6 R_j). This model represents a partial solution to an underdetermined inverse problem (spherical harmonic expansion to degree and order 6) combined with an explicit model of the magnetodisc. Another distinguishing feature of this model was the application of a weights $(r_c/a)^{n+2}$ to parameters of degree n, where r_c , the core radius, was taken to be 0.7 R_j. This weighting encourages a model solution with terms of degree n contributing equally to the mean squared field amplitude on the (presumed) core boundary, as is observed for Earth, through degree 12 or so [*Langel,* 1987]. The Schmidt coefficients of this model ("O_6" for octupole part of a degree 6 expansion) are listed in Table 3. The O_6 model is characterized by smaller quadrupole and octupole moments, relative to the comparable O_4 model of *Acuña and Ness* [1976], and in that regard is more Earth-like in harmonic content (Figure 13).

< Figure 13 near here >

The discovery of infrared emission at the foot of the Io Flux Tube (IFT) in Jupiter's ionosphere [*Connerney et al*, 1993] provided an entirely new and extremely valuable constraint on magnetic field models. This emission occurs in Jupiter's polar ionosphere where field lines threading the satellite Io (or its conducting ionosphere) at an orbital distance of 5.95 R_j intersect the Jovian ionosphere. The feature is observed to move across the disc of Jupiter in concert with Io's orbital motion, tracing a path around each magnetic pole (Figure 14). This path serves as a fiducial marker, providing an unambiguous reference on Jupiter's surface through which magnetic field lines with an equatorial crossing distance of 5.95 R_j must pass. Thus observations of the location (latitude, longitude) of the footprint offer a unique constraint on magnetic field models, precisely where it is most needed, on the surface of the planet, and otherwise unavailable. This emission is observable from Earth with ground telescopes (e.g., Infrared Telescope Facility on Mauna Kea) in the infrared region of the spectrum (*Connerney et al*, 1993; *Connerney and Satoh*, 2000), and with the Hubble Space Telescope in the ultraviolet [*Clarke et al*, 1995; *Clarke et al,* 1996; *Prange* et al., 1996; *Clarke et al.,* 2005, *Bonfond et al., 2009*]. Emission has now been detected at the foot of the Europa and Ganymede flux tubes as well [*Clarke et al.,* 2002; *Grodent et al.,* 2006; 2009]. The Europa and Ganymede observations are less useful constraints on the internal field because these field lines pass through the equator at greater radial distance (9.4 and 15 R_j, respectively) where relatively large and time variable external fields are encountered. However, the Ganymede footprint supplies a useful fiducial marker for locating the source of the main aurora (*Clarke et al.,* 2002), since emissions poleward of the Ganymede footprint must map to regions beyond its orbit (15 R_j).

< Figure 14 near here >

Over 100 determinations of the location of the IFT footprint were combined with in-situ magnetic field observations (Pioneer 11 VHM and Voyager 1 FGM) to obtain a partial solution to a 4th degree and

order expansion of the internal field [*Connerney et al,* 1998]. This model is referred to as the "VIP4" model (Table 3), reflecting use of Voyager 1, IFT, and Pioneer observations; and an internal spherical harmonic expansion to degree and order 4. This model was designed to fit the position of the IFT footprint very well (within ~1° latitude), so it has proven particularly useful in analyses of satellite interactions and aurorae. However, no one model fits the IFT footprints and *all* of the in-situ magnetic field observations to within their expected errors [*Connerney et al.,* 1998], although the IFT observations can be so fit with any *one* set of flyby observations.

Surprisingly, the IFT footprint observations are nearly sufficient *by themselves* to yield a useful field model. All that is needed is a means of constraining the magnitude of the field. The final model in Table 3 lists a model derived using over 500 IFT footprint locations [*Connerney et al*., in preparation]. This model constrains the field magnitude using a minimum of in-situ magnetic field data: the theta component, only, of the Voyager 1 closest approach data obtained within 7 R_j. The Voyager 1 data was chosen as the most accurate (~0.1%) available and the theta component, ranging from 1036 nT to a maximum of 3280 nT, as the least influenced by external fields (larger in magnitude than the radial component of the external field but easily modeled). This model ("VIT4", reflecting use of IFT observations, Voyager theta component, degree and order 4) is essentially a product of remote observation, using just enough in situ data to constrain the magnitude of the field. In principal, any constraint on the field magnitude – such as the field magnitude on the surface, deduced from the maximum frequency of radio emission at the electron cyclotron frequency – would serve as well if properly understood.

Spherical harmonic models confined to relatively low degree and order (e.g., 4) fit the Io flux tube footprint observations well, but there remains a systematic and non-negligible residual, particularly in the area of the "kink" that appears in UV observations of satellite footprints and aurorae near 110º Jovian system III longitude. Grodent and colleagues [*Grodent et al, 2008*] have suggested that a more localized source, or magnetic anomaly, might account for the "kink". This additional source could not be uniquely determined, but these authors demonstrated that the observations of satellite and auroral footprints in the northern hemisphere can be reproduced using a model of this type. The putative magnetic anomaly would necessarily have to be located near the surface in order to modify the satellite and aurora footprints without dramatically altering the field elsewhere. In perhaps the most ambitious such modeling attempt to date, Hess and coworkers utilized an extensive database of IFT footprints compiled by Bonfond et al [2009] to obtain a 5th degree and order spherical harmonic model ("VIPAL") of the Jovian magnetic field [*Hess et al., 2011*]. This model obtains a better fit to the satellite footprints, at the expense of a poorer fit to the in-situ magnetic field observations inward of about 4 R_j. However, VIPAL was also constrained to minimize the longitudinal separation between the observed and modeled IFT footprint, and it was adjusted to produce a prescribed surface magnetic field strength along the IFT "footpath". This was done

to better match the local electron gyrofrequency to observations of Jovian radio emissions, under certain assumptions regarding the source location, beaming, and generation of radio emissions. Thus models of this type (e.g., VIPAL) require a certain degree of confidence in our understanding of the propagation of Alfven waves from Io to Jupiter (longitude constraint), and a degree of confidence in our understanding of the generation and propagation of Jovian radio emissions (surface magnitude constraint) as the price to be paid for encompassing more observables. In contrast, models based entirely on in-situ magnetic field observations rest on our unshakable faith in the extrapolation of a potential field through a region of space devoid of local currents. In a few short years, the Juno spacecraft will enter Jupiter orbit (July 2016) to map Jupiter's magnetic field both closer to the planet and with much greater accuracy than ever before. Shortly after Juno's prime mission begins it should be possible to discriminate among the approaches outlined above.

Heading Three: Discussion

The magnitude of the surface field for the 0_6 model ranges from just over 3 G at low latitudes to just over 14 G at high (northern) latitudes. This is similar to the range of field magnitudes predicted by the O_4 model of *Acuña and Ness* [1976]. The variation of magnetic field magnitude on the surface of Jupiter is illustrated in Figure 15, which depicts contours of constant field magnitude on the dynamically flattened surface of Jupiter, computed using the VIT4 model. This model has a less prominent north-south asymmetry in the magnetic field magnitudes, compared to earlier models; the maximum field in the northern hemisphere (13.9 G) is comparable to that of earlier models (see *Acuna et al,* 1983), but the maximum field in the southern hemisphere (13.2 G) is considerably larger. The estimated uncertainty in the surface field magnitude is about +/- 1 Gauss (*Connerney*, 1981; *Connerney*, 1991), but could be appreciably greater if higher degree and order harmonics are larger than expected, or, equivalently, if a near-surface "magnetic anomaly" is invoked (e.g., *Grodent et al,* 2008). Modeled surface magnetic field magnitudes for the VIP4 and VIPAL models are not dramatically different [*Hess et al,* 2011]; however, Grodent et al.'s magnetic anomaly model results in much larger surface field magnitudes in the northern mid-latitudes near 150º – 180º system III longitudes. Figure 15 also illustrates the foot of the Io flux tube ("Io foot"), which is the path traced out on the planet's surface by the field lines which pass through the satellite Io as the planet rotates. The foot of the Io flux tube passes through the region of highest field strength in the northern hemisphere at a longitude of about 150° λ_{III}. The maximum surface magnetic field magnitude present along the Io foot is generally consistent with the maximum frequency extent (39.5 MHz) of Jovian decameter radio emission (DAM), assuming that the emission occurs at the local electron gyrofrequency and at the foot of the Io flux tube (f_c[MHz] = 2.8 B [G]). However, significant challenges

arise in modeling Jovian radio emissions using such models, and this has motivated a number of authors to explore alternatives.

< Figure 15 near here >

Heading Two: Saturn

Saturn, the second largest (R_S = 60,330 km) and most massive (M_j = *95 M_e*) planet in the solar system, resides at a mean orbital distance of 9.5 AU, nearly twice as far from the sun as Jupiter. Like Jupiter, Saturn exhibits the characteristics of a condensed object of solar composition. Saturn was assumed to be much like its larger companion before Pioneer 11 arrived. However, unlike Jupiter, Saturn as a radio source was relatively quiet. Saturn is too weak a radio source and too distant to be easily detected from Earth. In addition, Saturn emits at lower frequencies (a few kilohertz to 1 MHz) than Jupiter, well below the critical frequency required to penetrate the Earth's ionosphere. Thus Saturnian radio emissions cannot be routinely monitored using terrestrial radio receivers.

