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Numerical models of the geodynamo and observational constraints

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[1]0 **Abstract:** The past few years have seen the emergence of a large number of numerical simulations of the geodynamo. In parallel, both new and old geomagnetic, archeomagnetic, and paleomagnetic observations have been interpreted as actual geomagnetic features and used as constraints for dynamo models. Naturally, model predictions should be tested against actual characteristics of the geomagnetic field. Despite huge differences (sometimes in excess of a billion) between the values of parameters used in the simulations and those estimated for the Earth, it is intriguing that many available simulations succeed in producing largely axial dipolar magnetic fields with weaker nondipolar structures, in agreement with the first-order characteristics of the geomagnetic field. Yet, when considering finer characteristics, there are significant differences, and failures to actually produce a number of fundamental characteristic features. In this presentation, we first review numerical results obtained to date, then we attempt to summarize which field characteristics derived from observational data sets can be considered robust. On the basis of simple criteria used to evaluate the degree of confidence that can be placed in each datum, we sort presumably characteristic geomagnetic features into three categories (robust, controversial, and unlikely). We conclude that numerical models should be illustrated with a number of key “predictions,” averaged over at least 10 dipole diffusion times. These predictions should be tested against the subset of robust observations only. Controversial observations should await additional confirmation.

Keywords: Geodynamo; Earth core; geomagnetic field; magnetic records.

Index terms: Dynamo theory; time variation-secular and long term; paleomagnetic secular variation.

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1. Introduction

[2]0 After decades of work generally performed in parallel but with rather little interaction, specialists of magnetic field observation, on one hand, and modeling, on the other hand, have come to interact much more recently. Papers coauthored by dynamo specialists and geophysicists observing the field over timescales ranging from real time to geological time have appeared [*Gubbins and Coe*, 1993; *Ultré-Guérard et al.*, 1998; *Glatzmaier et al.*, 1999; *Coe et al.*, 2000]. Acceleration of both theoretical and numerical work has produced a num-

ber of numerical dynamo models that may have some geophysical relevance. Remarkably, most if not all of these numerical simulations share characteristics with the present or past Earth's field as it is now known to us: this includes the dominance of the axial dipolar component, weak nondipolar structures, and, in some cases, full polarity reversal. In order to validate these codes a benchmark comparison is being coordinated by Ulrich Christensen from Göttingen University.

[3] Yet some of the main parameters that control dynamo equations, such as the Ekman number (E) or the magnetic Prandtl number (Pm), that are used in the numerical calculations are extremely far from their estimated values for the Earth. It has even been shown in the case of simplified problems [*Dormy et al.*, 1998] that the numerical values commonly used for modeling do not lead to the equilibria that are appropriate for the Earth.

[4] For instance, the Ekman number is defined as the ratio of the rotation timescale (for the Earth, that is the day) and the viscous diffusion timescale. The day is a terribly short timescale compared to any other relevant timescale that applies to the Earth's core: the Ekman number is believed to be as small as 10^{-15} for the Earth [*Poirier*, 1994; *De Wijs et al.*, 1998]. Yet it is impossible with present computers to attain E values smaller than 10^{-6} for the dynamo problem. This implies that the viscous diffusion term in numerical models is too strong by a coefficient of the order of one billion. Viscous effects decrease very slowly with decreasing values of the Ekman number: as $E^{1/2}$ for boundary layers, as $E^{1/3}$ for axial shear. Thus the Ekman number must be extremely small for viscous effects to be negligible. When viscous effects are restricted to very small regions of shear, it is referred to as the "asymptotic regime" of small E. This regime is relevant to the Earth's core. Ekman numbers that can be achieved in current numerical simulations are not small enough to allow viscous effects to be neglected in the main flow (viscosity actually being used to ensure numerical stability). One may therefore expect the flows computed in numerical models to be of a large scale because of the important viscous dissipation of the small-scale flow.

[5] The magnetic Prandtl number is a measure of the ratio of two other characteristic timescales: the timescale for dissipation of electrical currents with respect to viscous diffusion timescale. The magnetic Prandtl number is estimated at about 10^{-6} for the Earth's core [see *Poirier*, 1994; *Braginsky and Roberts*, 1995] when in practice, it is always chosen to be of the order of or larger than 1 in numerical models (to allow dynamo action). For more Earth-like values, electrical current dissipation and thus field diffusion are very efficient and make it more difficult to achieve self-excited dynamo action. Also, this implies that the field will dominantly be of a large scale (independent of the scale of the flow) in the Earth's core itself as well as at its boundary (an important argument when downward continuing the field at the core-mantle boundary (CMB)).

[6] The numerical values of two essential parameters in dynamo equations are therefore off from their actual Earth values by factors in excess of one million! Yet increasing numbers of authors point out "Earth-like" features in their numerical models. Quoted features include a dominant dipolar axial component, so-called flux patches, westward drift, and in some cases polarity reversal. Most of these characteristics are validated with instantaneous snapshots of the models.

[7] Field observations are becoming simultaneously available at timescales ranging over at least 6 orders of magnitude (from annual to

millions of years) and a more modest spatial scale range (from global to spherical harmonic degree of order 10, i.e., "wavelengths" from 20,000 to 2000 km). However, these "observational constraints" may not be equally robust, and all groups of geomagnetists and paleomagnetists may not agree on their significance. Proper knowledge of all the pitfalls and sources of uncertainties in data acquisition (from magnetic observatory or satellite observations to sampling of rocks for paleomagnetic analysis, laboratory procedures, statistical treatment of raw data, and extraction of descriptive mean field models) is requested before these data can safely be used as constraints and guidelines by theoreticians and dynamo numericists.

[8] In section 2, we briefly summarize published dynamo models and recall the numerical values they use for characteristic parameters and the "Earth-like" features they are reporting. Section 3 is a critical review of field features uncovered from observations, organized in order of increasing timescale. The features we describe are sorted into three categories: robust, controversial, and rejected. In section 4, we confront the class of what we believe to be robust observations versus numerical predictions.

2. Status of Numerical Dynamo Models

[9] Modeling of magnetic induction in the Earth's core is a complex problem that involves the equations of magnetohydrodynamics in a rotating frame. It can be written using various approximations (compressible, Boussinesq, anelastic, etc.), heating, boundary conditions, and nondimensionalizations. In order to give a coherent picture of the models available so far, we have chosen to present them here using a common convention for nondimensionalization (we will use the core radius r_c as the length scale and define the Ekman number as $\nu/\Omega {r_c}^2$, where ν is the kinematic viscosity and Ω the Earth's angular velocity). When presenting these models, we will try to highlight the differences in the parameters they use (see Table 1), as well as in the approximations they apply. We point out in particular any Earth-like behavior they feature. We first introduce a few notions that are necessary to describe these models. The Ekman and magnetic Prandtl numbers have already been introduced. The Prandtl number is the ratio of the thermal versus kinematic diffusive timescales and is of the order of one or less than one in the Earth (between 0.01 and 1; or slightly above one if chemical diffusion is used instead of thermal diffusion). The ratio of the magnetic Prandtl number versus the Prandtl number is defined as the Roberts number; for the Earth's core it is of a magnitude comparable to that of the magnetic Prandtl number (about 10^{-6}). We will concentrate here on convection driven dynamos as this is the case in the papers presented here. It should be noted that other possible sources of energy are proposed for the geodynamos; these include chemical convection (with a shorter diffusive timescales) and precession. The Rayleigh number is a measure of the energy input to the system: the higher its value, the more vigorous motions will be. This parameter is defined using different conventions in numerical works (depending on the heating mode selected and on the normalizing quantity); it is very poorly known as far as the Earth's core is concerned [*Braginsky and Roberts*, 1995], but we know it must be high enough for motions to be able to sustain a magnetic field. Finally, we want to define what is meant by "strong field" and "weak field" dynamos. These notions were first introduced in asymptotic studies, with a meaning that is different from that used in numerical studies. Following *Roberts* [1978], for very rapidly rotating bodies like the Earth's core and when the magnetic Prandtl number is small, there would be two

Table 1. Parameter Values, Model Specifications, and Model Outputs for Numerical Three-Dimensional Dynamos

	Parameters			Model			Outputs				
	E $(\nu/\Omega r^2)$	Pr (ν/κ)	Pm (ν/η)	Equations	Boundary Conditions	Integration Time	Dipolar Field	Λ (F_B/F_Ω)	Reversals	Westward Drift	Others
Glatzmaier and Roberts [1995a, 1995b]	$\ell = 1$, 1.8×10^{-6} $\ell = 10$, 1.3×10^{-4} $\ell = 20$, 1.0×10^{-3}	5000	500	Boussinesq, no inertia	conducting I.C., Ins. M., rigid, with enhanced viscosity near boundaries	$3\tau_d$	yes	500	yes	yes	spectrum, I.C. super-rot.
Glatzmaier and Roberts [1996a]	$\ell = 1$, 1.8×10^{-6} $\ell = 10$, 1.3×10^{-4} $\ell = 20$, 1.0×10^{-3}	5000	500	Boussinesq, axial-inertia	conducting I.C., Ins. mantle, rigid	$0.8\tau_d$	yes		not yet	yes	spectrum, I.C. super-rot.
Glatzmaier and Roberts [1996b] and following	$\ell = 1$, 1.8×10^{-6} $\ell = 10$, 6.5×10^{-5} $\ell = 20$, 5.1×10^{-4}	725	725	anelastic, axial-inertia	conducting I.C., Ins. mantle, rigid	$15\tau_d$	yes	100	yes	yes	spectrum, I.C. super-rot.
Kuang-Bloxham	$\ell = 1$, 4.0×10^{-5} $\ell = 10$, 9.0×10^{-5} $\ell = 20$, 4.9×10^{-4}	1	1	Boussinesq, axial-inertia	conducting I.C., Ins. M.+conducting layer, stress free	$2\tau_d$	yes	$\mathcal{O}(1)$	not yet	yes	flux expulsion, Pacific window, spectrum
Busse et al.	5.1×10^{-4}–7.2×10^{-4}	5–20	100	Boussinesq, full-inertia	Ins. M. & I.C., stress free	$1.4\tau_d$	yes	2–20	not yet		
Kageyama et al.	4.0×10^{-4}	1	10.6–15	compressible, full-inertia	$B \wedge r = 0$, rigid	$50\tau_d$	yes	7×10^{-4}	yes		patches, sawtooth?
Christensen et al.	4.2×10^{-5}–4.2×10^{-4}	1	0.5–5	Boussinesq, full-inertia	Ins. M. & I.C., rigid or stress free	$3\tau_d$ to $10\tau_d$	yes	0.14–14	not yet	yes	patches
Kitauchi-Kida	5.6×10^{-3}	1	8.3–14.2	Boussinesq, full-inertia	vacuum, rigid	$100\tau_d$	yes	60	yes	yes	quasi threefold symmetry
Sakuraba-Kono	$\ell = 1$, 6.3×10^{-5} $\ell = 10$, 7.1×10^{-5} $\ell = 20$, 1.3×10^{-4}	1	20	Boussinesq, full-inertia	conducting I.C., Ins. mantle, rigid	$3\tau_d$	yes	0.1–10	not yet		Archean paleointensity
Katayama et al.	7.2×10^{-3}–7.2×10^{-2}	1	35	Boussinesq, full-inertia	Ins. M. & I.C., stress free	$6\tau_d$	yes	6×10^{-4}	not yet	yes	
Earth	10^{-15}	1/7	10^{-6}					$\mathcal{O}(1)$?			