Heading Three: Observations

The discovery of Saturn's magnetic field was thus left to Pioneer 11, arriving in September 1979, following its Jupiter swingby. The particle and fields investigations on board Pioneer 11 (see previous section) obtained observations along a spacecraft trajectory that remained within approximately 6° of the equator (Figure 16). The most remarkable feature of Saturn's magnetic field proved to be the close alignment of the magnetic dipole and rotation axes [*Smith et al.,* 1980*a*; *Acuña and Ness,* 1980; *Acuña et al.,* 1980], also noted by the charged particle investigations [*Simpson et al.,* 1980; *Van Allen et al.,* 1980]. Pioneer 11 approached to within 1.35 Rs, measuring a maximum field of ~8200 nT at -6° latitude [*Smith et al.,* 1980].

< Figure 16 near here >

The Voyager 1 encounter in November 1980 and the Voyager 2 encounter in August 1981 provided the first observations of Saturn's magnetic field at high northern and southern latitudes [*Ness et al.,* 1981, 1982]. Voyager 1 sampled relatively high latitudes (-40°) at close radial distance (3.07 R_s) but remained in the southern hemisphere while inside of 6 R_s radial distance. Voyager 1 measured a maximum field of 1093 nT at -40° latitude and 184° SLS longitude, just prior to closest approach [*Ness et al*., 1981]. Voyager 2's closest approach of 2.69 R_s occurred at 323° longitude, diametrically opposite those of Pioneer 11 and Voyager 1. Voyager 2 sampled relatively high latitudes in both hemispheres and measured a maximum field of 1187 nT just prior to closest approach [*Ness et al.,* 1982].

More than two decades passed before the arrival of Cassini at Saturn. The Cassini spacecraft has been in orbit about Saturn since its orbit insertion (SOI) on June 30, 2004, and has remained relatively distant

subsequent to the orbit insertion maneuver [*Dougherty et al*, 2004b], targeting multiple satellite flybys from the vantage point of a near equatorial orbit. During SOI, the spacecraft periapsis was just 1.33 Rs; this is the closest Cassini would come to Saturn throughout its prime mission. Cassini is instrumented with two magnetometers, a vector fluxgate and a helium magnetometer, the latter operated in a mode to measure field magnitudes near the planet [*Dougherty et al,* 2004a]. However, a large number of orbits have been executed thus far, many with periapses < 4 Rs.The spacecraft is just now executing a series of maneuvers designed to increase the inclination of the orbit, which will provide a more favorable distribution of observations for internal field analysis, particularly near the end of mission, with many orbits planned with periapses < 3 Rs.

Heading Three: Models

The Pioneer 11 magnetic field observations were consistent with a dipole field of moment ~ 0.20 G-$R^3{}_S$, slightly offset to the north of the planet's center [*Smith et al.,* 1980*a*; *Acuña and Ness,* 1980; *Acuña et al.,* 1980] by about 0.04 or 0.05 R_S. The simple northward-offset dipole model was affirmed by independent analyses of charged particle absorption signatures [*Chenette and Davis,* 1982] which serve as a probe of the geometry of the field. Analyses of the Pioneer 11 high-field fluxgate magnetometer observations [*Acuña et al.,* 1980] found no departure of the field from axisymmetry; however, in Saturn's relatively weak field, the resolution of the high-field fluxgate was quite limited. Working with higher resolution vector helium magnetometer observations, *Smith et al.* [1980*b*] proposed a dipole tilt of ~ 1°, although the orientation of the dipole could not be determined.

The combined Voyager 1 and 2 magnetic field observations were not consistent with the field of a simple displaced dipole, nor did these observations reveal any measurable departure from spin-axisymmetry (of the internal field). An axisymmetric, zonal harmonic model of degree 3 was found to be both necessary and sufficient to describe the magnetic field of Saturn [*Connerney et al.*, l982*b*]. This three-parameter model is referred to as the Z_3 model (zonal harmonic of degree 3) and it is listed in Table 4; the nonaxisymmetric terms (*m* > 0) are insignificant. Independent analyses of the Voyager 1 and 2 observations suggested an estimated uncertainty of ~ 100 nT for the Z_3 coefficients. This is perhaps the best way to estimate errors when model residuals are highly correlated – a comparison of independent analyses, using separate flybys. The Z_3 model parameter uncertainties are likely to be smaller, benefiting from the combined observations of both Voyager flybys. The magnetic and rotation axes are indistinguishable, aligned to within ~ 0.1°.

< Table 4 near here >

A series of analyses followed to test the high degree of axisymmetry of the Z_3 model (reviewed in *Connerney et al.* [l984*b*]). *Acuña et al.* [1983*a*] showed that the charged particle absorption signatures

observed in Saturn's magnetosphere were consistent with the Z_3 model. A reanalysis of the absorption signatures studied by *Chenette and Davis* [1982], allowing for the presence of a nonzero g_3^0 term, found these signatures to be best fit by a zonal harmonic model which was indistinguishable from the Z_3 model [*Connerney et al.,* l984*b*]. A re-analysis of the Pioneer 11 vector helium magnetometer observations [*Connerney et al.*, 1984*a*] identified a spacecraft roll attitude error (of 1.4° in magnitude) as the primary source of the discrepancy between the Pioneer 11 and Voyager results. After correcting the spacecraft roll attitude, these authors demonstrated that the Pioneer 11 VHM measurements were consistent with the Z_3 model field, and axisymmetry, to better than the accuracy of measurement [*Connerney et al.*, 1984*a*]. The roll attitude error was subsequently independently verified by analysis of Pioneer 11 imaging photopolarimeter data. After correcting for the Pioneer 11 roll attitude error, *Davis and Smith* [1985, 1986] combined the Pioneer 11 VHM and Voyager magnetometer observations and obtained a zonal harmonic model (SPV) in substantial agreement with the Z_3 model (Table 4).

Analysis of the Cassini fluxgate magnetometer observations obtained during SOI confirmed (to the level of estimated measurement accuracy, approximately 0.1% and 0.1°) the spin-symmetry of Saturn's magnetic field [*Dougherty et al.*, 2005]. An unfortunate command error prevented the helium magnetometer from making measurements of the field magnitude during SOI. A zonal harmonic model of degree 3 was obtained (Table 4) from the fluxgate observations; it is very similar to the earlier zonal harmonic models derived from Voyager and Pioneer observations, but for a smaller g_3^0 term. Figure 17, in which we plot the vector magnetic field observed throughout Cassini's orbit insertion maneuver, illustrates the remarkable spin symmetry of Saturn's internal field. The absence of a significant azimuthal field component during this close encounter suggests a negligible tilt of the dipole and the lack of higher order contributions to the field.

< Figure 17 near here >

More revealing, perhaps, is Figure 18, which plots the vector residual after the best fitting zonal harmonic model has been removed from the observations. Residuals that *are* observed can be understood by attention to engineering practicalities, for example spacecraft attitude maneuvers associated with turns to avoid impact of ring material on sensitive components ("turn turtle", i.e., assume a protected attitude crossing the ring plane) and the Earth point maneuver. Errors in the measured vector magnetic field can arise from attitude uncertainties associated with spacecraft maneuvers or with variations in sensor response as the field direction in the sensor reference frame changes. A turning spacecraft often depends on gyros for attitude information, lacking a celestial reference, and slight errors in gyro drift rates may result in attitude errors during maneuvers. Likewise, one sees an increase in variations of the field associated with the main engine burn and its aftermath. This figure also illustrates that the residuals are highly correlated, and dominated by systematic effects, most of which can be readily understood and

related to spacecraft engineering events; thus, estimates of model parameter uncertainties based on propagation of errors and Gaussian statistics are likely to be overly optimistic.

< Figure 18 near here >

Near the end of Cassini's prime mission, a number of orbits were acquired with periapses inside the orbit of Enceladus (3.95 Rs), providing much-needed close-in observations at higher latitudes (though still within about +/- 30º of the equator). More recent modeling efforts have assembled a large number of orbits obtained over a lengthy span of time, fitting zonal harmonic internal fields with external fields described by external spherical harmonics or an explicit ring current (or magnetodisc) model. These models are listed in Table 5. Burton et al. [2009] analyzed observations (45 individual orbits) acquired during Cassini's first three years (July 2004 – July 2007) utilizing an explicit ring current model to represent external fields. They find no departure from axisymmetry in the internal field and little evidence for secular variation of the field over the thirty years between the Pioneer/Voyager era and the Cassini era. A later study by the same authors [*Burton et al.,* 2010] used observations confined to the region inside the orbit of Enceladus at 3.95 Rs (to mitigate external effects) and represented external fields with a spherical harmonic expansion, reaching the same conclusions listed above. The latter study also explored a range of putative Saturn rotation periods, finding no preference for a different rotation rate and no significant non-axial internal field. Likewise, Cao et al. [2011] modeled observations obtained over a span of five years within the Enceladus L-shell, using a simple uniform external field representation. They also find that a zonal harmonic internal field of degree three is sufficient to describe Saturn's magnetic field. They arrive at a reduced upper limit for the tilt of the dipole (less than 0.06º). Their more recent reanalysis of the Cassini SOI observations [*Cao et al.,* 2012] – fixing the zonal harmonics with degree 1-3, and performing a Monte Carlo search for a best fitting g_4^0 and g_5^0 – resulted in the identification of non-zero 4th and 5th degree zonal harmonic terms. These are both very small in magnitude, however, and not much larger than the estimated parameter uncertainties (which are likely to be quite optimistic; see, e.g., Figure 18).

< Table 5 near here >

Cassini, in its "Solstice" extended mission, now enters a phase in which the orbital inclination increases, providing close-in observations of the magnetic field at progressively higher latitudes. With better spatial coverage, and as more observations accumulate, one may expect some refinement of the spherical harmonic models derived from Cassini.