Each row from the left corresponds to one model as described in a paper or series of papers. The last row recalls parameter estimates for the Earth's core. The first three columns display the parameters described in the text: the Ekman number, the Prandtl number, and the magnetic Prandtl number. The next three columns present the model used in the computations: equations, boundary conditions, and integration times (scaled with respect to the dipole free decay time $\tau_d \simeq 20$ kyr). Finally, the last five columns state the character of the generated field found in each model through the answer to the following questions: "Is the field dipolar?", "What is its strength?" (measured by the Elsasser number, defined as the ratio of the Lorentz force to the Coriolis term), "Did it reverse during the simulations?", "Does it drift westward?", etc. Ins., insulating; I.C., inner core; M, mantle; rot., rotation.

branches of dynamo solutions (i.e., two kinds of solutions): the weak field branch and the strong field branch. Let us start from a non-magnetic state and let the Rayleigh number increase (starting from zero). At first the fluid is at rest, then, beyond a first critical value of this parameter, convection starts (this appearance of a new kind of solution is referred to as a "bifurcation"). The amplitude of convection increases as the Rayleigh number is further increased (the nature of this bifurcation is characterized by the asymptotic study of *Soward* [1977]). Beyond a second critical value, motions become energetic enough to maintain a magnetic field (starting with a perturbation): this is the dynamo bifurcation. This second bifurcation produces a weak field dynamo, in the sense that the amplitude of the field is controlled by viscous effects (though viscosity is very small). If the Rayleigh number is further increased, this branch disappears beyond a third critical value, and the solution quickly evolves to another branch (through runaway growth of the magnetic field intensity). On this new branch, viscous effects are negligible and the amplitude of the field is now controlled by magnetostrophic equilibrium (between Lorentz and Coriolis forces). The nondimensional number measuring the ratio of the Lorentz force to the Coriolis acceleration, namely the Elsasser number, is then close to unity [*Soward*, 1979]. This second solution is referred to as the strong field branch. Given the strength of the measured geomagnetic field, it is clear that the Earth's dynamo corresponds to that kind of solution. The geometry of the solution on that branch should significantly differ from the previous one. In fact, once on that strong field branch, the magnetic field has relaxed the constraints imposed by rapid rotation on the flow, and dynamo action is expected to exist even if the Rayleigh number is now decreased below its first critical value (onset of convection). It is worth noting that such behavior (subcritical dynamo action) has not been observed so far in any numerical models with a spherical geometry.

[10] This whole scenario has been established for small magnetic Prandtl numbers and in the limit of small Ekman numbers or vanishing viscosity (relevant to the Earth), and it is only relevant to that regime. For numerical work one refers to a strong field dynamo if the magnetic field term is on the same order of magnitude, or even much larger than the Coriolis term (the Elsasser number is then equal or larger than unity); this, of course, does not imply that viscosity is negligible (as would be needed with the original definition). Viscosity is never negligible in current three-dimensional numerical models, whose stability always rely on the viscous term. The strong field notion used to discuss numerical models must therefore not be confused with the well-defined one used in the asymptotic regime (see *Childress and Soward* [1972] for a rigorous definition between parallel planes). Indeed, studies of simplified problems strongly suggested that in the asymptotic regime, the Elsasser number should settle to values very close to unity on the strong field branch (see *Eltayeb and Roberts* [1970], *Eltayeb and Kumar* [1977], *Fearn* [1979a, 1979b], and *Soward* [1979], and for a review, see *Soward* [1998]). We will see that such is not always the case for numerically strong field dynamos.

[11] It is important to note here that molecular values were used to evaluate all diffusivities (i.e., viscosity as well as thermal and magnetic diffusivities). It can be argued that small-scale turbulent flow will lead to equivalent large-scale diffusivities that could be several orders of magnitude larger than molecular values. Through a study of the small-scale thermal instability of the core, *Braginsky and Meytlis* [1990] obtained a qualitative picture of the anisotropic turbulent motions and field. They expect that the effects of small-scale turbulence

on the large-scale flow will be highly anisotropic and that the equivalent thermal diffusivity tensor could compare in magnitude with magnetic diffusivity. If this is true and if one neglects the important anisotropy, then the equivalent Roberts number at large scale would be close to one. It is also often argued that vigorous convective models lead to order one values for the equivalent Prandtl number. A simplified argument would then be that all equivalent diffusivities are of the same order, which would imply order one Roberts, Prandtl, and magnetic Prandtl numbers and an Ekman number close to 10^{-9}. The "turbulent" Ekman number evaluated this way would still be several orders of magnitude smaller than can be achieved so far numerically. For ratios of diffusivities (especially the Roberts number and the magnetic Prandtl number), the turbulent estimates are much easier to use than the values based on molecular diffusivities.

[12] This turbulent modification of the parameters can be justified with the qualitative picture described above, but it is not theoretically established. It does not take into account the strong anisotropy of the small-scale flow. Also, experimental studies show that setting a turbulent Prandtl number to unity is an oversimplification, even for nonrotating and nonmagnetic low Prandtl number turbulent convection [*Cioni et al.*, 1997].

[13] Of course, the present goal of numerical simulations cannot yet be to achieve geophysical values of the parameters based on molecular diffusivities! Only the asymptotic behavior, and a well-approximated large-scale solution, constitute a realistic objective. One can then wonder whether this aim can be better achieved using magnetic Prandtl numbers based on either isotropic turbulent diffusivities (i.e., magnetic Prandtl number of order 1) or molecular values (i.e., 10^{-6}). This remains an open question. However, part of the difficulty clearly is to understand how dynamo action is maintained despite the very efficient magnetic diffusivity (compared with other ones). Assuming an order 1 magnetic Prandtl number is a way to get around this issue.

2.1. Zhang and Busse

[14] In 1988 and 1989, Zhang and Busse presented magnetic field generation results in a rotating spherical shell with imposed symmetry with respect to the equator and imposed periodicity in longitude (this is therefore not fully three dimensional). They studied magnetic Prandtl numbers above which they obtained steady or oscillatory dynamos with either quadrupolar or dipolar fields [see also *Zhang et al.*, 1989]. Their approach (later extended by *Hirshing and Busse* [1995]) consisted in following the consecutive bifurcations, starting from the nonmagnetic solution. Being close to the mathematical/physical understanding described above, this approach should allow one to describe solutions on the weak field branch and to safely establish the strong field branch (after runaway growth of the field intensity). This seems to be a safe approach to ensure that parameters different from Earth values indeed lead to the appropriate equilibria (e.g., magnetostrophic balance). The numerical scheme used for resolving the nonlinearity (Galerkin) required moderate values of the Rayleigh number; also in order to study the dynamo bifurcation (starting with arbitrarily small magnetic field), a very high resolution would be needed for the convective regime at high values of the Rayleigh number. When trying to increase the Rayleigh number and/or decrease the Ekman number, authors noted [*Zhang et al.*, 1989] that magnetic fields tend to decay. In these earlier studies, little attempt was made to compare numerical solutions with geomagnetic features. They only concerned westward drift of the nondipole part

of the field [*Zhang et al.*, 1989] and suggestions that reversal maybe due to the close competition observed between dipolar and quadrupolar field geometries.

2.2. Glatzmaier and Roberts

[15] In a series of papers, Glatzmaier and Roberts showed that chaotic dynamos exhibiting features similar to those of the geomagnetic field can be simulated with current computers. Their approach was different from that of Zhang and Busse. Instead of following successive bifurcations (or modifications) of the solution as the Rayleigh number was increased, starting with a nonmagnetic solution, they fixed the Rayleigh number and exhibited a dynamo solution starting with large initial magnetic fields. The Elsasser number (defined on the maximum value of the field) was found to be larger than one.

[16] They successively used two approaches to model the liquid outer core. The first approach [*Glatzmaier and Roberts*, 1995a, 1995b, 1996a] relied on the Boussinesq approximation. Essentially, all density variations were neglected, except for the buoyancy effect. This approximation has been used in most subsequent numerical models. It is not quite appropriate for the Earth, since density varies with depth in the core, but it is a reasonable approximation. The second approach [*Glatzmaier and Roberts*, 1996b and following] relied on the more appropriate anelastic approximation, allowing for varying properties of the Earth with depth. This approximation and its application to the Earth's core had been fully introduced by *Braginsky and Roberts* [1995]. It was not found to yield drastic changes on the solution.

[17] The first model [*Glatzmaier and Roberts*, 1995a, 1995b] was the first to exhibit reversals in a configuration trying to mimic the Earth's core. This model is very sophisticated and incorporates many different features (hyperviscosity, rotation of the inner core, thin conducting layer at the base of the mantle, etc.). Numerical integration of the model was performed over some three dipole diffusion times (to attest dynamo action). Inertia was dropped in this integration, because of the low value (10^{-6}) of the coefficient scaling it (for geophysically realistic values of parameters). This term was discarded, whereas numerical reasons imposed that a 10^9 times smaller term (viscous effects) be retained. The model incorporated enhanced viscosity in boundary layers. The authors later noted [*Glatzmaier and Roberts*, 1996a, 1997b] that this feature was at least partly responsible for destabilizing the solution (which led to the reversal) and significantly altered the field spectrum. When removing this additional enhanced turbulent viscosity in boundary layers, *Glatzmaier and Roberts* [1996a] noted that the solution got more strongly dipolar and did not reverse for the remainder of their simulation. Other modifications included reintroduction of inertia for the zonal axisymmetric components of the flow and a significant reduction of the electrical conductivity of the lower mantle [see *Glatzmaier and Roberts*, 1996a]. It was integrated further over less than one dipole diffusion time. They observed flux patches in these models and noted that the nondipolar part of the simulated field at the outer boundary (CMB) was qualitatively similar in structure to that on Earth and had westward drifting features [*Glatzmaier and Roberts*, 1997b].

[18] The second model [*Glatzmaier and Roberts*, 1996b, 1996c] was similar to the first one (in its 1996a version), but it used an inhomogeneous (or anelastic) approximation. This approximation is based on the assumption that velocities in the fluid flow are small compared with acoustic velocities. The temperature gradient at the outer sphere was superadiabatic and uniform. The associated

heat loss generated thermal and compositional sources at the inner sphere boundary by cooling the whole domain. Both thermal and compositional driving mechanisms were considered, and model parameters were slightly modified, the hyperdiffusivity was reduced. The field generated by this model again showed a tendency to drift westward. For both models the behavior was very different inside and outside the tangent cylinder (the cylinder having the same axis as the axis of rotation and the same radius as the inner core). The most intense magnetic activity was confined to the interior of this cylinder (in contrast with most other models reported below). This second model was also studied in different configurations [see *Glatzmaier and Roberts*, 1997a]: heterogeneous heat flux at the outer sphere (inspired by lower mantle tomography), subadiabatic model, and subadiabatic model with varying heat flux.

[19] Important ingredients of all these simulations were assumptions concerning the way the small-scale turbulent flow, which could not be numerically resolved, affected the large-scale flow and field. Crucial to the success of the numerical work was the introduction of enhanced dissipation of the higher spectral components, a procedure described as using "hyperviscosity." The same was also applied to magnetic and thermal diffusivity, and referred to as using "hyperdiffusivity." This procedure, though essential for the numerical work, is difficult to justify on physical grounds. In very approximate terms we might say that the simulations introduce enhanced turbulent diffusivities in the horizontal directions, rather than in the directions given by the rotation axis and the large-scale magnetic field, as identified by *Braginsky and Meytlis* [1990]. The effective Ekman number varies with the spherical harmonic degree of the field. The effect of hyperviscosity on the large-scale solution is unknown, though *Zhang and Jones* [1997] showed that the introduction of hyperviscosity in convection simulations not only has some impact on the effective Ekman number (see Table 1 for the Ekman number at various degree l) but also modifies the force balance in the momentum equation and thus the asymptotic behavior itself (see the discussion by *Sarson and Jones* [1999]). This issue was recently addressed in the dynamo regime by *Grote et al.* [2000b] (see section 2.4). They found that the effects of hyperdiffusivities increased as the Ekman number was decreased.

[20] In both models the authors observed that the solid core rotates faster than the mantle. There is an ongoing debate in the seismological community to determine whether such super-rotation of the inner core is actually observed in the Earth [*Song and Richards*, 1996; *Su et al.*, 1996; *Souriau et al.*, 1997; *Souriau*, 1998a, 1998b; *Poupinet et al.*, 2000]. Magnetic observations of a polar vortex were recently reported [*Pais and Hulot*, 1997, 2000; *Olson and Aurnou*, 1999]. These observations appear to be compatible with a thermally driven rotation of the inner core [*Aurnou et al.*, 1996; *Pais and Hulot*, 2000] but also with a nonrotating inner core [*Olson and Aurnou*, 1999]. *Kuang* [1999] argued that this effect could be an artefact, due to excessive viscous coupling in numerical models. The issue of a possible rotation of the Earth's inner core is therefore not yet resolved.