Heading Three: Discussion

Magnetic field magnitudes (Z_3 model) on the surface of a dynamically flattened Saturn (flattening 1/10.6) range from a minimum of 0.18 G at the equator to maxima of 0.65 G and 0.84 G at the southern

and northern poles, respectively (Figure 19). Note that Saturn is the most dynamically flattened of all the planets, and the polar magnetic field is substantially increased by the decreased radial distance (0.9 R_S) at the poles. The equatorial field magnitude is less than that of a simple dipole or offset dipole, and polar field magnitudes (both north and south) exceed those of either a the centered or offset dipole. An eccentric dipole representation of the Z_3 model has a northward displacement of the dipole of 0.038 R_S along the rotation axis (*Connerney*, 1993), in agreement with the Pioneer 11 OTD representations. However, the ratio of g^0_3/g^0_2 is too large by a factor of 30 to be attributed to a dipole offset. The Z_3 model's large antisymmetry with respect to the equator plane cannot be removed by a simple coordinate system translation. The same can be said of all of the Saturn zonal harmonic models listed in tables 4 and 5.

< Figure 19 near here >

The high degree of spin axisymmetry of Saturn's magnetic field is unique among the planets. This unanticipated symmetry is somewhat disquieting from an observational perspective, as it lends no simple explanation for observations of periodic magnetospheric phenomena, e.g., periodicities in magnetic field and particle observations in the distant magnetosphere [*Galopeau et al.,* 1991; *Connerney and Desch*, 1992; *Giampieri and Dougherty*, 2004; *Giampieri et al.*, 2006; *Cowley et al.,* 2006; see also *Kaiser et al*, 2005 and *Giampieri et al*, 2005], the modulation of Saturn kilometric radio emission [*Desch and Kaiser,* 1981; *Kaiser and Desch,* 1982], optical spoke activity in Saturn's B ring [*Porco and Danielson,* 1982], and ultraviolet aurorae [*Sandel et al.,* 1982]. All of these phenomena would presumably be easier to explain if a magnetic anomaly could be found. However, it is not clear how such a slight departure (e.g., near the ~0.1° limit of detection) from axisymmetry might so dramatically influence these phenomena, which occur at different magnetic latitudes and as far as 1 or 2 R_S above the surface.

The uncertainty regarding the origin of Saturn's quasi-periodic magnetospheric phenomena has spilled over into a related topic: Saturn's rotation period. Accurate measurement of the rotation period of gaseous planets relies on the long term observation of magnetospheric radio emissions, which are assumed to be locked in phase with the magnetic field and interior of the planet. Long term observation of Jupiter's radio emissions, for example, yield a consistent estimate [*Riddle and Warwick*, 1976] of rotation period (9 hr, 55m, 29.71s) regardless of frequency of emission measured or method of analysis [*Carr et al.,* 1983]. Note that the uncertainty associated with this estimate is about 0.02s [*Carr et al*, 1983], reflecting observation over many years. Saturn's rotation period was measured in the Voyager era (1980-1981) using periodicities in kilometric radio emissions (SKR), yielding an estimate of the rotation period (10h, 39m, 24s) with an uncertainty of +/- 7s [*Desch and Kaiser,* 1981; *Kaiser et al.,* 1984]. The larger uncertainty of this estimate is due to the limited span of time (9 months) over which SKR could be monitored. Continued observation of Saturn's radio emissions, however, in the Ulysses era [*Galopeau*

and Lecacheux, 2000] and during the Cassini mission [*Gurnett et al.*, 2006; *Gurnett et al.,* 2007; *Kurth et al*, 2007] revealed variations of about +/-1% (~6 minutes) in the apparent radio period – far in excess of that allowed by statistical uncertainties. Likewise, direct measurement of periodicities in the magnetic field [*Giampieri and Dougherty*, 2004; see also *Kaiser et al.* 2005 and *Giampieri et al*, 2005; *Giampieri et al,* 2006; *Cowley et al.*, 2006] yielded inconsistent rotation periods, at best, with little promise of establishing a true rotation rate [*Sterenborg and Bloxham*, 2010].

Saturn's radio emissions are influenced by variations in the solar wind [*Desch,* 1982; *Desch and Rucker*, 1983] and are emitted from a source region fixed in local time [*Kaiser and Desch,* 1982; *Kaiser et al,* 1984; *Galopeau et al,* 1995]. It has thus been proposed that variations in the apparent periodicity of radio emissions are related to variations in the solar wind [*Cecconi and Zarka,* 2005]. Gurnett et al. [2007] observed that the SKR period derived from Cassini radio observations differed substantially from that deduced from the Voyagers; even worse, the SKR-derived period was found to be anything but constant in time [*Gurnett et al., 2007*]. Thus it is not clear which of the proposed rotation periods, if any, closely matches that of the deep interior of the planet.

Not surprisingly, then, interest has turned to other means of establishing Saturn's rotation period, and even questioning what it means to assign such a number to a fluid planet [*Stevenson, 2006*]. The Voyager era rotation period was never warmly embraced among atmospheric scientists because this period results in an atmosphere that superrotates at nearly all latitudes [*Smith et al., 1982*; *Melendo et al., 2011*]. In contrast, Jupiter's atmosphere is characterized by alternating eastward and westward jets, referenced to its system III (radio) period. Thus a number of proposals have emerged that would have Saturn rotating significantly faster. Anderson and Schubert [2007] treat Saturn's rotation period as a free variable and find the period (10 h, 32 m, 35±13 s) that best reconciles gravity, radio occultation, and wind observations; they propose this as a more accurate period of rotation of Saturn's interior. The validity of this method is somewhat bolstered by application to Jupiter [*Helled et al. 2009*], where it results in a period that agrees well with the radio-derived period. A similar Saturn rotation period (10 h, 34 m, 13±20 s) is obtained from a consideration of atmospheric dynamic stability [*Read et al., 2009;* see also *Showman, 2009*]. Either proposal results in a more Jupiter-like system of eastward and westward flowing jets that would be expected of a giant planet with a water-rich atmosphere driven by large scale latent heating [*Lian and Showman, 2010*].

This is the subject of intense activity at present, so perhaps a better understanding of Saturn's rotation will be forthcoming soon. An accurate value for the rotation period is essential to arriving at an understanding of Saturn's interior structure and composition.

The magnetic fields of Earth, Jupiter, and Mercury all favor dipole tilts of approximately 10° in

magnitude; Saturn's dipole tilt is not yet determined but < 0.06° or so. The high degree of spin axisymmetry is surprising but not in and of itself a problem for dynamo theory [e.g., *Lortz,* 1972]. The Cowling antidynamo theorem [*Cowling,* 1937, 1957], which states that regenerative dynamo action cannot occur if the magnetic field and fluid motions are axisymmetric, applies to the field in the dynamo region, not the field observed above. It is, however, necessary to explain why Saturn's magnetic field is so dramatically different from Jupiter's, given the similarity in size, composition, and rotation.

Stevenson proposed a model in which the field is axisymmetrized by the differential rotation of an electrically conducting shell above the dynamo region [*Stevenson,* 1980, 1982; also *Kirk and Stevenson,* 1987]. This nonconvecting shell forms above the dynamo region as a result of the immiscibility of helium and hydrogen under the temperature, pressure conditions in this region (the larger planet, Jupiter, would have no such region). In this model, the formation of helium raindrops results in a stably-stratified conducting shell which, in differential rotation with respect to the core, attenuates any nonaxisymmetric fields generated below. *Schubert et al.* [2004] treated the more general case of dynamo generation in a convective core within an electrically conducting shell, taking into account the mutual coupling between the two. These authors conclude that the observable characteristics of planetary dynamos are largely determined by the nature of the (electrically conducting) flow above the dynamo region. That is to say, the mere presence of a stably stratified, conducting shell in differential rotation above the dynamo is not sufficient to axisymmetrize the field [*Stanley and Mohammadi*, 2008]; but it may act to axisymmetrize the field if it is sustained by a thermal wind as a result of a pole to equator temperature difference [*Stanley*, 2010; *Christensen and Wicht*, 2008]. The dynamo models studied thus far do not go quite far enough in axisymmetrizing the field, producing dipoles with 1-2º angular separation between magnetic and spin axes, compared to Saturn's < 0.1º dipole tilt.

The Z_3 model of Saturn's field can be used (with caution) to predict the field inside the planet, but above the dynamo (or regions of current flow). The radial field component is of particular interest since it is continuous across the core boundary, and if the frozen flux assumption is satisfied, serves as a tracer of horizontal fluid motion at the outer radius of the dynamo [*Backus,* 1968; *Benton,* 1979; *Benton and Muth,* 1979]. Figure 20 shows the latitudinal variation of $|B_r|$, computed using the Z_3 model, on the surface of a sphere of radius $r = 0.5\ R_S$, which is the approximate outer boundary of the metallically conducting core [*Stevenson,* 1983]. The Z_3 model (via the g^0_3 term) enhances the field at the poles and reduces the radial field near the equator, relative to a simple dipole. In fact, the Z_3 model is very nearly identical to that required to minimize the low latitude unsigned flux across the core surface [*Connerney,* 1993]. Cao et al. [2012] reach a similar conclusion with their preferred zonal harmonic model, albeit with a dynamo core radius (0.4 R_s) that is significantly smaller than the radius (0.64 R_s) of metallically conductive Hydrogen [*Weir et al.,* 1996]. The concentration of radial flux at the spin poles, and minimization of radial flux at

low latitudes, may be the signature of a dynamo axisymmetrized by differential rotation of a conductive layer above.