[21] *Glatzmaier et al.* [1999] studied reversals with this model and various patterns of heat flux at the core-mantle boundary. They extended time integration over some 15 dipole diffusion times. Surprisingly, they found that the case with uniform heat flux at the core-mantle boundary appeared the most Earth-like. The simulation using a prescribed heat flux inferred from seismic velocity anomalies (tomography) in the lowermost mantle did not

produce a very realistic field (a reversal lasted more than one dipole diffusion time), plus the excursion frequency appeared much too high. The authors suggested that variations in heat flux at the core-mantle boundary could thus be smaller than previously thought. Interestingly, varying numerical resolutions, they noted that the field strength depended on the resolution used. The dipole moment, on average lower than that stimated for the Earth, was found to be greater or similar when numerical resolution was increased. Very recently, *Coe et al.* [2000] performed a detailed comparison of reversals in these numerical simulations. Both homogeneous and heterogeneous heat fluxes at the outer boundary were found to feature low intensities during reversals, with longer-term intensity variation resembling sawtooth behavior. Statistical studies of virtual geomagnetic pole (VGP) paths, i.e., paths followed by the poles of a dipole that would produce the observed directions at a given site, were performed during reversals. In the case of the heterogeneous model it was suggested that VGP paths seem to correlate with high heat flux areas, though only two reversals were available.

[22] Parameters used for these models are presented in Table 1. The choice of these parameters is of course not unique and other parameters could be added (e.g., the Roberts number, introduced in section 2). An interesting number is the ratio of the Ekman to the magnetic Prandtl number, sometimes referred to as the magnetic Ekman number, and representing the ratio of the rotation timescale (the day) to the magnetic timescale. This number is important when comparing numerical with actual time units (as will be discussed in section 3). It is close to 10^{-9} for the Earth. Parameters used in the Glatzmaier and Roberts models were set to approach this value (using nominal coefficients, i.e., neglecting hyperdiffusivities). It is useful to recall that viscosity does not enter the definition of this number.

2.3. Kuang and Bloxham

[23] *Kuang and Bloxham* [1997] presented another model that they called "an Earth-like numerical dynamo model." They studied a Boussinesq model including a conducting inner core. They assumed a conductive layer above the core-mantle boundary and used hyperviscosity, only axial inertia was retained. Reported integration times range over four dipole diffusion times. Their solution differs greatly from the Glatzmaier and Roberts models: convection and induction occur mainly outside the tangent cylinder. Many differences exist between both models (definition of hyperviscosity, heat flux, rotation rate, boundary conditions, etc.). *Kuang and Bloxham* [1997] attributed most of the differences in the numerical results to the kinematic boundary conditions. Moreover, they produced a solution comparable to the Glatzmaier and Roberts model when using the same kinematic boundary conditions. Glatzmaier and Roberts used a no-slip (or rigid) kinematic boundary condition: this condition is physically realistic (motion of a fluid near a solid wall should vanish). Kuang and Bloxham, on the other hand, used free-slip (or stress-free) boundary conditions (no penetration but also no tangential stress). This condition has the convenient property of suppressing sharp boundary layers (numerically difficult to resolve). The authors claimed that since viscosity is very small in the Earth's core, the effects of these layers could be neglected, and thus the layers be suppressed from the computations. They suggested that viscous effects (associated with high Ekman number values) in the Glatzmaier-Roberts dynamo models were especially important near boundaries (in boundary layers). They claimed that by using stress-free boundary conditions, hence suppressing strong viscous effects at the boundary, they would reduce viscosity effects in their model to an almost negligible level, bringing it closer to expected Earth behavior [*Kuang and Bloxham*, 1997].

The possibility of boundary layer instabilities in the parameter regime thought to be valid for the Earth [*Desjardins et al.,* 1999, also Instability of Ekman-Hartmann boundary layers, with application to the fluid flow near the core-mantle boundary, submitted to *Physics of the Earth and Planetary Interiors*, 2000] however raises questions on the validity of this simplification.

[24] *Kuang and Bloxham* [1998] next presented a comparison of their results with the actual Earth's magnetic field. They proposed an interpretation in terms of physical quantities, under the assumption that viscous effects were negligible in their computation (this scaling is comparable to the one used by Glatzmaier and Roberts). They focused their comparison on westward drift, episodes of flux expulsion, and behavior similar to the so-called "Pacific dipole window" (i.e., low secular variation in the Pacific hemisphere, see below). They compared the spectrum of the field generated by their dynamo with that based on historical measurements of the Earth's magnetic field. While other groups generally compared the spectrum of their dynamo fields with the Earth's field for 1980 (well known because of MAGSAT data), Kuang and Bloxham used the *Bloxham and Jackson* [1992] reconstruction of the mean field over the last 300 years (this field model incorporates some additional assumptions, including damping of higher degree terms). They also noted that flow at the core surface was similar to that inferred from secular variation. However, they did not report any polarity reversal. The full description of their numerical model was recently published [*Kuang and Bloxham*, 1999].

[25] *Kuang* [1999] investigated the role of boundary conditions in this dynamo model. He presented two computations, one with stress-free boundaries and the other with no-slip boundaries. His main concern was to show that when viscous couplings at the boundaries are eliminated (free-slip conditions), the axial Lorentz torque acting on cylindrical surfaces (coaxial with the axis of rotation) almost balances fluid inertia, leading to the suggestion that viscosity had little effect on the solution. Kuang showed that with free-slip boundary conditions the viscous torque was smaller than the Lorentz torque (by a factor of 2) on most of these cylinders. Using free-slip boundary conditions led to an oscillating (rather than rotating) inner core. Note finally that the Elsasser number was measured using averaged field values.

2.4. Busse and Coworkers

[26] *Wicht and Busse* [1997] presented the study of numerical dynamos, using a Boussinesq approximation with full inertia, as well as stress-free and insulating boundary conditions. Their study was limited to even azimuthal wave numbers (implying symmetry of the solution about a meridional plane). The authors concentrated on numerical resolution and sequences of bifurcations (see the discussion of the model by Zhang and Busse above) at low Prandtl number values ($Pr \simeq 0.1$). They showed that the subcritical dynamo effect disappeared when numerical resolution was increased. Thus they could not produce numerical dynamos for Rayleigh numbers below the critical value for convection (except when restricting their comparison to the dominant azimuthal wave number). Later, *Busse et al.* [1998] used a Chebychev (instead of Fourier) radial scheme and studied the influence of the Prandtl number. They investigated chaotic dynamos and noted a strong dependence of the mean zonal flow on the Prandtl number. More strikingly, they noted that the ratio of magnetic to kinetic energy density is proportional to the Prandtl number (after nondimensional reasoning and confirmation by their numerical results). Because the inertial term scales as the inverse of the Prandtl number, magnetic energy density would necessarily be larger than kinetic energy density in models where the inertial term is dropped (e.g.,

Glatzmaier and Roberts' first model, Jones et al.) or partly dropped (e.g., Glatzmaier and Roberts' following models, Kuang and Bloxham). This emphasizes how the popular distinction between weak and strong field dynamos in numerical models can be very misleading (see also the nontrivial definitions of these notions in the beginning of this section). The authors also noted that the ratio of magnetic energies of the axisymmetric to nonaxisymmetric components of the induced field increased strongly with the Prandtl number. *Busse et al.*, [1998, p. 212] stressed that "the impression currently en vogue in parts of the scientific community that the problem of the origin of the Earth's magnetic field has been solved is overly optimistic" and insisted on the need for numerical dynamos with magnetic Prandtl numbers lower than unity (not yet achieved so far). *Grote et al.* [1999] showed that for (1) Prandtl number unity, (2) magnetic Prandtl numbers in the neighborhood of unity, and (3) with Ekman numbers ranging from 6×10^{-5} to 2.4×10^{-4}, the field generated by dynamo action is dominantly axial quadrupolar. They argue that this preference is related to the suppression of convection in polar regions observed at lower Ekman numbers. *Grote et al.*, [2000a, p. 270] present a more detailed parameter space and distinguish between regular and chaotic dipolar dynamos, quadrupolar dynamos, and hemispherical dynamos (where the magnetic field nearly vanishes in one hemisphere due to "roughly equal contributions from the axial quadrupolar and dipolar fields"). The type of solution is found to depend mainly on the magnetic Prandtl number and on the Rayleigh number. Dipolar dynamos were only reported for magnetic Prandtl numbers above 10 [see *Grote et al.*, 2000a, Figure 1]. *Grote et al.* [2000a, p. 271] insisted that in such simulations "the magnetic field does not seem to influence the convection velocity very much." *Grote et al.* [2000b] investigated the role of hyperviscosity on their dynamo model and noted that unrealistic dynamical effects could be introduced. The ratio between axisymmetric and nonaxisymmetric components of the magnetic energies was also found to be larger in hyperdiffusive simulations (by a factor 4). The authors note that the differences between hyperdiffusive and regular dynamos increase when the Ekman number is decreased and the Rayleigh number increased.

2.5. Kageyama and Sato

[27] *Kageyama and Sato* [1997a, 1997b, 1997c] presented a dynamo model in a rotating spherical shell filled with a conductive ideal gas; thus they solved equations in the compressible case. They used rigid kinematic boundary conditions and included full inertia. Magnetic boundary conditions in their model were not very geophysical: only the radial component of the field was allowed to be nonzero at the boundary (the authors argue that this ensures that the Poynting vector flux through the boundaries is zero). They first [*Kageyama and Sato*, 1997a] concentrated on the linear growth phase (before saturation) and described how the field was increased by convective columns. They next [*Kageyama and Sato*, 1997b] studied a (numerical) strong field dynamo, in which the magnetic field was found to be strongly dipolar, the second largest multipole being the octupole (representing 77% of the dipole in terms of energy). Kageyama and Sato concluded that dipolar fields were a natural consequence of columnar convection in a rotating spherical geometry. *Kageyama and Sato* [1997c] focused on the magnetic field distribution at the CMB and in the equatorial plane. Recently, *Kageyama et al.* [1999] extended integration time from some 15 to about 50 dipole diffusion times and presented a numerical polarity reversal in their model (the magnetic Prandtl number was here increased to 15). They observed that the dipole reversal was followed by a sequence of octupole reversals,

each one correlated with energy jumps, and suggested that paleomagnetists should look for such features in data.

2.6. Christensen-Olson-Glatzmaier

[28] *Christensen et al.* [1998] published a dynamo model without hyperdiffusivity, retaining full inertia and using no-slip boundary conditions. Boundary conditions were rigid and insulating (including the inner core). The authors noted that their model qualitatively reproduced several structural characteristics of the present-day geomagnetic field: a dipole-dominated magnetic field, with distinct bundles of strong radial flux at high latitudes, low flux over the poles, and paired spots of reversed flux near the equator. The magnetic flux contribution to the dipole moment was concentrated in distinct patches at about 60° latitude. The field generated appeared to be more dipolar than the present geomagnetic field. Its structure appeared to be comparable to that found in the work of Kuang and Bloxham's model.

[29] In a later study, *Olson et al.* [1999] exhibited two regimes for dynamo action, which depend on the Rayleigh number (energy input). The first one (low Rayleigh number) corresponds to a strongly columnar regime (comparable to laminar convection with no magnetic field). The dipole-dominated external field then appears to be formed from a superposition of flux bundles. The authors observed a pattern of reversed flux patches that propagated in the westward direction at low latitudes. At higher Rayleigh numbers they observed the development of convection inside the tangent cylinder, characterized by polar upwelling and azimuthal winds. This second regime reproduced some of the features thought to be typical of the present magnetic field: concentrated flux patches, polar minima in field intensity, and episodes of westward drift (this regime contains some elements of similarity with the Glatzmaier and Roberts dynamo).

[30] Recently, *Christensen et al.* [1999] published a systematic parameter study of their dynamo model and varied the kinematic boundary conditions. Again, they obtained mainly axial dipolar fields (with a few quadrupolar exceptions). Interestingly, they found little influence of the kinematic boundary conditions (no-slip versus free-slip) in their model, contrary to what Kuang and Bloxham observed. They suggest this discrepancy is due to the use of an insulating inner core, which suppresses magnetic torque on the inner core. They studied various magnetic Prandtl numbers (they decreased, for the first time, this parameter slightly below one and presented dynamos at Pm = 0.5). As the magnetic Prandtl number was decreased below a critical value for a given Ekman number, the authors observed that no self-sustained dynamo action could occur. This is an important result, because the minimum critical values they obtained are about one million times larger than the magnetic Prandtl number relevant to the Earth's core. They observed, however, by decreasing the Ekman number by a factor of 10, that the critical value decreased as well. By computing the exponent for the evolution of this parameter (over a rather short range), they estimated that an Ekman number as small as 10^{-12} would be necessary to sustain dynamo action in the regime of small magnetic Prandtl number (10^{-6}). This confirms the need to study high rotation rates (i.e., small Ekman numbers) to observe dynamo action in a medium with efficient magnetic field diffusion (small magnetic Prandtl number), as is the case for the Earth's core.

[31] The magnetic fields they obtained at the outer boundary for two of their dynamo models were of smaller scale, with flux spots stretched in a meridional direction. The authors sug-

gested that the field at the Earth's core-mantle boundary could be small scale as well (despite the much larger magnetic diffusivity) but would appear large scale because of the limited resolution used to represent the core field (largely because of difficulties in removing crustal effects). Broad features (such as patches) in a low-pass filtered representation of their results compared well with those observed in the 1980 field model. No reversal nor significant excursion was observed in any of these studies.

[32] *Kutzner and Christensen* [2000] studied the influence of different driving mechanisms on this model. They reported dipolar dynamos when modeling chemical convection or thermal convection with an imposed temperature difference between the ICB and the CMB. When using purely internal heating (e.g., radioactive heat sources in the core), they only reported dynamos producing quadrupolar (or more complex) fields (see also, in section 2.4, quadrupolar dynamos reported by Grote et al.).

2.7. Kitauchi and Kida

[33] *Kitauchi and Kida* [1998] presented numerical simulations with a Boussinesq model with full inertia. Boundary conditions were somewhat ungeophysical: vacuum was assumed outside the shell. Their solution took the form of a weak field dynamo with periodic reversals of the dipole moment. The flow consisted in convection columns (with a fivefold symmetry); they observed that the magnetic field did not share the same symmetry and studied the mechanism of field intensification. Kitauchi and Kida noted that the structure of the external magnetic field they obtained bore some similarities with the Earth's field: the pattern of the magnetic field was antisymmetric with respect to the equatorial plane and strong magnetic flux in the middle latitudes drifted westward. *Kida and Kitauchi* [1998] increased the magnetic Prandtl number and reported chaotic polarity reversals of their solution. Four observational constraints were considered in this article: dominance of the dipole component, chaotic nature of dipole reversals, quasi-threefold symmetry around the rotation axis, and westward drift of the magnetic field. They noted that these features were all well reproduced by their model except for the quasi-threefold symmetry. Using the reported magnetic energy, one can estimate the Elsasser number to be close to 60.

2.8. Sakuraba and Kono

[34] *Sakuraba and Kono* [1999] presented the study of a numerical dynamo model in a spherical shell and in a full-sphere simulation. They used a Boussinesq model and retained full inertia. The authors used hyperviscosity (see definition section 2.2). The aim of their work was to study the effect of the conducting and free-to-rotate inner core on field generation. They observed that field induction was particularly important near the zone of strong shear associated with boundary layers (precisely those boundary layers suppressed in models using free-slip kinematic boundary conditions). They reported an "Earth like dipole field" and concluded that the thin boundary layers were important in the magnetic field generation process. They observed the stabilizing effect of the inner core on the magnetic field and suggested that growth of the inner core could help interpret the low paleointensity records from the Archean.

2.9. Katayama and Coworkers

[35] *Katayama et al.* [1999] presented a Boussinesq dynamo model with full inertia and an insulating inner core. Stress-free (or free-slip) boundary conditions were used (suppressing boundary layers). The authors examined the magnetic field generation mechanism of their solutions. The magnetic field appeared to be confined in cyclonic convection columns (in a

manner comparable to the models of Zhang and Busse or Kageyama and Sato presented above). They describe the "numerical weak field" regime (the Elsasser number can be estimated, based on the magnetic energy, to be less than 10^{-3}). In this regime, and with Ekman number of 0.07, they still obtained a dominantly dipolar axial field. They also noted that patterns of velocity and magnetic field drift westward in their simulation, although they did not observe any reversal (so far) in the parameter range they studied.

2.10. Jones and Coworkers

[36] We also would like to briefly report some results obtained with a mean field approximation model introduced by *Jones et al.* [1995]. In this approach, only an arbitrary azimuthal component was considered in addition to the mean axisymmetric component. The authors used a two-mode approximation, which is self-consistent and is referred to as a 2.5-dimensional model. Although this model is not a full three-dimensional solution to the MHD equations, some of the results they reported have Earth-like behavior and can shed some light on the behavior of fully 3-D models. The main advantage of their approach is to make numerical integration lighter, thus allowing the extention of the range of studied parameter values. *Jones et al.* [1995] studied different parameter regimes and found a (numerical) "weak field" regime as well as a (numerical) "strong field" regime. The fields were largely axial and dipolar but had an axisymmetric dipolar component about twice as large as in the Earth and a larger value of the nonaxisymmetric components. *Jones et al.* [1995] also studied a model in which a stable layer was introduced at the top of the core in order to decrease these discrepancies.

[37] *Sarson et al.* [1998] presented a systematic study of the 2.5-D model, using hyperdiffusivity for some simulations only. This model produced solutions very similar to some of the previously published 3-D numerical dynamos, yet at much lower cost. At the higher values of the Ekman and Roberts numbers and moderately supercritical Rayleigh numbers, results were very similar to the *Zhang and Busse* [1988, 1989] solutions. At lower values of these parameters (with hyperviscosity) the solution reproduced most of the characteristics of the *Glatzmaier and Roberts* [1995a, 1995b, 1996a, 1996b] model. *Sarson and Jones* [1999] studied magnetic field reversals in more detail in the latter parameter regime. The origin of reversals was interpreted in terms of "buoyancy surges" occurring inside the tangent cylinder. It was demonstrated that suppressing the Lorentz force did not significantly alter the basic form of the flow, again showing that viscosity has a very important effect on numerical solutions and that the magnetic field exerts a relatively minor influence on the solution (despite the dynamo being a "numerically" strong field).

2.11. Conclusions on Geodynamo Models

[38] We summarize the results presented above in Table 1. The first three columns of Table 1 recall parameter values used in the models (using the common nondimensionalizations introduced above); the next three columns summarize the differences in the model definitions and integration times; the last five columns summarize the outputs from these models. All models succeed in producing dipolar fields. Most of them are strong field models according to the definition used by numericists (i.e., Elsasser number above unity). Note that the Elsasser numbers reported here are those given by the authors (or derived from values of magnetic energy). They are based in most cases on the mean value of the magnetic field throughout the core, although some authors prefer to use its maximum value (e.g., Glatzmaier and Roberts). Actually, most models produce Elsasser num-

bers much larger than unity, opposed to what is expected in the asymptotic regime (see discussion at the beginning of this section). Also, several authors [*Grote et al.*, 2000a; *Sarson et al.*, 1998] have pointed out that the Lorentz force could be suppressed from numerical integration without affecting much the geometry of their solutions. Reversals were observed in some models but relatively short integration times (below 5 dipole diffusion times for most models) do not allow a significant distinction between models. Most models are found to have westward drifting features, and a few other Earth-like features have also been noted in the last column.

[39] We wish to stress that the most fundamental point to have emerged from numerical simulations within the last few years is probably that most of the available numerical models, even with Ekman numbers close to 0.1 [see *Katayama et al.*, 1999] succeed in producing an external magnetic field largely dominated by an axial dipole component (with a few quadrupolar exceptions, see *Grote et al.* [1999, 2000a] and *Kutzner and Christensen* [2000]), with weaker nondipolar structures and in some cases even polarity reversal of the dipole. With perhaps exaggerated optimism one could argue that despite drastic differences in the parameter regimes, numerical dynamos capture the essential features of the Earth's magnetic field generation mechanism. Somewhat puzzling, though, are the discrepancies on the generation mechanisms involved in these models. Some models generate the magnetic field by chaotic motions mainly inside the cylinder tangent to the inner core (e.g., Glatzmaier and Roberts), others by chaotic motions mainly outside this cylinder (e.g., Kuang and Bloxham); some propose a giant alpha effect in columnar structures comparable to nonmagnetic convection rolls (e.g., Christensen et al. or Kitauchi and Kida); some stress the role of boundary layers in the field generation mechanism (e.g., Sakuraba and Kono), whereas others artificially suppress these layers (e.g., Kuang and Bloxham, Busse et al., or Katayama et al.). This stresses the need for robust yet detailed constraints to describe what an Earth-like magnetic field actually should be and hopefully establish whether any of the available models fulfills these constraints. This is the object of the following two sections.

3. Evaluation of Observational Constraints

[40] A picture of the present-day geomagnetic field would, of course, not be sufficient to describe the Earth's magnetic field, much in the same way it would be impossible to determine the climate of a region with only a snapshot of the weather at an arbitrary time. We therefore present here a short review of observational facts established with geomagnetism, archeomagnetism, and paleomagnetism, and attempt to find objective criteria that allow one to propose which observations can be considered as robust and can (should) be used in order to efficiently constrain numerical models. Data provided by geomagnetism, archeomagnetism, and paleomagnetism are based on instrumental and interpolation methods whose heterogeneity must be taken into account; their accuracy and time constants vary over a very large range. Some simple criteria must be sought to provide an objective basis to decide which of these observations are indeed robust and relevant. There is not much worry regarding present-day and historical field determinations over the last centuries, although improving accuracy and spatial coverage are of course important. *Bloxham and Gubbins* [1985] have considerably extended our knowledge of the geomagnetic field over the last three centuries, using ancient ship logs. Confidence in the results is a function of irregular data distribution and, for instance, lack of intensity

measurements prior to 1830. Most controversies or ambiguous observations occur for older periods and longer timescales, when magnetic measurements are linked to acquisition of magnetization by artefacts or rocks. This is due to uncertainties in dating and in the complex process of magnetization acquisition.

[41] For this reason, global geomagnetic features must be properly documented by multiple records from different locations (and different materials whenever possible). It is also important that additional testing be provided by different approaches or alternative parameters. Converging observations reached independently by various teams should be considered as a prerequisite to the evaluation of the records. It is obviously difficult to reach any conclusion about the actual interpretation of the records, as long as there is controversy going on about the validity of the observations or the relevance of the interpretations. It cannot be excluded that in the near future, new data or theories could invalidate some features that we presently retain as actual characteristics of the field. Conversely, some ongoing controversies could be settled by new data. On the basis of these ideas we will attempt to sort observations in three categories: A, the most robust ones that are supported by multiple locations/analyses/authors; B, those that are still at the center of live ongoing controversies, and C, those that have at some time been proposed but are now considered most unlikely. We believe there is overall agreement within the community on the first (A) category.

[42] Analyses based on present or historical field data are useful to document most variations with time constants of a few hundred years, which seem to characterize the evolution of the main part of nondipole terms. The mean archeomagnetic field documents longer field changes and appears to resemble the paleomagnetic field [e.g., *Carlut et al.*, 1999]. Its geometry is also much simpler (i.e., lower harmonic degrees prevail) than the complex historical field. As recalled above, such timescale-dependent magnetic behavior, with complex short-term secular variation opposed to longer-term smoother changes, is reminisent of the contrast between the complexity of weather patterns and longer-term evolution of the climate [see *Cox*, 1975]. At all timescales, one seeks to characterize the steady and transient parts of the field, to describe their amplitude and space variations, and to relate one timescale to the next. For instance, *Hulot and Le Mouël* [1994] and *Carlut et al.* [1999] have attempted to identify the typical time constants of individual spherical harmonic components of the field; these components are noted g_n^m and h_n^m and are respectively defined as the cosine and sine terms in the spherical harmonic expansion of the magnetic potential in the mantle (assumed to be an insulator). These timescales are of the order of 500 years for the equatorial dipole (g_1^1,h_1^1) and less for other higher-order terms, whereas the axial dipole (g_1^0) changes much slower. In the following, we summarize our analysis of available data and observations based on their characteristic duration, starting with the shortest ones.

3.1. *From 0.1 Year to 1 kyr*

3.1.1. Secular Variation Impulses

[43] Many summaries of the present-day and historical field time variations are available [e.g., *Merrill et al.*, 1996; *Courtillot and Le Mouël*, 1988]. It is now generally accepted that the conductive mantle efficiently attenuates and therefore screens out variations with internal origin with a time constant shorter than a few months. The shortest phenomena of unquestionable internal origin are the so-called geomagnetic secular variation impulses or jerks [*Courtillot and Le Mouël*, 1984], which have occurred (irregularly) some seven times in the

last 150 years [*Alexandrescu et al.*, 1995, 1996a, 1996b, 1997; *Mandea-Alexandrescu et al.*, 1999; *McMillan*, 1996]. These worldwide events have a total duration of the order of, or possibly less than, 1 year. They possibly propagate and are attributed to brief, major changes in the large-scale flow in the core [*Hulot et al.*, 1993]. They are well established and thus fully satisfy category A requirements. Although the sequence of identified jerks is very short, their occurrence appears to be quite irregular, with seven events between the present and 1870, and none in the preceding 150 years [*Alexandrescu et al.*, 1997; *Mandea-Alexandrescu et al.*, 1999]. The impulses, which are separated by quieter intervals, shape much of recent secular variation. This is found to correlate well with predictions based on time-dependent models of flow in the core, and with observed length of day changes, implying core-mantle coupling with exchange of angular momentum, as was first shown for the 1969–1985 period by *Jault et al.* [1988] and extended to the 1840–1990 period by *Jackson et al.* [1993], then developed in more detail by *Pais and Hulot* [2000].