< Figure 20 near here >

The high degree of symmetry (about the rotation axis) of Saturn's magnetic field is a distinguishing feature that may have some unexpected consequences throughout the system. For this planet, a specific radial distance in the equatorial plane maps (along the magnetic field) uniquely to a corresponding latitude on the surface [*Connerney,* 1986], as is illustrated in Figure 21. Thus one may associate a radial distance (ring plane conjugate) with any northern or southern latitude. From a practical standpoint, all of the zonal harmonic models introduced subsequent to the Z_3 model yield a very similar mapping from equator to ionosphere. Narrow dark bands that encircle the planet at 44.2° N, 46.3° N, and 65.5° N have been attributed [*Connerney,* 1986] to an influx of water guided along magnetic field lines from sources in the ring plane: two locations at the inner edge of the B ring (1.525 R_s, 1.62 R_s) where high charge to mass ratio particles (e.g., sub-micron ice or dust) become unstable [*Northrop and Connerney*, 1987] and a third (3.95 R_s) at the orbit of Enceladus and the E ring. The B-ring sources are identified with an electromagnetic erosion mechanism that sculpted the rings [*Northrop and Connerney*, 1987], a process that over tens of millions of years removed most of the mass from the C-ring, for example. Recent Keck observations of H_3^+ emissions from Saturn's ionosphere [*O'Donoghue et al.,* 2013] reveal a latitudinal variation in ion density that suggests transport of water in the form of high charge-to-mass ratio particles along magnetic field lines from sources in the rings [*Connerney,* 2013]. The Enceladus/E-ring source may reflect persistent activity of the recently discovered geysers on Enceladus [*Kivelson*, 2006; *Kargel,* 2006].

< Figure 21 near here >

Heading One: Ice Giants

Heading Two: Uranus

Uranus, residing at a mean orbital distance of 19.2 AU, was a mystery prior to the Voyager 2 encounter with Uranus in January 1986. The planet appears bland and featureless when observed from the Earth even under the best circumstances. Uranus and Neptune are similar in size (approximately 25,000 km radius), mass (about 15 Earth masses), and presumably composition (largely low-temperature condensates, or "ice": H_2O, NH_3, CH_4). One might therefore expect that their magnetic fields would be similar. Uranus is unique in that its rotation axis lies very nearly in its orbital plane, which is within 1° of the ecliptic plane. The actual inclination of its equatorial plane to that of its orbit is 82°. Twice during each orbit about the sun (84 years), the rotation axis is oriented very nearly along the planet-Sun line, so that we on Earth look upon the pole. This was the case during the Voyager 2 encounter in 1986; the

southern hemisphere was directed toward the sun.

Uranus was not known to have a magnetic field prior to the Voyager encounter, having maintained "radio silence" prior to encounter. It is a relatively weak radio source, emitting only at lower frequencies (below about 1 MHz), making detection from Earth difficult. In addition, the peculiar "pole-on" encounter geometry kept the Voyager spacecraft beyond the reach of radio emissions. Voyager 2, on a flyby trajectory (Figure 22) approaching to within 4.2 R_u, provided the first and thus far only measurements of the magnetic field [*Ness et al.,* 1986]. These measurements led to the rather surprising discovery of the first "oblique rotator," a planet with a large angular separation between magnetic and rotation axes.

< Figure 22 near here >

Heading Three: Models

Preliminary analyses of the Voyager magnetic field observations used an OTD representation of the field. This model is characterized by a dipole of moment 0.23 G R^3_u (1 R_u = 25,600 km), displaced along the rotation axis by 0.33 R_u and inclined by 60° with respect to the rotation axis [*Ness et al.,*1986]. Since Voyager remained somewhat distant, an OTD representation is expected to be a good approximation to the field, and indeed it fits the observations well [*Ness et al.,* 1986]. The very large dipole offset of the OTD is an indication of a planetary magnetic field with a large quadrupole, and likely higher degree and order terms, i.e., a very complex field.

A spherical harmonic analysis of the field required an expansion to degree and order 3 (octupole) to adequately represent the measurements [*Connerney et al.,* 1987], which is somewhat surprising in view of Voyager's relatively large periapsis (4.2 R_u). The Voyager trajectory provided a fairly good spatial distribution of observations, sufficient to resolve, or determine, spherical harmonic terms of degree 1 and 2, the dipole and quadrupole. This partial solution to the underdetermined inverse problem, designated "Q_3" (Quadrupole part of an expansion to degree and order 3), is listed in Table 6. The relative magnitude of the quadrupole is quite large; the octupole is likely to be large as well, but cannot be meaningfully constrained by the observations alone. Holme and Bloxham [1996] obtained octupole coefficients utilizing a regularization technique that imposes a smoothness constraint on the field; the resulting model is quite similar to that obtained by Connerney et al. [1987] among the lower degree coefficients, and presents a plausible extension of the model field to higher degree and order. Lack of knowledge of higher degree and order terms may be an important consideration, particularly in estimation of the field near the planet. The Q_3 model was extended to higher degree and order using the mapped location of near-surface ultraviolet auroral emissions which were assumed to lie on conjugate L-shells [*Herbert, 2009*]. The Q_3 model provides a low degree and order approximation only, and it is likely that surface fields are more

complex than illustrated here.

< Table 6 near here >

The magnetic field magnitude on the surface of Uranus reaches a maximum of approximately 1 G at mid-latitudes in the south, and a minimum of ≤0.1 G at middle to high northern latitudes (Figure 23). Figure 23 also illustrates the foot of the Miranda flux tube, which is the path traced out on the planet's surface by the field lines passing through the satellite as the planet rotates. Also shown are estimates of the position of the aurorae (stippled regions) and the paths of multiple-dip equators (where $B_r = 0$). Uncertainties in the higher degree and order components of the field, which are substantial, may lead to large differences in the actual surface field, particularly in the weaker field regions. Note that for Uranus, the convention adopted by the Voyager particles and fields investigations is at odds with the International Astronomical Union (IAU) latitude convention. The Voyager system has planetary latitudes positive over the hemisphere in which the angular momentum vector resides, i.e., the hemisphere oriented toward the Sun (and Earth) at the time of the Voyager encounter. The IAU defines the south pole as the rotation pole south of the ecliptic, without regard to the direction of rotation.

< Figure 23 near here >

Heading Two: Neptune

Neptune, residing at a mean orbital distance of 30 AU, is the most distant of the Jovian planets. It is similar in size (approximately 25,000 km radius), mass (about 15 Earth masses), and composition (largely low-temperature condensates, or "ice": H_2O, NH_3, CH_4) to Uranus. Very little was known of Neptune prior to the Voyager 2 encounter in August 1989. The first evidence of Neptune's magnetic field appeared in radio observations [*Warwick et al.*, 1989] obtained by Voyager 2 just days prior to its closest approach to the planet on August 25, 1989.

The in situ magnetic field observations revealed a magnetic field of internal origin with characteristics not unlike those of the planet Uranus [*Ness et al.,* 1989]. The most striking difference between the Uranus and Neptune encounters was a difference of our own design: the Voyager 2 flyby trajectories at each planet. Where Voyager 2 at Uranus remained relatively distant from the planet, Voyager 2 at Neptune approached to within 1.18 R_n of the planet (Figure 22). The presence of higher degree and order components of the field in the latter observations is overwhelming, simply a consequence of the close periapsis. Preliminary analyses of the Voyager magnetic field observations used an OTD representation of the field, valid at moderate distances from the planet (4 to 15 R_n radial distance; 1 R_n = 24,765 km), and fitted to data in excess of 4 R_n from the planet only. In this first approximation, the field may be characterized as that of a dipole of moment 0.13 G R_n^3, offset from the center of the planet by a rather surprisingly large 0.55 R_n and inclined by 47° with respect to the rotation axis [*Ness et al.*, 1989].

Heading Three: Models

It was necessary for *Ness et al.* [1989] to exclude from their OTD analysis all observations of the field obtained at $r \leq 4\,R_n$. One indication of how nondipolar the field of Neptune is can be had by comparison of the actual field measured at closest approach (approximately 10,000 nT) to that one would expect if the field were dipolar (6500 nT, for the OTD model). In fact, the field measured near closest approach was quite complex, as is illustrated in Figure 24. Two local maxima appear in the magnitude of the field observed along the flyby trajectory, instead of the one maximum one would observe in a dipolar field. Spherical harmonic analysis of the field required an expansion to degree and order 8 to adequately represent the measurements [*Connerney et al.,* 1991]. The large number of parameters associated with a spherical harmonic expansion of degree and order 8 (80 Schmidt coefficients) and the rather limited observations leads to a severely underdetermined inverse problem which has only a partial solution [*Connerney et al.,* 1991]. The distribution of observations is sufficient to resolve, or determine, all of the spherical harmonic terms of degree 1 (dipole), some of the terms of degree 2 (quadrupole), and some terms of degree 3 (octupole) and higher. The resulting model, designated "O_8" (octupole part of an expansion to degree and order 8), is listed in Table 7. For the Neptune field model, it is necessary to add a measure of the resolution of each parameter listed in Table 7. For this we use the magnitude of the corresponding diagonal element of the resolution matrix, a measure of the uniqueness of the solution, the extent to which an individual model parameter can be represented by the basis vectors included in the partial solution. Note that some of the quadrupole and octupole coefficients of the O_8 model are not well resolved, but included out of necessity (see discussion of parameter resolution in *Connerney et al.* [1991]).

< Figure 24 near here >

< Table 7 near here >

As anticipated, the relative magnitude of the quadrupole is quite large; the octupole is large as well, but can only be loosely constrained by the Voyager observations. The addition of a smoothness constraint [*Holme and Bloxham, 1996*] allows one to more confidently estimate the octupole coefficients, again resulting in a model that agrees well with the parameters in Connerney et al.'s [1991] model that are well resolved.