[44] Interestingly, the fastest impulses have been inferred not from geomagnetic but from paleomagnetic observations: they are those associated with the polarity transition recorded 16 Myr ago at the Steens Mountain in Oregon. In a series of papers [e.g., *Coe and Prévot*, 1989; *Coe et al.*, 1995], a French-U.S. team investigated the origin of a puzzling progressive evolution of paleomagnetic directions in the interiors of two transitional lava flows. Each lava unit recorded a complete sequence of directions going all the way from that of the underlying flow to the direction of the overlying flow. In the absence of any clear evidence for anomalous rock magnetic properties [*Camps et al.*, 1999] these features have been interpreted in terms of very fast geomagnetic changes, which would have reached 10° and 1000 nT per day. For comparison, values typical of the present-day secular variation are of the order of 0.1° and 50 nT per year, i.e., some 10^4 times slower. Such paleomagnetic impulses would thus have been even larger and sharper than geomagnetic jerks.

[45] These geomagnetic time constants could be constrained by estimates of the cooling times of individual flows. However, such rapid changes do not seem to be compatible with accepted values of mantle conductivity [see *Ultré-Guérard and Achache*, 1995]. Additional detailed investigations have therefore been conducted in order to see whether this situation could not have arisen because of remagnetization of the flows. Remagnetizations of lava flows, yielding complex or unusual directions, have been detected at several localities where reversals have been recorded. A first interesting example on Oligocene basaltic rocks was given by *Hoffman* [1984]. Recently, *Valet et al.* [1998] observed the coexistence of both polarities within flows marking the last reversal boundary (0.78 Ma) and also in a lava flow associated with the onset of the upper Réunion subchron (2.13 Ma) in Ethiopia. In these cases a purely geomagnetic interpretation would imply that a full reversal took place in only a few days. As in the Steens Mountain case, there is no striking difference between the rock magnetic properties of these units and the rest of the sequence, but a scenario involving thermochemical remagnetization in the early stages of magnetization is not incompatible with the results. Thermochemical magnetic overprinting can be particularly serious when it affects a flow emplaced at a time of very low field intensity, sandwiched between flows emplaced at a time of full (stable polarity) intensity. The same effects may as well unfortunately alter features of secular variation recorded in lava flows. Although difficult to demonstrate, such remagnetization events are far more likely to be identified when a reversal has occurred than during a period of constant polarity. A recent investigation of additional

flows at Steens by members of the original study team casts doubt on the geomagnetic significance of the paleomagnetic directions [*Camps et al.*, 1999]. Therefore the existence of very large and rapid changes that have been documented from a single site by a unique team remain controversial and thus cannot be retained as a robust characteristic of reversals (category C). The occurrence of complex and hardly detectable remagnetization effects remains the most reasonable assumption.

3.1.2. Westward Drift and Flux Patches (Decades to Millenia)

[46] Among the first observations that have been made over the longer time range of decades to millenia are the decrease in geomagnetic axial dipole intensity (of the order of 5% per century) and the famous westward drift of the nondipole field [see, e.g., *Merrill et al.*, 1996]. This drift of isoporic lines and foci was discovered by Halley in 1692 and quantified by *Bullard et al.* [1950], who evaluated its global rate at 0.18°/yr. *Bloxham and Jackson* [1992] [see also *Bloxham and Gubbins*, 1985, 1986] have emphasized the fact that over three centuries, westward drift has been far from uniform. The observations that the most intense westward drift was found over the Atlantic hemisphere and that the historical nondipole field had been small in the Pacific Ocean have been made by a number of authors [e.g., *Bloxham and Gubbins*, 1985; *Walker and Backus*, 1996; *McElhinny et al.*, 1996; *Johnson and Constable*, 1998]. *Le Mouël* [1984] showed that the large-scale geometry of flow at the surface of the core, under the geostrophic approximation, could account in a rather simple way for this apparent discrepancy between both hemispheres. Since there is controversy regarding its interpretation this observation has been ranked in category B.

[47] *Gire et al.* [1984] tried to estimate the "instantaneous" rate of westward drift and its time changes based on separate individual field components. They showed that this was an ill-posed problem, without a unique solution. Drift estimates are significantly larger when the Y (east) component is used and smaller when X (north) is used. Drift estimates also vary with time, with values at the time of the 1970 secular variation impulse being a minimum. If this is attributed to bodily drift of core fluid, then the drift rate in 1980 was about twice as large as that before. More importantly, *Yukutake* [1967] showed that drift estimates also depended strongly on latitude (see also the more recent analysis of *Jault et al.* [1988]). Under the frozen flux and geostrophic approximations, *Jault et al.* [1988] were able to show that this could be due to differential rotation of core fluid with time constants of the order of 10 years and overall symmetry with respect to the equator. The characteristic latitudinal patterns of drift estimates based on components X, Y, and Z can be interpreted in terms of rigid rotation of axial cylindrical annuli about the Earth's rotation axis. The corresponding zonal flow model has those annuli intersecting the CMB at latitudes less than 40° rotating westward at about 0.1°/yr, whereas annuli at higher latitudes rotate at a similar velocity but eastward. As a result, the overall changes in the angular momentum of these core annuli roughly balance changes in the angular momentum of the mantle (or length of day). The various results reported, i.e., the large variability of estimates as a function of geophysical area (Pacific versus non-Pacific hemispheres), latitude, geomagnetic component, and time, cast serious doubt on the physical reality and meaning of a (bodily) westward drift of geomagnetic features. A recent investigation of torsional oscillations [*Pais and Hulot*, 2000] illustrates this issue: the variation of angular velocities with latitude is far from constant, neither is the time variation of the sign of the component of the flow corresponding to an axial rotation (T_1^0). A systematic and

global westward drift certainly enters the category C.

[48] Evidence for drift has been sought in archeomagnetic data. Relevant references and discussions can be found in the work of *Merrill et al.* [1996]. On the basis of 2000 years of archeomagnetic data, *Evans* [1987] argued that they demonstrated westward drift at some 0.3°/yr, but this was shown to be a highly nonunique interpretation [*Merrill et al.*, 1996].

[49] The most intriguing feature that emerged from the historical compilations of *Gubbins and Bloxham* [1987] was a concentration of magnetic flux at high latitudes in both hemispheres. These are particularly prominent in the Northern Hemisphere near 120° (both east and west) longitudes and 60° latitude and obviously fit category A. They have been attributed to convection columns in the fluid core, parallel to the rotation axis and tangent to the (solid) inner core. A third patch near 0° longitude, which would be required to complete some form of (threefold) symmetry, and the corresponding mirror images in the Southern Hemisphere are either missing or not well developed. In the Southern Hemisphere this could partly be due to relatively sparse data coverage.

3.1.3. Dipole Versus Nondipole Field

[50] A number of authors have tried to identify persistent features in archeomagnetic field models, which extend beyond the range of historical data. Naturally, this entails less detailed representations of the field, with much lower spatial resolution: in general, the mean amplitudes and variations of the lower degree spherical harmonic components are evaluated and tested against a priori statistical models of the field. Such a model has been proposed (actually based on paleomagnetic data) by *Constable and Parker* [1988] and extended by *Hulot and Le Mouël* [1994]. Spherical harmonic components of the field are supposed to be independent stationary central random Gaussian variables with known correlation functions, zero mean (except g_1^0 and g_2^0) and variances depending only of degree n. This approach allows one to evaluate the time intervals over which the time series become uncorrelated. On the basis of historical and archeomagnetic data, *Hulot and Le Mouël* [1994] and *Hongre et al.* [1998] have shown that the correlation times for terms with degree 2 or higher were less than 150 years; correlation times for the equatorial and axial dipoles are respectively thought to be of the order of 500 years and much larger. *Hongre et al.* [1998] (see also *Carlut et al.*, 1999] show that historical and archeomagnetic data are compatible with zero long-term averages for the equatorial dipole and for the quadrupole and octupole. When the field is averaged over, say, 2000 years or more, only the axial dipole remains. As no study exists to the contrary, we assign an A rating to the persistence of an axial dipolar field over times over 2000 years.

3.2. *From 1 to 10 kyr*

[51] As one attempts to jump from scales of thousands to tens of thousands of years, one crosses the boundary where archeomagnetism and artefacts are replaced by paleomagnetism and natural rocks. Yet finding rocks that record field changes with a ~10 kyr resolution is very difficult. The most limiting factor is the general difficulty in correlating records in time over large distances. This is typically the case for secular variation records covering the past 10 kyr. Sediments with high deposition rates are required to document nondipole field changes with sufficient resolution. As a consequence the corresponding database has been mostly obtained from lake sediments. A large number of records has been published from western Europe, North America, Australia, and Japan (see *Opdyke and Channel* [1996] for a more de-

tailed account). Correlation of individual records even over short distances requires very accurate dating. The better dated data sets from Britain have been merged together into a well-dated "master curve" of directional changes for the past 10 kyr. The variations seem to match over distances of up to 2000 km, from western Europe to Iceland, but are incompatible with the record from the Sea of Galilee. Secular variation from Australia is more difficult to correlate because of the lower amplitude of directional changes. The general lack of similarity of records from Argentina, Australia, and the North American Great Lakes confirms that patterns of secular variation cannot be correlated over large distances. In summary, when considered together, these paleomagnetic records display large-amplitude variations, up to 60° in declination and 30° in inclination and suggest the possibility that secular variation waveforms could be recurrent at some sites [*Lund et al.*, 1988].

3.2.1. Westward Drift

[52] *Runcorn* [1959] championed the use of (D,I) or Bauer diagrams to relate anticlockwise looping to westward drift of nonaxial dipolar sources (see also *Thompson and Barraclough* [1982]). However, *Dodson* [1979] showed that this Runcorn's rule suffered a number of violations. Work on lake sediments on the 10,000 to 30,000 year timescale [e.g., *Denham*, 1975; *Thouveny et al.*, 1990] has suggested alternating periods of clockwise and anticlockwise rotation in Bauer diagrams, which would correspond to alternate periods of eastward and westward drift. These have been sometimes interpreted as evidence for propagating dynamo waves [e.g., *Parker*, 1979; *Hagee and Olson*, 1989]. However, none of these studies have been clear or systematic. In any case, the significance of westward drift, which was emphasized in the 1950s and until the 1970s, has decreased. It is no longer considered as a characteristic feature of the field and its secular variation, and thus enters category C. It is not a robust feature that can be imposed on dynamo models, despite the fact that it seems to have remained ingrained in some recent presentations of geomagnetic field behavior.

3.2.2. Excursions and Reversals

[53] In contrast to reversals, which correspond to a full polarity change, excursions reflect deviations from the geocentric axial dipole (GAD) that are only considered to be "larger than normal" paleosecular variation. The meaning of such deviations remained controversial for a long time because they were mostly observed at a single location and associated with a large reduction in magnetization intensity, which could be indicative of remagnetization. It became clear that excursions of the geomagnetic field did exist when the same events could be detected at distinct locations, and in some cases in different materials, i.e., volcanics and sediments [*Verosub and Benerjee*, 1977; *Verosub*, 1982; *Champion et al.*, 1988; *Quidelleur and Valet*, 1994]. Excursions and reversals share several common characteristics, which suggest that they can be treated as manifestations of similar processes within the core. Alternatively, excursions are regarded by some authors as an intrinsic part of secular variation, whereas reversals would involve longer time and larger spatial scales [*Lund*, 1997]. The main question is thus to know whether paleomagnetic records are detailed and accurate enough to make a clear distinction between these two possibilities.