The magnetic field magnitude on the surface of the planet ranges from approximately 0.9 G at southern mid-latitudes to ≤0.1 G near the equator and at northern mid-latitudes (Figure 25). Figure 25 also shows estimated auroral zones (stippled regions) which differ greatly in size as a result of the hemispherical asymmetry in field magnitudes. In this representation, it must be recognized that only the low degree and order part of the field is portrayed on the surface of the planet, and the likely presence of

higher degree and order terms may be expected to significantly alter the surface field. It is clear that the magnetic fields of Uranus and Neptune are quite similar in the range of surface field magnitudes, the tilt of the dipole, and the appearance of one strong pole at southern mid latitudes. Both Uranus and Neptune have extended weak-field regions, multiple-dip equators, and multipolar structure in evidence at the surface of the planet.

< Figure 25 near here >

Heading One: Satellites and Small Bodies

Heading Two: Moon

The Moon, with a radius of 1738 km and a mass of little more than 1 percent that of the Earth (m/M_e = 0.0123), is the largest satellite among the terrestrial planets. It is not so much smaller than Mercury (2439 km) in size, with a little more than 1/3rd the volume of Mercury, but only about 1/5th of the mass. It is a differentiated object similar in composition to the Earth's mantle.

The Moon has been a frequent target of space exploration spanning nearly half a century, beginning with a series of robotic probes in the 1960's (Ranger series of the US space program, Luna and Zond series of the USSR program) and culminating in the first missions of human exploration of the Moon, the US Apollo Program. For a brief moment in history, beginning with Apollo 11 on July 20, 1969, and ending with Apollo 17 on December 14, 1972, humans walked (and drove) on the surface of another world, deploying instruments and collecting soil and rock samples for return to Earth. Much of what we know of lunar magnetism evolved from study of the lunar samples returned from the Moon, combined with in situ observations collected both on the surface and in orbit about the Moon. Robotic scientific exploration of the Moon was rekindled decades after the Apollo Program ended, beginning with the orbiting Clementine spacecraft in February and March of 1994. The Lunar Prospector (LP) spacecraft was in polar mapping orbit from January 13, 1998 through July 31, 1999, mapping crustal magnetic fields via both direct (magnetometry) and indirect (electron reflectometry) methods. Initially LP operated in polar orbit at a nominal 100 km mapping altitude, followed by an extended mission (January – July of 1999) at reduced altitudes of between 15 and 45 km. ESA's SMART-1 spacecraft was in orbit about the Moon from November 15, 2004 through impact on September 3, 2006. In October, 2007, Japan's SELENE spacecraft was inserted into a ~100 km altitude, polar mapping orbit to map crustal magnetic anomalies, characterize the lunar plasma environment, and probe its electrical conductivity structure [Tsunakawa et a., 2010]. SELENE was a product of the Institute of Space and Astronautical Science (ISAS) and the National Space Development Agency (NASDA), both now part of the Japan Aerospace Exploration

Agency (JAXA). After 20 months in orbit, SELENE impacted the lunar surface near the crater Gill, having completed its mapping mission.

The first complete lunar magnetic field map (Figure 26) to utilize the electron reflectometry method was compiled using 1.5×10^5 estimates of the surface field magnitude obtained over Lunar Prospector's 18 month mission [*Mitchell et al.,* 2008]. This map extends the electron reflection technique, correcting for charging of the lunar surface and spacecraft [*Halekas et al.*, 2002], significantly improving the dynamic range of field magnitudes mapped relative to previous maps [*Halekas et al.* 2001]. The surface field magnitude is estimated with a resolution of about 0.2 nT in magnitude. This map is still undersampled, in the sense that substantially more observations would be required to obtain complete spatial coverage at the resolution implied by the electron gyroradius (about 5 km at 300 eV). However, it provides a good sense of the global distribution of magnetic sources in the lunar crust and it demonstrates that the characteristic scale length of coherent magnetization in the lunar crust is quite small. Where the lunar crust is magnetized, it is weakly magnetized (on spatial scales of 5 km and greater), compared to either the Mars crust, or Earth's. Lunar crustal magnetization also appears to be less coherent at large spatial scales, compared to Earth and Mars. The late, large impact basins (e.g., Orientale, Imbrium) appear non-magnetic, implying that the impact melt cooled in zero field (no lunar dynamo at the time of impact); and yet, the regions of strongest field magnitude appear *antipodal* to the largest impact basins [*Lin et al.,* 1988].

Figure 26 near here >

So it seems that lunar crustal magnetization and large impacts are related; perhaps impacts magnetize the crust not at the impact site, but on the opposite side of the Moon. One theory for how this might occur [*Hood and Vickery*, 1984; *Hood and Huang*, 1991] uses the impact-generated plasma cloud to intensify the weak ambient magnetic field at the antipode, if only briefly (~ 1 day), as it expands from the impact site to envelop the Moon. In this theory, the expanding plasma cloud, being a good electrical conductor, sweeps up the ambient field as it converges on the antipode, where the crust then acquires remanence via shock magnetization [*Fuller et al*, 1974; *Cisowski et al.,*1983]. The efficacy of this process (the acquisition of remanence via shock in the presence of an applied magnetic field) has been demonstrated in shock experiments performed on lunar mare basalts [*Gattacceca et al.*, 2010]. Shock remanent magnetism (SRM) may be acquired by typical lunar rocks (dominated by multidomain iron-nickel mineralogy with low nickel content) subjected to low pressure shocks (<10 GPa) attributed to lunar impacts. A detailed study of the magnetic field associated with multi-ring impact basins, estimated using the electron reflectometry technique, finds an extensive magnetic low of ~1.5 – 2 basin radii associated with each basin, attributed to shock demagnetization of the impact area [*Halekas et al.*, 2003]. A subset of these impact basins formed in the early Nectarian era (~3.9 Ga) also evidence a more compact, central magnetic

high that is attributed to thermal remanence (TRM) in an ambient field, i.e., evidence for a lunar dynamo operating at that time. The lack of such features in younger (Imbrian) impact basins (~3.2 – 3.85 Ga) and older (Pre-Nectarian) impact basins (>3.92 Ga) suggests that the lunar dynamo was active only briefly, for perhaps a few hundred million years during the period of time spanned by the (large) impact record. A detailed study of LP magnetometer observations acquired over these same basins at low altitude (~25-40 km) confirms the presence of central features in these Nectarian aged basins [*Hood*, 2011]. It is worthwhile to note that these features are very weak – just a few tenths of a nT over most of the lunar surface – at satellite altitude. The most intense magnetic anomalies measured directly by the LP magnetometer range from a few to ~25 nT in magnitude [*Hood et al.*, 2001; *Richmond and Hood*, 2008], sensed at ~15 to 20 km above the lunar surface. The most extensive magnetic anomalies are mapped antipodal to the Imbrium, Crisium, and Serenitatis basins; and over Rima Sirsalis and the Reiner Gamma Formation. The ER observations sense (surface) field magnitudes that are about an order of magnitude greater but even these estimates may be dramatically biased, i.e., underestimated [*Halekas et al.*, 2010], particularly if the anomalies sensed are of smaller spatial scale (<10 km) than sensed by the ~200 eV electrons utilized by the ER technique.

< Figure 27 near here >

The loose association of lunar magnetic anomalies with impact site antipodes [*Lin et al.*, 1988] has been the most perplexing aspect of lunar magnetization, in part due to the somewhat exotic nature of the proposed mechanism by which an impact on one side of the Moon magnetizes the crust on the opposite side [*Hood and Vickery*, 1984; *Hood and Huang*, 1991]. Wieczorek et al [2012] have proposed instead that the global distribution of lunar crustal magnetism is better understood as a consequence of the giant impact event that formed the largest and oldest impact crater on the Moon: that of the South Pole – Aitken basin. They modeled the oblique impact of an ~200 km diameter object, and find a distribution of impactor mass emplaced on the lunar surface that resembles the distribution of the most intense lunar magnetic anomalies. They propose that the magnetized material in the lunar crust was delivered by this differentiated impactor, and acquired remanence by cooling in the presence of the Moon's early dynamo field. This proposal neatly circumvents the problem posed by the paltry thermoremanence susceptibility of typical lunar materials, substituting material from the impacting projectile with orders of magnitude greater thermoremanent susceptibility [*Wieczorek et al.,* 2012]. It is also consistent with the presence of widely scattered and isolated anomalies of small spatial scale, downrange of the South Pole – Aitken basin and elsewhere.

Complete global maps of the lunar magnetic field were also compiled using direct measurements obtained by the fluxgate magnetometer [*Raymond and Hood,* 2008; *Purucker,* 2008; *Purucker and Nicholas*, 2010] at a nominal altitude of ~30 km provided by LP's extended mission. These maps (Figure

27) agree well with each other and with the map of field magnitudes derived from ER measurements if one allows for the differences in mapping altitudes. All evidence, to some degree, the broad weak field associated with Imbrium on the near side, and the relatively strong crustal fields northwest of the South Pole-Aitken basin on the far side. More subtle features associated with other impact basins can also be found, particularly if one uses the ER map (top) as a guide in interpretation of the others. Tsunakawa et al. have applied a regional mapping method to map lunar surface magnetic fields [*Tsunakawa et al.*, 2014], using both LP and SELENE (Kaguya) observations obtained in low lunar orbit (<50 km altitude above the South Pole-Aitken Basin). They obtain a rather remarkable correspondence between fields mapped to the lunar surface using the two (independent) data sets, for both weak and strong field regions alike. Weak fields are characterized by surface field magnitudes of a few nT, whereas the strong field region to the northwest of the South Pole-Aitken Basin is characterized by small scale features with surface fields of ~200 nT. Both regions evidence elongated features, most apparent in the vector components of the mapped field, suggesting lineated source material.