3.2.3. Apparent Duration of Excursions and Reversals

[54] Excursions and reversals are, geologically speaking, rapid phenomena and thus their exact duration is extremely delicate to evaluate. In their recent review on polarity transitions, *Merrill and McFadden* [1999] have summarized

four distinct approaches to estimate the duration of reversals. Estimates derived from cooling of intrusive igneous rocks are not very reliable because of uncertainties in cooling rates. Statistical approaches involving the number of occurrences of intermediate directions are also not very meaningful. Indeed, in many cases, records have been obtained from single lava flows so that it is impossible to distinguish intermediate directions associated with excursions from transitional directions during reversals. Actually, it is common to encounter lava flows with directions pointing farther away from the geographic pole than expected for usual secular variation during periods without reversals. Direct absolute dating of long sequences with well-identified events is the most reliable approach with volcanics. However, most records are incomplete owing to the episodic nature of volcanic activity. Oversampling of particular field directions is frequent, as is the absence of recording during other periods. Above all, the uncertainties inherent to dating techniques are at least equivalent to the duration of the event. Combining and averaging datings of parallel records does not reduce uncertainties and thus does not solve the problem.

[55] An interesting case is the Laschamp event, which was initially detected in several lava flows from the French Massif central at Laschamp, Olby, and Louchardière [*Bonhommet and Babkine*, 1967]. The Laschamp event is by far the most extensively dated excursion. Three sets of measurements performed at three laboratories on the same sample sites [*Gillot et al.*, 1979] indicate that absolute ages of the Laschamp and Olby flows lie between 38 and 50 ka with a mean estimate of 46.6 ± 2.4 ka [*Levi et al.*, 1990]. However, there is strong overlap of the determinations when dispersion is taken into account. This conclusion is reinforced when adding results from Icelandic lava flows, which very likely recorded the same event with a mean age of 42.9 ± 7.8 ka obtained from 19 determinations [*Levi et al.*, 1990]. Thus, despite the existence of detailed records, it is impossible to obtain better estimates for the duration of this event.

[56] Records from sediments could be more appropriate. The best and most detailed records of the Laschamp event have been published recently by S. Lund et al. (Deep sea sediment records of the Laschamp geomagnetic field excursion (~40,000 14C years BP), submitted to *Geophysical Journal International*, 2000) based on marine sediments with very high deposition rates. The results indicate a duration of 2000 years for the period with low intensity and a shorter one for directional changes. The fact that intensity changes would last longer than directional changes is readily understood [*Mary and Courtillot*, 1993], if one accepts that the (axial) dipole plays a special role among other field components [*Courtillot et al.*, 1992]. Indeed, the intensity of the (axial) dipole needs to be reduced by a very significant factor (of the order of 3 to 5), compared to its "normal" values, before dipolar dominance in field geometry at the Earth's surface is significantly affected (and field directions or VGPs start departing strongly from a dipole-dominated field).

[57] These estimates obviously require that the mean deposition rate was high enough and more or less constant and also that smoothing resulting from postdepositional reorientations can be neglected. Actually, whereas records with very high deposition rates are supposed to provide the best estimates, in such cases, there is an increased probability that the deposition rate was highly variable. It is often claimed that field reversals last longer than excursions because excursional directions are rare in sediments while transitional directions are more frequent, which would explain also why there are so few estimates of their duration. However, several factors can affect the sediment thick-

ness over which the magnetic event has been recorded. Significant delay in the acquisition of magnetization can generate smoothing of non-antipodal pre- and posttransitional directions and thus create artificial intermediate directions [*Rochette*, 1990; *Langereis et al.*, 1992]. This smoothing effect is more pronounced for reversals than for excursions for which the pre- and postevent directions are close and not antipodal [*Quidelleur and Valet*, 1994]. In the same vein, complications can arise from early diagenetic processes [*Van Hoof*, 1993], with particular intervals being affected by dissolution and/or precipitation of minerals, resulting in the obliteration of an excursion or alternatively in a longer apparent duration.

[58] Not only are there differences between the apparent duration of different events but also for the same event recorded at different locations. This is particularly well documented for the last reversal. Such variability could reflect the fact that the apparent duration of reversals or excursions actually depends on site location. Indeed, the intensity of the axial dipole decreases during reversals and excursions and higher-order terms become more prominent or may even dominate field geometry. Because they fluctuate with a much shorter spatial wavelength, the duration of directional changes can be strikingly different at different locations [*Quidelleur and Valet*, 1996; *Carlut et al.*, 1999]. This would explain why very different durations (from a few years to 10 kyr) have been published for the Brunhes-Matuyama reversal and also why excursions are so difficult to observe.

[59] In summary, we do not find convincing evidence for significant differences between durations of excursions and reversals. In most cases, directional changes would take place in less than 5 kyr, which justify that this upper limit actually fits category A. However, durations as short as 1 kyr or even less have been proposed based on high-resolution records [*Love and Mazaud*, 1997; *Merrill and McFadden*, 1999] and cannot be totally excluded. There are also longer durations such as for the Gauss-Matuyama [*Glen et al.*, 1999]. *Gubbins* [1999] suggested recently that the field would reverse in the liquid outer core during excursions but not in the inner core, whereas reversals would involve the entire core. A similar suggestion, based on 3-D simulations of these events, was discussed by *Glatzmaier et al.* [1999]. In fact, Gubbins divides the field into the outer core field and the inner core field, whereas Glatzmaier et al. simulations show that the two different field polarities tend to divide along the tangent cylinder (that is, the field outside the tangent cylinder can reverse its polarity without the field inside the tangent cylinder reversing). These appealing suggestions are not yet testable, since the overall variability of the duration of reversals still does not allow one to argue that they would last significantly longer than excursions. The first critical parameter that should be explored is the apparent duration of a same event at different sites, which depends on details of field morphology.

3.2.4. Field Intensity During Excursions

[60] The absence of records with sufficiently good geographic coverage implies that the documentation of the "excursional" field is relatively limited and thus prevents us from drawing firm conclusions. There is no systematic feature emerging from the present excursion database. Virtual geomagnetic poles (VGP) are often used as a simple way to characterize field geometry during excursions and reversals. The compilation of excursional VGPs does not show any preference for a particular sector of longitude [*Quidelleur and Valet*, 1994]. In fact, the present data set remains fully compatible with the simplest hypothesis, which links excursions to amplified directional secular variation directly related with a large decrease in the geocentric axial dipole (GAD) [*Courtillot et*

al., 1992; *Mary and Courtillot*, 1993; *Courtillot and Valet*, 1995].

[61] Further indications about the origin of excursions come from relative paleointensity records, which demonstrate that most acknowledged excursional events are coeval with a large decrease in the GAD. Recent papers [*Langereis et al.*, 1997; *Lund*, 1997; *Lund et al.*, 1998] report compilations of excursions in the most recent Brunhes (0–780 ka) normal chron. Six excursions can now be reliably considered as worldwide events, whereas five others have not been globally correlated [*Langereis et al.*, 1997]. The six global events are associated with the major intensity lows in the composite paleointensity record of *Guyodo and Valet* [1999]. This is consistent with the observation that nondipole components did not emerge everywhere in the five cases with a larger GAD and also that the apparent duration of each excursion is significantly different at each site [*Quidelleur et al.*, 1999]. Thus the frequency of excursions in the last Myr appears to be much higher and to yield a much more unstable image of the field than was assumed before. Several results indicate that the same characteristics prevailed during periods older than 1 Myr, and it appears likely that excursions punctuated the entire history of the geomagnetic field. We therefore believe that excursions, and their frequency of occurrence, is a major field characteristic that must be considered as an actual constraint for dynamo modeling (category A).

[62] Very few detailed records of absolute paleointensity have been obtained across excursions [*Roperch et al.*, 1988; *Chauvin et al.*, 1991; *Carlut et al.*, 1999]. A noticeable feature is the absence of field recovery (during the excursional period) before the field returns to initial polarity. This is also the case in the sedimentary records. However, we are well aware that the brevity of the event in comparison with the resolution of the records prevents any strong inferences. One can speculate that the absence of recovery could be a sufficient condition for a rapid return to initial polarity; conversely, intense recovery following a geomagnetic instability could be a necessary condition to induce a stable polarity. However, if excursions and reversals are primarily linked with strong decreases in (axial) dipole intensity with little changes in the amplitudes of other (nondipolar) terms, a whole spectrum of excursions will simply be due to fluctuations in dipole intensity. The smaller the fluctuation, the more evasive the excursion. In that case the notion of field recovery is part of the description and definition of the excursion, with no other special physical causal link. Given the very few detailed records obtained so far, it is premature to consider such observations as systematic and hence to give them too much weight in dynamo modeling (thus C rating).

3.2.5. Field Intensity Variations Across Reversals

[63] A large number of paleomagnetic studies have been carried out for about 20 years in order to document field variations during reversals. It is now relatively well established, on the basis of sedimentary as well as volcanic records, that field intensity drops significantly and that these changes last longer than directional changes (see *Merrill and McFadden* [1999] for a complete review). Again, this is simply because a large decrease in dipole intensity is necessary before directions depart significantly from dipolar ones [*Mary and Courtillot*, 1993]. There are also indications that recovery following a transition is short and culminates in higher values than before the decay phase. Indeed, the four detailed volcanic records including determinations of absolute paleointensity provide support for such an asymmetry. Strong posttransitional field values have been reported in a Pliocene

reversal recorded at Kauai [*Bogue and Paul*, 1993] and in the upper Jaramillo subchron recorded from Tahiti [*Chauvin et al.*, 1990]. The comparable characteristics emerge also from the 15 Ma old Steens Mountain reversal [*Prévot et al.*, 1985] and the Matuyama-Brunhes transition recorded from La Palma in the Canary islands [*Valet et al.*, 1999]. Similar observations have been drawn from sediments but over much longer periods (we will return to this later). In the present situation we are only dealing with relatively short time periods preceeding and following the reversals. In this case the present database shows asymmetrical field changes. Since additional records are still needed on this matter, this observation has been classified in category B.

3.2.6. Field Morphology During Reversals

[64] Many reversal studies have attempted to constrain field geometry on the basis of the configuration of VGP paths. After early publications of a few sparse records from volcanic flows, most attention was focused on sedimentary records. It was rapidly accepted that the field was not dipolar during reversals [*Dagley and Lawley*, 1974; *Hillhouse and Cox*, 1976]. Despite still ongoing discussions the nondipolar character of the transitional field is more and more widely recognized and seen as a direct consequence of the overall and systematic drop of dipole intensity. We consider that this is an A level constraint that must be produced by numerical simulations of reversals. Syntheses of paleomagnetic records published about 10 years ago yielded several very interesting and still controversial observations [*Clement*, 1991; *Tric et al.*, 1991; *Laj et al.*, 1992; *Gubbins*, 1994; *Gubbins and Love*, 1998], such as the existence of preferred longitudinal bands for VGP paths, over the Americas and eastern Asia, possibly in relation with the cold circum-Pacific regions in the lower mantle (outlined by seismic tomography). It was shown subsequently [*Valet et al.*, 1992; *McFadden et al.*, 1993] that this coincidence cannot be interpreted without taking into account the poor distribution of paleomagnetic sites, which lie mostly 90° away from the preferred bands. Actually, many factors related to acquisition of magnetization in sediments in the case of low field intensity can generate VGP paths 90° away from the observation sites [*Quidelleur and Valet*, 1994; *Quidelleur et al.*, 1995; *Barton and McFadden*, 1996; *Langereis et al.*, 1992; *Rochette*, 1990]. *Prévot and Camps* [1993] observed also that there is no preferred longitudinal band emerging from the database of volcanic records. Recently, *Love* [1998] questioned this interpretation, arguing that similar directions from successive flows should not be averaged, because they do not necessarily result from a very rapid succession of eruptions. Instead, he treated each individual flow as a single time event (implying no correlation between successive flows, or identically that the duration between flows was longer than the typical correlation times of secular variation; see above). This is a drastic assumption and a violation of the Occam's razor principle. The most extensively dated flows from Hawaii show evidence that such is not the case [*Holcomb*, 1987] and a recent study of the same reversals [*Herrero-Bervera and Valet*, 1999] recorded in three parallel sections show that identical successive transitional directions are not reproduced between parallel sections. This is clear evidence of fast sampling of geomagnetic changes by irregular and spatially limited volcanic flows. Actually, both analyses are limited by the number of sites and by the identification of actual transitional directions. Because VGPs obtained from volcanics can reach latitudes as low as 45–50° during episodes of large secular changes (and of course excursions), one should restrain reversal analyses to "actual" transitional directions, i.e., corresponding to VGP latitudes less than 45° in order to support a more robust analysis. There

is actually no clear distinction between excursional and transitional VGPs; *Love*'s [1998] subset of data shows significant scatter when restrained to "pure" transitional VGPs. Would we wish to rely on VGP positions lower than 45° (and not 60°), the number of points would become far too small to perform any robust analysis. An answer to this question should rely exclusively on volcanic records with well-defined pre- and posttransitional directions and a sufficient number of intermediate directions. Only those records have some significance in terms of transitional field characteristics, without generating unacceptable uncertainties about the origin of the directions and their stratigraphic relationship. The distribution of poles extracted from this very limited database does not display any preferred location nor does it show any evidence for systematics in the reversal process. For these reasons the hypothesis of preferred longitudinal VGP paths fits category B.