The moon's crustal magnetism can be understood as a result of impact processes resulting in both SRM and TRM, with rather weak magnetization intensities of a few tenths to a few A/m [*Hood*, 2011; *Purucker and Nicholas*, 2010] and with spatial scales at or below the mapping resolution currently available (~5 km). Surface field magnitudes of a few hundred nT have been detected by the ER method but Halekas et al. [2010] suggest that fields of order 1000 nT might be measured at the surface (with better resolution). One measure of the relative intensity of crustal magnetization is provided by the Lowe's spectrum (Figure 28), Rn, which is equal to the mean-squared magnetic field amplitude attributed to harmonics of degree n. At all spatial scales measured, and modeled with the aid of spherical harmonics, the lunar crustal magnetic field is about an order of magnitude weaker than that of Earth (n>14) and two orders of magnitude weaker than that of Mars. This comparison likely underestimates the great differences in crustal magnetization among these bodies, however. The crustal fields of the moon and Mars are due to remanent magnetism, whereas the earth's crustal field is at least partly, if not mostly, induced. The Lunar magnetic anomalies appear to be due to discrete magnetic sources scattered in an otherwise non-magnetic (or very weakly magnetized) crust, whereas the Mars crust must be broadly magnetized throughout, but for regions characterized by large impacts or volcanic emplacements.

< Figure 28 near here >

Heading Two: Ganymede

Ganymede, with a radius of 2634 km, is the largest satellite in the solar system. It is slightly larger than the smallest planet in our solar system, Mercury (R_m= 2439 km), but only about half as massive (M_g=0.025 M_e). Ganymede may have an Fe or FeS core extending to as much as half the satellite's radius,

overlain by a silicate mantle and topped with ~800 km layer of ice [*Schubert et al.,* 1996; *Anderson et al.,* 1996].

The Galileo spacecraft was targeted for several close flybys of Ganymede, the first of which occurred in June of 1996. The first two flyby encounters, with minimum altitudes of 838 and 264 km, demonstrated conclusively the presence of an intrinsic dipolar magnetic field and associated magnetosphere [*Kivelson et al*, 1996; *Gurnett et al*, 1996, *Kivelson et al,* 1997]. Initial analysis of the magnetometer observations revealed a dipole field with equatorial surface field of 750 nT, nearly (within 10 degrees) aligned with the rotation axis. A field of this magnitude is sufficient to produce a Ganymede "mini-magnetosphere" with all of the characteristics normally associated with a trapped particle population. Ganymede's immediate environment (Figure 29) includes field lines with both ends anchored within Ganymede, hosting a local trapped radiation environment; "open" field lines that extend to the Jovian ionosphere; and Jovian field lines. A more comprehensive analysis, performed after the discovery of induced magnetic fields associated with Europa and Callisto [*Neubauer,* 1998; *Khurana et al.,* 1998; *Kivelson et al.,* 1999, *Zimmer et al,* 2000], allowed for the presence of an induced field in addition to the main field [*Kivelson et al.,* 2002]. This analysis demonstrated that Ganymede has both: an inductive response to the time-varying magnetic field, due to currents induced in a hidden global ocean, and an intrinsic field likely due to a dynamo in the deep interior. The intrinsic field is a permanent dipole with a surface equatorial field of 719 nT, tilted by 176° with respect to the rotation axis and rotated 24° toward the trailing hemisphere from the Jupiter-facing meridian plane. The permanent field is perhaps the best evidence that Ganymede has a substantial Fe or FeS core in the interior [*Schubert et al.,* 1996]; dynamo action implies that it is fluid, and convecting, with sufficient energy input to sustain the dynamo. Zhan and Schubert [2012] constructed Ganymede dynamo models in which inner core convection is driven by Fe-snow and FeS flotation, in addition to solid inner core growth. They find that the Fe-snow mechanism is an unlikely candidate for powering Ganymede's dynamo, as it results in a magnetic field that is both weaker and more complex (multipolar) than is observed. Thus, core convection is mainly driven by FeS flotation or inner core growth, both of which concentrate the bouyancy flux near the inner core boundary, where it is most effective [*Zhan and Schubert,* 2012]. However, it is possible (but not likely) that Ganymede's dipolar magnetic field is due to remanent magnetism [*Crary and Bagenal*, 1998].

< Figure 29 near here >

Heading Two: Io

Io, with a radius of 1821 km, is the most geologically active satellite in the solar system. It is substantially smaller than the smallest magnetic planet in our solar system, Mercury (R_m= 2439 km), and less than half as massive as the largest satellite, Ganymede (M_{io} = 0.39 M_g). It is quite similar in size and

mass to the Moon. Yet it is the most volcanic body in the solar system, evidencing extensive and ongoing volcanic activity owing to tidal heating. Io is a differentiated body with an Fe or FeS core extending to as much as one third or one half the satellite's radius, overlain by a silicate mantle and topped with a young, pockmarked crust a few tens of km thick [*Anderson et al.,* 2001].

The Galileo spacecraft executed a number of close flybys of Io, probing fields in Io's immediate vicinity when it was both above and below the magnetodisc current sheet, where it experiences the maximum inducing field. Io, like the icy satellites Europa, Ganymede, and Callisto, evidences an inductive response to the time-varying magnetic field, due to currents flowing in an electrically conductive interior [*Khurana et al.,* 2011]. But in contrast to the icy satellites, Io's currents flow in a global subsurface magma layer just beneath the crust, rock heated to ~1200º C by tidal heating caused by its orbital resonance with Europa and Ganymede [*Peale et al.,* 1979]. Khurana et al. [2011] considered the possibility that Io (like Ganymede) sustained a weak dynamo in addition to an inductive response, finding an upper limit of ~100 nT for a surface (equatorial) field, about 20% of the magnitude of the maximum inductive response. They conclude that if Io has a permanent dipole field, it is extremely weak. Why would Io, with a fluid, electrically conductive Fe or FeS core, and prodigious heat flow, not sustain an active dynamo? A likely explanation has to do with where the heat originates. Tidal heating deposits heat just beneath the crust, inhibiting transport of heat from the deep interior [*Weinbruch and Spohn*, 1995; *Schubert and Soderlund*, 2011]. Convection in the core (and an active dynamo) cannot be maintained absent a temperature differential across the core-mantle boundary.

Heading One: Discussion

A convenient measure of the complexity of a planetary magnetic field, often used in studies of the magnetic field of the earth and planets, is the "harmonic spectrum," sometimes referred to as a "Lowes spectrum," defined as follows [*Lowes,* 1974; *Langel and Estes,* 1982]:

$$ R_n = (n+1) \sum_{m=0}^{n} \left\{ \left(g_n^m \right)^2 + \left(h_n^m \right)^2 \right\} $$

This quantity is equal to the mean squared magnetic field intensity over the planet's surface produced by harmonics of degree *n*. Scaled to the core-mantle boundary with the factor $(a/r_c)^{2n+4}$, this quantity represents the mean squared magnetic field intensity at the dynamo surface, one of several measures of interest in dynamo theory [e.g., *Radler and Ness*, 1990; *Schultz and Paulikas*, 1990]. The Earth's field is

well known to high degree and order ($N_{max} = 23$). Scaled to the core-mantle boundary, the spectrum becomes almost flat for $n \le 14$, suggesting a "white" spectrum for the dynamo at the core-mantle boundary [e.g., *Lowes,* 1974]. (The Earth's spectrum beyond $n \sim 14$ is attributed to crustal sources). It is assumed that the core-mantle boundary, the location of which is very accurately known, represents the outer boundary of the geodynamo. Thus it has often been assumed (prior to knowledge of Uranus' and Neptune's magnetic fields) that a white spectrum at the core-mantle boundary is a common feature of all planetary dynamos [*Elphic and Russell,* 1978], although in the earth's case the quadrupole is considerably less than expected. In Figure 30, R_n is calculated using the GSFC 12/83 model for the Earth [*Langel and Estes,* 1985], the Z_3 model for Saturn [*Connerney et al.*, 1982*b*], and the Q_3 model for the magnetic field of Uranus [*Connerney et al.,* 1987]. For Jupiter both the O_4 model [*Acuña and Ness,* 1976] and O_6 models [*Connerney,* 1992] are shown, illustrating a range of possible values for the quadrupole and octupole. For Uranus, no estimate of the magnitude of higher degree *(n > 2)* moments can be made, from magnetic measurements alone (but see *Holme and Bloxham, 1996* and *Herbert, 2009*) while for Neptune, terms of higher degree are appreciable, but quite uncertain.

< Figure 30 near here >

For Jupiter and Saturn, we assume that the dynamo core radius is near the radius of the metallically conducting hydrogen core, approximately 0.75 and 0.5 planet radius, respectively [e.g., *Stevenson,* 1983]. The transition from molecular hydrogen to the metallic state is likely to be a gradual transition, and for Jupiter, especially, the molecular hydrogen envelope at depth is likely to be electrically conductive and capable of dynamo generation [*Hide*, 1967; *Smoluchowski*, 1975; *Stanley and Glatzmaier*, 2010]. So Jupiter's dynamo may extend to ~0.85 Rj or so. For Uranus and Neptune, we take $r_c = 0.75\ R_p$ with the expectation that the dynamo for these planets operates in the ice mantle [*Hubbard and MacFarlane*, 1980; *Hubbard and Marley*, 1989; *Podolak and Reynolds*, 1981; *Podolak et al*., 1991], where ionic conduction prevails [*Smoluchowski and Torbett*, 1981]. The pressure-induced ionization of H_2O provides sufficient electrical conductivity for dynamo generation [*Mitchell and Nellis*, 1982; *Nellis et al*., 1988] though it is orders of magnitude less conductive than an iron-rich terrestrial core or a metallically conducting hydrogen core.