[65] Using another selection of records, *Hoffman* [1991, 1992, 1996] pointed out the existence of VGP clusters in the vicinity of south America and above western Australia, which could be due to persistent inclined dipolar field configurations during the reversal process. This suggestion would establish some link between sedimentary and volcanic records. Such features could imply some form of indirect control of the reversal processes by the mantle, generating heterogeneous flow at the CMB. Such heterogeneous constraints at the CMB are not taken into account in most present dynamos. However, VGP clusters could simply indicate that a series of volcanic flows were emitted over a short interval of time compared to the time constants of secular variation (years rather than centuries or millenia). As described above, clusters of directions are not reproduced between nearby sections of the same reversals in Hawaii [*Herrero-Bervera and Valet*, 1999] and are therefore most likely due to a brief spurt of volcanic activity. These cannot be considered as statistically uncorrelated data for the kind of analyses we wish to perform. The apparent concentration of VGPs close to south America and western Australia could therefore still be considered as a coincidence (or a sampling artefact). Thus it cannot be considered as certain yet that long-lived transitional states represent an actual characteristic of the reversing field, and this leads us to rank this observation within category B.

[66] A dominant feature emerging from the most detailed sedimentary [*Valet et al.*, 1986; *Tric et al.*, 1991; *Channell and Lehman*, 1997) and volcanic (*Mankinen et al.*, 1985; *Chauvin et al.*, 1990; *Herrero-Bervera and Valet*, 1999] records is their complex structure, with large directional variations (particularly in inclination) preceeding and/or following the transition. The presence of such large loops shows similarity with secular variation of the present nondipole field. This observation reinforces the concept of a rather complex transitional field that would be dominated by nondipole components, subsequent to the large decrease of intensity that has been observed in all records of reversals. Therefore the simple model that considers the field variations during reversals as identical to historical secular variaton [*Dagley and Lawley*, 1974; *Valet et al.*, 1989; *Courtillot et al.*, 1992] remains in our view the most plausible. The large loops are easily modeled as soon as the drop in (axial) dipole field intensity is large enough to let the typical pattern of secular variation of nondipole terms emerge. This observational fact should in our view be considered as an actual characteristic of transitional field morphology, and could be a useful constraint for numerical dynamos (category A).

3.3. From 10 to 100 kyr

3.3.1. Oscillations of Dipole Moment

[67] In principle, field intensity changes should not be decoupled from directional

changes and the magnetic field vector should be studied as a whole. Unfortunately, paleointensity studies were only developed extensively during the past decade. Except for the past 40 kyr, which are well documented by a large number of full vector data, not much information has been gained from records of absolute paleointensity. Recent development in the measurement of relative paleointensity from sediments has significantly changed this picture. Thirty-two records from different locations around the globe have now been integrated into the Sint-800 composite curve, which describes the variations of the GAD field intensity up to 800 kyr [*Guyodo and Valet*, 1999]. The resolution of the curve is limited by the subset of records with the poorest resolution and is not better than a few thousand years. Directional variations mostly reflect changes in dipole field intensity. The geomagnetic nature of the signal is strongly supported by its agreement with an independent estimate derived from the variation in ^{10}Be production during the past 200 kyr [*Frank et al.*, 1997; *Frank*, 2000].

[68] Several features inherent to these records can thus be considered as actual geomagnetic characteristics and thus fit category A. The first one is the very variable overall character of the field, which depicts a succession of large 20–60 kyr long oscillations and changes in amplitude, that can exceed a factor of 5. Within the limits imposed by the low resolution of the curve to document short-term variations we could not identify any periodicity in any time interval (long enough to allow proper spectral analysis). Thus dipolar field intensity has varied in a nonperiodical, erratic manner (periodicity would enter category C). Another interesting observation (see above) is that the weakest intensities always correspond with geomagnetic excursions.

3.3.2. Average Dipole Moment

[69] Another aspect is the time-averaged value of the dipole moment. Between 50 and 15 ka B.P. the number of volcanic records is large enough to average out nondipole components within 5 kyr windows. For this period the dipole moment has an average value of $(4.5 \pm 0.6) \times 10^{22}$ Am^2, nearly half the $(8.7 \pm 1.6) \times 10^{22}$ Am^2 value found over the past 10 kyr.

3.4. From 0.1 to 5 Myr

3.4.1. Average Dipole Moment

[70] For older periods, temporal coverage of the volcanic database is much poorer than with sediments. The calibration of Sint-800 has been performed by adjusting the younger part of the curve with the volcanic data set. Since Sint-800 incorporates a very large number of records, it is reasonable to consider that the differences in amplitude between distinct records have been averaged out. Thus the mean value of 6.0×10^{22} Am^2 with a standard deviation (s.d.) of 1.5 obtained for the past 800 kyr can be considered with good confidence. Unfortunately, this calculation cannot be extended as yet over a longer period by using sediments because of the absence of worldwide records. The number of volcanic records is also far too sparse and poorly constrained in time. In addition, a typical problem inherent in the volcanic data set is that dispersion increases with the number of results within each temporal window. Overall averages performed from volcanics tend to indicate higher values than those derived from calibrated sedimentary records. The 4 Myr long record from the Pacific ocean [*Valet and Meynadier*, 1993] has an average value of 4.0×10^{22} Am^2 (s.d. 1.9). This is almost half the 7.4×10^{22} Am^2 value (s.d. 4.3), which is often quoted [*Merrill et al.*, 1996] and in satisfactory agreement with the determination of 5.49×10^{22} Am^2 (s.d. 2.36) recently pub-

lished by *Juarez and Tauxe* [2000] after careful data selection. Note, however, that the volcanic database derived from this subset has very poor temporal coverage and large dispersion. Sedimentary records calibrated with volcanic data remain at the present time the most promising approach to extract the field mean intensity over a long period of time.

[71] In summary, it is likely that the average field intensity was about 6×10^{22} Am^2 during the past 0.8 Myr, which is 30% lower than the present-day field. Despite large uncertainties and poor geographical and temporal coverage of the existing data, there is no evidence as yet for any substantial long-term decrease (or increase) of mean field intensity for, say, the past 10 Myr. Of course, there is no reason why present-day field intensity should be considered as representative of the long-term average. One can thus consider that the mean intensity varies between 4 and 6×10^{22} Am^2 on this timescale (category A). This is reasonably well established for the past 1 Myr and is likely to hold also the past few million years.

3.4.2. Sawtooth Pattern of Intensity Variations

[72] No attention could be paid to field intensity changes across reversals until the recent emergence of long and well-dated sedimentary sequences. It is essential to detect whether specific conditions prevail before the field reverses and if there is any influence of reversals on dynamo operation. The most intriguing feature is the sawtooth shape of the records that features a long-term decrease of the field intensity during periods of stable polarity and a large and rapid increase immediately after the reversals. Such characteristics have been observed in several records from the Pacific and Indian Oceans [*Valet and Meynadier*, 1993; *Valet et al.*, 1994; *Thibal et al.*, 1995; *Verosub et al.*, 1996]. However, there are also records that do not exhibit such asymmetry between the pre- and postreversal periods [*Laj et al.*, 1996]. *Raisbeck et al.* [1994] attempted to observe the changes in ^{10}Be production that are modulated by the screening effect of the geomagnetic field against penetration of cosmic rays. However, the results were inconclusive, since they did not reveal any variation accompanying the large-intensity changes that prevail during reversals. Several alternative nongeomagnetic explanations have also been proposed [*Meynadier and Valet*, 1995; *Mazaud*, 1996; *Meynadier and Valet*, 1996; *Kok and Tauxe*, 1996a, 1996b], but they have not been convincingly demonstrated so far [*Meynadier and Valet*, 1996; *Meynadier et al.*, 1998]. Of course, this does not validate the geomagnetic origin of the sawtooth shape of relative paleointensity, which has thus been ranked among the B level observations. Detailed records of ^{10}Be production for this period would certainly be extremely useful. Another promising approach may be provided by detailed study of young marine magnetic anomalies: recent studies [*Gee et al.*, 1996; G. Pouliquen et al., A geomagnetic record over the last 4 million years from deep-tow magnetic anomaly profiles across the Central Indian Ridge, submitted to *Journal of Geophysical Research*, 2000] have shown that the oceanic crust formed at ridges can provide a reliable record of field intensity changes. The 4 Myr records obtained so far from sediments depict a field evolution dominated by a succession of highs and lows, similar to the past 800 kyr. The intensity lows coincide either with reversals or with excursions and/or short events. Thus it appears that the major characteristics of variations in field intensity (time constants, amplitude, averaged value) remained unchanged during the past four million years. In other words, no drastic changes in field intensity variations need to be expected in numerical dynamos over periods of a few million years.

3.4.3. The Last 5 Myr (Very Long Term Averages of Spherical Harmonic Coefficients)

[73] Paleomagnetists have assembled large directional databases from sediments and lava flows to search for persistent features in the geomagnetic field in the time window from hundreds of thousands to millions of years. Recent analyses [e.g., *Johnson and Constable*, 1997; *Kelly and Gubbins*, 1997; *Carlut and Courtillot*, 1998] of some of these bases using diverse inverse techniques come to somewhat different conclusions. All agree on the presence and robustness of a long-term axial quadrupole g_2^0, on the order of 5% of g_1^0. *Carlut and Courtillot* [1998] estimate that all other terms have long-term zero averages, compatible with the *Constable and Parker* [1988] Gaussian process. *Carlut et al.* [1999] find that the persistent axial quadrupole emerges from noise when the length of the available time series exceeds about 6 kyr. This would be why archeomagnetic data are not sufficient to isolate it. *Johnson and Constable* [1997] and *Kelly and Gubbins* [1997] find flux concentrations at the core surface under Canada and Siberia, similar to those observed over historical times and possibly related to columns parallel to the rotation axis in the core. They propose that the present pattern of secular variation is typical of the past millions of years. *Johnson and Constable* [1997] find significantly smaller higher-order and nonzonal terms than *Kelly and Gubbins* [1997], yet they do believe that they are significant. Therefore, given ongoing debates, in addition to g_1^0, only a small persistent g_2^0 (about 5% of g_1^0) can be considered as a robust constraint on the 10^5 to 10^6 year timescale for modeling purposes (category A). Values of other g_n^m and h_n^m remain controversial. However, they can all be safely considered to be below 1 μT (a rather weak category A constraint).

3.5. *Beyond 5 Myr*

3.5.1. Changes in Reversal Frequency

[74] There have been many studies dealing with statistics of the reversal sequence, some claiming that reversals form a Poissonian process, whereas others favor some form of determinism. The distinction is sometimes subtle. The succession of polarity intervals is closely approximated by a Poisson process or at least by a gamma process [*McFadden*, 1984; *McFadden and Merrill*, 1984, 1986, 1993; *McFadden et al.*, 1987; *Lutz and Watson*, 1988], assuming that short events have been missed in the reversal sequence. This implies that subsequent refinement of the magnetic polarity timescale could bring different conclusions. Thus the distribution of the length of polarity intervals cannot be considered as a major constraint to build an Earth-like dynamo.

[75] However, reversal frequency slowly changes between well-defined values. The long-term (hundreds of Myr) average is of the order of 1 reversal per million years, and the maximum value when averaged over a few million year does not seem to exceed 6 reversals per million years (category A).

3.5.2. Superchrons

[76] It could be tempting to take into consideration the occurrence of very long periods without reversals, such as the Cretaceous or Kiaman superchrons. These are periods when the field apparently did not reverse for tens of millons of years (though, see *Tarduno et al.* [1992], on the presence of events within the Cretaceous superchron). There is no doubt that these are extreme occurrences in the time series of geomagnetic chrons (category A). However, there is no agreement on how extreme they are. They might correspond to possible values with low probabilities; the small total number of available reversals precludes a firm statistical con-

clusion. They might correspond to times when the process responsible for geomagnetic reversals passed below a certain critical threshold, and reversals could start again only when that threshold was exceeded again. Various other analyses [e.g., *McFadden and Merrill*, 1995] suggested that these intervals would reflect cessation of the reversal process. In other words, the geodynamo would have a reversing and a nonreversing state. Alternatively, *Gallet and Hulot* [1997] proposed an explanation in terms of perturbations of the reversal process. Such changes may depend on factors directly connected with boundary conditions that affect the flow pattern within the core (such as the distribution of heterogeneities within the D″ zone of the lower mantle) and therefore may not be related to the intrinsic time constants of the geodynamo. Consequently, we feel that taking these superchrons as constraints for dynamo modeling is not yet warranted, particularly in the case of homogeneous conditions at the CMB.