In this presentation, the planets fall into two distinct groups, one in which the core surface field is dominated by the dipole (Earth, Jupiter, Saturn) and one in which the core surface field is dominated by higher-degree contributions (Uranus and Neptune). However, in recognition of the peculiarity of Saturn's axisymmetric field, one might better think in terms of three classifications. In *Stevenson's* [1982] model, differential rotation of an outer conductive shell attenuates the nonaxisymmetric components of the dynamo within, thereby reducing the observed tilt by an order of magnitude or more from the "nominal"

value. This model has an appealing physical basis (material properties of the H-He fluid) and appears to satisfy other constraints, e.g., helium depletion of the gaseous outer envelope and a relatively high planetary heat flux. The dipole-dominated group includes the planets with near alignment of dipole and rotation axes (≤10); the other includes the oblique rotators Uranus (59°) and Neptune (47°), the planets with a very large angular separation of the dipole and rotation axes. Table 8 provides a comparison of the gross characteristics of the magnetic fields of the planets and satellites with active dynamos, using the eccentic dipole representation [*Connerney,* 1993]. The eccentric dipole is computed from a spherical harmonic model and provides a useful characterization of the field at a distance, the dipole magnitude, tilt and offset. Estimated minimum and maximum surface fields are computed using spherical harmonic representations, taking into account dynamical flattening, which is most significant for the gas giants.

< Table 8 near here >

Heading One: Summary

The magnetic planets (and satellites) of the solar system provide a rich diversity of objects of relevance to dynamo theory. Mercury, Mars, Earth, and Ganymede have substantial iron cores with varying amounts of S or some other alloying element; each has, or once had, an active dynamo. In this group Mars and Earth appear similar, although the Mars dynamo is no longer active, and we have only the crustal remanence that remains today as evidence of its dynamo. Mercury and Ganymede are similar in size as well as magnetic properties; both have diminutive dynamos often considered so weak that alternatives to an active dynamo are proposed [*Giampieri and Balogh,* 2002; *Aharonson et al.,* 2004; *Crary and Bagenal*, 1998]. Mercury's weak field has been attributed to a dynamo operating beneath a stably stratified and electrically conductive shell [*Christensen,* 2006] that serves to attenuate the field, or, alternately, a thin shell dynamo operating above a solid and electrically conductive inner core [*Stanley et al.,* 2005]. Both Mercury and Ganymede have dipolar, or nearly dipolar fields more or less aligned with the spin axis.

Unmagnetized Venus reminds us that in dynamo theory, the "devil is in the details".

In Jupiter and Saturn, we have two planets of nearly the same composition and comparable size but with magnetic fields that are worlds apart. Jupiter's dynamo produces a magnetic field resembling Earth's, while Saturn's is uniquely spin-axisymmetric. Stevenson's account of Saturn's internal structure and magnetic field is appealing, as it provides a vehicle for understanding the differences between these gas giant planets. In Uranus and Neptune we have again two planets of similar composition and size, this time with similar magnetic fields (fortunately!). These icy twins form a category of "oblique rotators"

with large dipole tilts and eccentric magnetic fields. Dynamo generation within these planets takes place in the relatively poorly conducting "ice" mantle that sits atop a stably stratified and electrically conductive "super-ionic" water interior [*Redmer et al.,* 2011]. Dynamos operating in spherical shells can produce dipole fields with large dipole tilts [*Aubert and Wicht*, 2004] and the harmonic content characteristic of ice giant dynamos [*Stanley and Bloxham*, 2006], but none thus far mimic the hemispheric asymmetry observed of such planetary fields. Heuristically, it is tempting to associate dipole dominance with small dipole tilt and dipole deficiency with large dipole tilt; it appears that dynamos operating in less conductive fluids are in the latter group. If dynamos operate with roughly comparable magnetic Reynolds numbers and convective velocities, those operating in poorly conductive fluids will require relatively large spatial scale lengths [*Connerney et al.*, 1987; *Connerney*, 1993]. If so, an ice giant dynamo, characterized by a relatively small number of large convective cells, might be expected to have a large dipole tilt [e.g., *Parker*, 1969].

In little more than three decades, we've sampled the magnetic fields of every planet in the solar system, some several times over. This diverse sample set challenges our understanding of the generation of magnetic fields in planetary interiors. Dynamos are ubiquitous and evidently remarkably responsive to subtle differences in interior composition, heat flow, and convection, and as such offer clues to a planet's interior structure and evolution. For the present, dynamo theory matures on the heels of such discoveries [*Schubert and Soderlund*, 2011; *Stanley and Glatzmaier*, 2010]; for the near future, we may all look forward to the global mapping of Jupiter's magnetic field in the coming years, and that of Saturn, Mercury, and Ganymede as well. Remote Uranus and Neptune may hold onto their secrets a bit longer.

Figure Captions:

Figure 1: Mariner 10 trajectory in Mercury centered cylindrical coordinates for the first (M1) and third (M3) encounters. The spacecraft distance from the equator (z) as a function of distance from the z axis (ρ) is given in units of Mercury radius. The part of the trajectory within the Hermean magnetosphere is indicated by the thick line segment.

Figure 2: M1 encounter trajectory and magnetic field observations in Mercury ecliptic coordinates, viewed from the sun (top) and viewed along Mercury's rotation axis (bottom). The vector magnetic field observed along the Mariner 10 trajectory is illustrated by vector projection onto the z-y (top) and x-y (bottom) planes; vectors originate at the spacecraft position in ME coordinates at the time of observation. Observed bow shock (BS) and magnetopause (MP) positions are shown along with scaled nominal bow shock and magnetopause boundaries.

Figure 3: M3 encounter trajectory and magnetic field observations in Mercury ecliptic coordinates, viewed from the sun (top) and viewed along Mercury's rotation axis (bottom). The vector magnetic field observed along the Mariner 10 trajectory is illustrated by vector projection onto the z-y (top), x-z

(middle), and x-y (bottom) planes; vectors originate at the spacecraft position in ME coordinates at the time of observation. Observed magnetopause (MP) positions are shown along with scaled nominal bow shock and magnetopause boundaries.

Figure 4: MESSENGER orbits about Mercury slowly evolve in local time (dawn-dusk at left; noon-midnight at right) as Mercury orbits the Sun. Closest approach to Mercury at altitudes of ~200 to ~500 km occurs at high northern latitudes (60º – 74º).

Figure 5: Magnetic field lines of a model Hermean magnetosphere (blue) in meridian plane projection (x-z plane) compared with the vector magnetic field observed along the M1 passage through the magnetotail.

Figure 6: Coplanar polar orbits of the BepiColombo Mission spacecraft. The three axis stabilized Mercury Planetary Orbiter (MPO) will complete an orbit about Mercury in about 2.3 hours, while the Mercury Magnetospheric Orbiter (MMO), a spin-stabilized spacecraft, will have an orbit period of about 9.3 hours; both spacecraft will have a closest approach altitude of 400 km. Launch is scheduled in July, 2016.

Figure 7: Mars Global Surveyor trajectory and magnetic field observations in an orbit plane projection for day 264, 1997. The vector magnetic field observed along the MGS trajectory is illustrated by vector projection onto the orbit plane; vectors originate at the spacecraft position at the time of observation. From Connerney, J. E. P., Acuña, M. H., Ness, N. F., Spohn, T., and Schubert, G. (2004). Mars crustal magnetism. *Space Science Reviews* **111**(1-2): 1-32.

Figure 8: Global distribution of the most intense magnetic sources in the Mars crust. Magnitude of the radial magnetic field measured during the Mars Global Surveyor aerobraking passes below 200 km altitude (to ~ 100 km minimum altitude). Most intense sources are found in the heavily cratered southern highlands, south of and near the dichotomy boundary (thin line). From Acuña, M. H., Connerney, J. E. P., Ness, N. F., et al. (1999). Global distribution of crustal magnetism discovered by the Mars Global Surveyor MAG/ER Experiment. *Science* **284**: 790-793.

Figure 9: Map of the magnetic field of Mars observed by the Mars Global Surveyor satellite at a nominal 400 km (mapping) altitude. Each pixel is colored according to the median value of the filtered radial magnetic field component observed within the 1º by 1º degree latitude/longitude range represented by the pixel. Colors are assigned in 12 steps spanning two orders of magnitude variation. Where the field falls below the minimum value a shaded MOLA topography relief map [*Smith et al,* 1999] and contours of constant elevation (-4, -2, 0, 2, 4 km elevation) are shown. From Connerney, J. E. P., Acuña, M. H., Ness, N. F., Kletetschka, G. *et al.,* (2005). Tectonic implications of Mars crustal magnetism. *Proceedings of the National Academy of Sciences* **102**(42): 14970-14975.

Figure 10: Vector magnetic field measured by MGS during aerobraking on calendar day 20, 1999. Periapsis occurred at 68.0º S, 181.2º W and 106 km altitude. The x (north) and z (down) components of the vector field sampled every 3 seconds (crosses) are compared with a model fit (solid line). Altitude variation in kilometers indicated with a dashed line. Observations plotted as a function of distance x north and south of an origin at 53º S and at the longitude of periapsis. Model consists of 20 uniformly magnetized slabs aligned with the y axis (east-west) and infinite in extent along y. The x and z components of the model crustal magnetization per unit volume (A/m) are indicated in the bar graphs between the two panels. Figure Connerney, J. E. P., Acuña, M. H., Wasilewski, P. J., Ness, N. F., *et al.,* (1999). Magnetic lineations in the ancient crust of Mars, *Science* **284**: 794-798.

Figure 11: Spacecraft flybys (Pioneer 10 & 11, Voyager 1, and Ulysses) of Jupiter in a magnetic

equatorial (cylindrical) coordinate system. The z axis is aligned with Jupiter's magnetic dipole axis. The spacecraft distance from the magnetic equator (z) as a function of distance from the z axis (ρ) is given in units of Jupiter radius. The Galilean satellites Io, Europa, and Ganymede trace out arcs (not shown) in this coordinate system crossing the equator at the radial distances indicated. The shaded region approximates the washer-shaped system of azimuthal ring currents that gives rise to the magnetodisc configuration of the Jovian magnetosphere.

Figure 12: Orthographic projection of the near Jupiter ($r < 5R_j$) portion of Juno's 11-day science orbits. Orbits 2-16 (blue) are acquired in the first six months of mapping and are equally spaced in longitude (at closest approach) by 24º. Orbits 17-31 (black) are acquired after a trajectory correction that places these passes between those already acquired, resulting in complete global coverage with longitude spacing of ~12º. Juno's insertion into orbit about Jupiter (JOI) occurs in July, 2016.

Figure 13: Relative harmonic content (Lowe's spectrum) of spherical harmonic models of the Jovian field, compared with that of Earth and Saturn.

Figure 14: Image of Jupiter at 3.4 μm obtained at NASA's IRTF at Mauna Kea, Hawaii, using the NSFCAM facility imager. At this wavelength, bright H_3^+ auroral emissions originating above the homopause appear agains a planetary disc darkened by methane absorption. Magnetic field lines are drawn in meridian plane projection to illustrate where field lines crossing the equator at 30 R_j and 6 R_j (Io's orbit) intersect Jupiter's surface. A bright emission feature at the foot of the Io Flux Tube (IFT) can be identified near the dawn limb, as well as faint emission extending along the Io L-shell downstream of the instantaneous IFT. From Connerney, J. E. P. and Satoh, T. (2000). The H_3+ ion: A remote diagnostic of the Jovian magnetosphere. *Philosophical Transactions of the Royal Society of London* **358**: 2471-2483.

Figure 15: Contours of constant magnetic field magnitude (Gauss) on the dynamically flattened (1/15.4) surface of Jupiter, computed using the VIT4 model field (see text). The top panel shows the field magnitude in orthographic projections viewed from the north (left) and south (right); below the colorbar is a rectangular latitude-longitude (System 3 west longitude) projection. A trace (dashed) indicates the position of the magnetic equator. Solid lines join points on the surface that trace along field lines to the orbits of the Galilean satellites Io, Europa, and Ganymede, with filled circles for increments of 30 º in the satellite's System 3 longitude.

Figure 16: Vector magnetic field observed during Cassini Saturn Orbit Insertion (SOI) on day 183, 2004. The observed field is dominated by the smooth variation of the radial (red) and theta (black) components of the field. The azimuthal component of the magnetic field (blue) is very small throughout the equatorial passage, and is plotted again increased by a factor of 100 and offset by 10 nT for clarity.

Figure 17: Vector residual magnetic field during the Cassini Saturn Orbit Insertion on day 183, 2004. The best fitting zonal harmonic (axisymmetric) model of Saturn's planetary magnetic field has been removed from the observations, and the components of the residual field plotted as a function of time. The radial and theta components are offset (by 10 and 20 nT, respectively) for clarity. Significant spacecraft maneuvers and events are indicated. Systematic differences between the best fitting zonal harmonic model and the observed field are evident; the residuals in all three components are clearly associated with spacecraft turns and propulsive events.

Figure 18: Spacecraft flybys (Pioneer 11, Voyager 1 & 2, and Cassini) of Saturn in a cylindrical coordinate system in which the z axis is aligned with Saturn's magnetic dipole axis and rotation axis. The spacecraft distance from the equator plane (z) as a function of distance from the z axis (ρ) is given in units

of Saturn radius. The positions of satellites Mimas, Enceladus, Tethys, Dione, and Rhea are indicated. The shaded region approximates the innermost portion of a washer-shaped region of azimuthal ring currents.

Figure 19: Magnetic field magnitude (Gauss) as a function of planetocentric latitude on the dynamically flattened (1/10.6) surface of Saturn computed using the axisymmetric Z_3 (zonal harmonic of degree 3) model of Saturn's magnetic field [*Connerney et al*, 1982b], compared with that of a simple offset dipole.

Figure 20: Unsigned magnetic flux computed on a sphere of radius 0.5 R_s (presumed dynamo core radius) using the Z_3 magnetic field model, compared to that for a simple dipole. Saturn's magnetic field is very nearly identical to that which (with 3 terms) minimizes the low latitude (<45º) unsigned magnetic flux across the core boundary. From Connerney, J. E. P. (1993). Magnetic fields of the outer planets. *Journal of Geophysical Research* **98**(E10): 18659-18679.

Figure 21: Saturn's magnetic field is very nearly symmetric about its rotation axis but not about the equator. All field lines passing through the ionosphere at one latitude cross the ring plane at a unique radial distance (ring plane conjugate) and map to a corresponding (but different) latitude in the opposite hemisphere. The magnetic equator is at 5.9º N (planetocentric) latitude [*Connerney,* 1986].

Figure 22: Voyager 2 spacecraft flybys of Uranus and Neptune in a cylindrical coordinate system. The spacecraft distance from the equator plane (z) as a function of distance from the rotation axis (ρ) is given in units of planet radius. The positions of satellites Miranda, Ariel, and Umbriel in the Uranus system are indicated.

Figure 23: Contours of constant magnetic field magnitude (Gauss) on the dynamically flattened (1/41.6) surface of Uranus, computed using the Q_3 spherical harmonic model (see text). The top panel shows the field magnitude in orthographic projections viewed from +z (left) and -z (right); below the colorbar is a rectangular latitude-longitude (west longitude) projection. A trace (dashed) indicates the position of the magnetic equator. Solid lines join points on the surface that trace along field lines to the orbit of the satellite Miranda, with filled circles for increments of 30 º in the satellite's longitude.

Figure 24: Magnitude of the observed magnetic field (crosses) as a function of time (lower axis) and spacecraft radial distance (upper axis) for the Voyager 2 encounter with Neptune, compared with a model fit (O_8 model) and the relative magnitudes of the dipole (dashed), quadrupole (solid) and octupole (dot-dashed) terms in the spherical harmonic expansion.

Figure 25: Contours of constant magnetic field magnitude (Gauss) on the dynamically flattened (1/47.6) surface of Neptune, computed using the O_8 spherical harmonic model (see text). The top panel shows the field magnitude in orthographic projections viewed from +z (left) and -z (right); below the colorbar is a rectangular latitude-longitude (west longitude) projection. A trace (dashed) indicates the position of the magnetic equator.

Figure 26: Magnetic field magnitude on the lunar surface as measured by the electron reflection technique. Field magnitudes from 0.2 to 250 nT are represented by colors according to the colorbar to the left. The top panel shows the field magnitude in orthographic projections viewed from +z (left) and -z (right); below is a rectangular latitude-longitude projection. The white circles denote major impact basins, and the black circles are drawn at the antipodal site. From Mitchell, D. L., Halekas, J. S., Lin, R. P., *et al.* (2008). Global mapping of lunar crustal magnetic fields by Lunar Prospector. *Icarus* **194**: 401-409.

Figure 27: Lambert equal area projections (near side: left, far side: right) of the magnitude of the magnetic field above the lunar surface measured by the Electron Reflectometer (ER) and Magnetometer

instruments on Lunar Prospector (LP). The ER map (top) shows estimated field magnitudes at the lunar surface, remotely sensed via the electron reflection technique (smoothed over 3º x 3º in latitude and longitude). The LP magnetometer observations are represented in two forms. The middle set is color coded to the magnitude of the magnetic field observed at ~30 km altitude after applying a correction for external fields (*Richmond and Hood*, 2008). The bottom set represents the field magnitude (also at 30 km altitude) derived from a 170 degree spherical harmonic model fitted to LP magnetometer data (*Purucker and Nicholas*, 2010). All measurements were acquired by LP at low altitude and primarily on the Moon's nightside, including passages through the Earth's magnetotail, when magnetically quiet conditions were experienced.

Figure 28: Relative harmonic content (Lowe's spectrum) of the magnetic fields of Earth, Moon, and Mars. The Earth's (internal) field at low degree is produced by the dynamo but above n = 14 crustal magnetic fields dominate the spectrum. At all spatial scales, the lunar crust is very weakly magnetized, compared to Earth and (especially) Mars.

Figure 29: Magnetic field of Ganymede and its immediate environment in meridian plane projection. Ganymede's internal magnetic field moment is indicated with a filled arrow. Low latitude field lines close within the satellite's interior; high latitude field lines (blue) form a flux tube with one end rooted in Jupiter's ionosphere; Jovian field lines (black) are separated from those linking Ganymede by a sepratrix (heavy black). Figure adapted from *Kivelson et al*, [2002].

Figure 30: Relative harmonic content (Lowe's spectrum) of spherical harmonic models of Jupiter, Saturn, Uranus, and Neptune, compared with that of earth, normalized to the assumed core radius (in parentheses) for each planet. Two Jovian magnetic field models are chosen to illustrate the range of possibilities for that planet.

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