[77] The same remark also holds for the hypothesis that there would be a long period of low dipole intensity between 120 and 180 Ma [*Perrin et al.*, 1991; *Courtillot and Besse*, 1987]. Note that the accumulation of new results suggests that the duration of this period (if ever it existed) has been considerably overestimated, so that no reliable estimate can be derived as long as a large independent set of data has not been obtained (category B).

[78] The observational constraints summarized in this section are listed in Table 2, together with what we estimate to be their robustness and reliability and whether they can actually be used in constraining numerical dynamo models.

4. Discussion and Conclusion

[79] A comparison between characteristics of numerical models and observations requires additional care, both when interpreting results from numerical models and when selecting relevant constraints. In this section, we first try to highlight the difficulties of such an exercise; we then attempt a comparison based on previously published models (when relevant data, i.e., numerical results, are available, which is far from being the general situation); finally, we propose a short list of the features that we feel should be presented with each numerical model in the future to allow a more efficient comparison with observational constraints.

[80] The scaling of quantities extracted from numerical models requires some attention before they can be compared with Earth values. Equations are usually solved in nondimensional form (to broaden potential use of the solutions and reduce the number of parameters involved). If all nondimensional numbers used in the computations were equal to their geophysical values, an obvious transformation would then allow one to scale numerical results and compare them with actual observations. It is clear at this point that this is not the case and that parameters used in the numerical simulations may significantly differ from geophysical estimates. In this case, the "scaling" needed to compare results with observations is not straightforward anymore and one can interpret the discrepancies in the nondimensional numbers in terms of different physical quantities. For example, if the Ekman number ($\nu/\Omega r_c^2$) used in the numerical model is larger than the Earth value, this could be interpreted as either due to a higher value of viscosity or to a smaller size of the core or to slower rotation of the Earth.

[81] Moreover, interpretations need to be consistent with the definition of all other nondimensional numbers relevant to the system. For instance, we may want to scale the results of a computation to compare the intensity of the

Table 2. Major Characteristics of the Geomagnetic Field

Timescale	*F*	*D, I*	Field Characteristic	Multiple Sites	Volcanics and Sediments	Different Authors	Controversy	Status
0.1–1 kyr	x	x	geomagnetic impulses	y	n.a.	y		A
		x	geomagnetic impulses during reversals	n	y?	n	y	C
	x	x	Pacific dipole window	y	n.a.	y	y	B
	x	x	global and systematic westward drift	n	n.a.		y	C
			columns and patches	y	n.a.	y		A
			dipole dominates	y	n.a.	y		A
1–10 kyr		x	global and systematic westward drift	n	n.a.	y	y	C
			excursions and reversals have similar durations shorter than 5 kyr	y	y	y		A
	x	x	excursions associated with intensity lows	y	y	y		A
	x	x	reversals dominated by nondipole components	y	y	y		A
	x	x	asymmetrical decay and recovery immediately before and after reversal	y	y	y		B
		x	preferred longitudes of transitional VGPs	y	n	y	y	B
		x	long-lived transitional states	y	n	y	y	B
10–100 kyr	x	x	20 to 60 kyr long oscillations of dipole moment	y	y	y		A
	x	x	periodic variations of dipole moment	n	n.a.	n	y	C
	x	x	dipole dominates	y	y	y		A
0.1–5 Myr	x		long-term decrease (saw-tooth) during stable polarity	y	y	y	y	B
	x		average dipole moment between 4 and 6 $\times$ 10^{22} A.m^2	y	y	y		A
		x	columns and patches	y	y	y	y	B
	x	x	dipole dominates	y	y	y		A
		x	$1000 < g_2^0 < 2000$ nT	y	y	y		A
		x	$g_n^m\ h_n^m < 1000$nT	y	y	y		A
>5 Myr		x	reversal frequency < 6/Myr	y	y	y		A
		x	superchrons	y	y	y		A
		x	Mezosoïc dipole low	y		y	y	B

Each line corresponds to a proposed characteristic feature of the field. These are arranged (as in the text) in the order of increasing timescales (first column). Next two columns indicate whether these involve measurements of field intensity (*F*) and/or direction (*D*,*I*). The characteristic feature (next column) is then evaluated through the answer to the following questions (four next columns): "Was it observed at multiple sites?", "Was it observed both in volcanics and sediments?", "Was it reported by different authors?", "Is it an ongoing controversy?". The last column is a direct consequence of the preceeding ones and suggests a classification of field characteristics into three categories: A (established), B (still discussed), and C (unlikely). Only category A will be retained for comparison with numerical models. (Abbreviations used are as follows: n.a., not applicable; y, yes; n, no.)

induced magnetic field with that observed for the Earth. Field intensity is usually based on rotation rate and magnetic diffusivity. If we want to interpret our results using Earth values for both quantities, we must consider all nondimensional numbers using these values. Let us assume that we describe our system with three nondimensional numbers: the Ekman, Prandtl, and magnetic Prandtl numbers. Because rotation is now fixed, viscosity is given by the value of the Ekman number used in the computation. Because magnetic diffusivity is also fixed (to determine field intensity), thermal diffusivity is given by the value of the magnetic Prandtl number used in the computation. As a consequence, because viscosity and thermal diffusivity are given, there is only one consistent (imposed) value for the Prandtl number. When inertia is dropped from the computation, the Prandtl number is ‘‘free’’ (although it should be kept small to justify the simplification); but if such is not the case, its value is already independently fixed in the computation, leading to a possible inconsistency in interpreting the results (see discussions by *Glatzmaier and Roberts* [1995a, 1995b] and *Kuang and Bloxham* [1998]).

[82] Timescale constraints are particularly difficult to take into account with numerical models. The reason for this difficulty is again a consequence of the parameters used in the simulations (that can be readily interpreted as ratios of characteristic timescales). Numerical results are often published using ‘‘year’’ as unit of time. By year, authors generally mean ‘‘magnetic year,’’ i.e., the timescale constructed using the dipole decay time, or magnetic diffusion timescale in the Earth's core (roughly 20 kyr). A consequence is that the ratio of the rotation timescale (the day) to the magnetic timescale (directly proportional to the magnetic Ekman number defined above) is much larger than it should be. One can count much less than a day per year in most models.

[83] The fact that numerical models use parameters far from geophysical estimates must be taken into account when establishing which observational constraints can effectively be used for comparison with numerical predictions. For example, it is difficult to directly incorporate constraints based on timescales in numerical results, for the reasons stated above, and we will therefore prefer to present them in terms of ratios of characteristic times or scale them using the dipole diffusion time.

[84] In section 2 on generally accepted facts versus ongoing controversies, we propose a classification of magnetic features (i.e., geomagnetic and paleomagnetic ‘‘observations’’; see Table 2) in three main categories: those that are established independently by different teams and essentially agreed on by most authors and that can safely be considered as robust constraints for numerical dynamo modeling (A); those for which the extent of controversy or disagreement is such that they should not be used pending confirmation (B); and those that are most unlikely (C). Some features that had been proposed in the past can clearly now be rejected. The decision to place data into one category should be as objective as possible, and we have outlined a few criteria that we believe should be prerequisites. However, there is clearly room left for ambiguity and personal bias (or preference), and we do not claim to be free of such limitations.

[85] A sensible choice appears to be to retain only the first category of observations (A) as constraints for numerical models. We believe features placed in this category to be free of controversies. Following the presentation adopted in Table 2 and starting with the 0.1–1 kyr scale, the existence of geomagnetic impulses should clearly be retained as a strong and well-established constraint. Features such as the Pacific dipole window low, columns and patches, or even the ratio g_2^0/g_1^0 (as estimated

over this timescale) need confirmation from longer timescale observations. On the 1–10 kyr scale the existence of excursions and reversals and the fact that both have similar durations (shorter than half of the dipole decay time) clearly is a robust and efficient constraint and so are as well the existence of intensity lows associated with excursions and the nondipolar character of the field during reversals. On the 10–100 kyr scale, oscillations of dipole intensity with time constants of the order of 1 to 3 dipole decay times is a robust constraint. On the 0.1–5 Myr scale the average intensity of the field between 4×10^{22} and 6×10^{22} Am^2 is well established as well as the ratio g_2^0/g_1^0, close to 5%. From longer scales the only robust observations are the maximum reversal frequency ($\sim$5 Myr^{-1}) and the existence of superchrons. Because such timescales are comparable with the characteristic timescales of mantle convection and evolution, there is a possibility that superchrons are not only due to some internal mechanism of the dynamo but are linked to boundary conditions imposed by the mantle on the CMB. This issue is still the focus of an ongoing debate, and it should not, at this stage, be retained as a robust constraint. This prevents us from establishing a meaningful minimum reversal frequency.

[86] In summary, according to the above discussion, there appear to be currently 10 and only 10 well-established, robust constraints for dynamo models: (1) Dipolar character of the field; (2) existence of geomagnetic impulses or jerks with a short time scale (less than 10^{-4} times of the dipole decay time); (3) existence of excursions on timescales of the order of one tenth to one half the dipole decay time; (4) excursions associated with intensity lows; (5) existence of reversals, with a duration comparable to that of excursions, and occurring at a rate of at most 1 per 10 dipole decay times; (6) nondipolar character of the field during reversals; (7) nonperiodic long-term oscillations of dipole intensity (1 to 3 dipole decay times) with maximum amplitude 5×10^{22} Am^2; (8) long-term average dipole intensity between 4×10^{22} and 6×10^{22} Am^2 (when averaged over more than ten dipole decay times); (9) long-term average value of g_2^0 between 1 and 2 μT (i.e., close to 5% of that of g_1^0), when averaged over more than 10 dipole decay times; and (10) long-term average values of all higher-degree terms lower than 1 μT, when averaged over more than 10 dipole decay times.

[87] Clearly, there is no proof that any magnetic field satisfying all these requirement will necessarily correspond to the same generation mechanism as that actually operating within the Earth's core; comparisons with asymptotic studies constitute the only robust confirmation. Also, one could imagine a numerical dynamo model could be based on the correct physical processes occurring in the Earth's core but not satisfy some constraints, simply because some parameter values would not be appropriate (and as a result, some outputs would not scale properly). Nevertheless, this appears to be a minimal set of well-established necessary conditions for a numerical model to claim Earth-like behavior.

[88] Most numerical models were integrated over a few dipole diffusion times (about 3) to attest self-sustained dynamo action. However, it is clear from section 2 that to test for Earth-like behavior, relevant features and numerical predictions should be averaged over at least 10 dipole diffusion times.

[89] Comparing the above list with Table 1, it is striking that apart from dipole dominancy, not many features are common to both. Many numerical models appear to concentrate on rather controversial features that cannot be considered as indisputably Earth-like (such as westward drift of the field) but also outputs from these models were not formatted in a way that would allow direct comparison with ob-

servational constraints. No impulses (jerks) seem to have been uncovered in numerical models. Excursions have only been reported by the recent work of *Glatzmaier et al.* [1999]. When models presented reversals (e.g., Glatzmaier and Roberts; Kageyama et al.; Kitauchi and Kida), the maximum reversal frequency of about 1 per 10 dipole diffusion times (τ_d) can be used directly as a constraint. Clearly, the *Glatzmaier and Roberts* [1995a, 1995b] model (one reversal in $\sim$3 τ_d) is not testable, because it is available over too short a time period. *Glatzmaier et al.* [1999] (two reversals in $\sim$15 τ_d) and *Kageyama et al.* [1999] (one reversal in $\sim$50 τ_d) are on the right side according to this test, although time integration is obviously too short to establish a reversal frequency. *Kitauchi and Kida* [1998] presented a polarity time series with some 30 reversals in $\sim$100 τ_d, long enough to suggest that the frequency value is significantly above the requirement for Earth-like behavior.

[90] Figure 1e presents dipole moment fluctuations as observed over the last 3 million years, i.e., approximately 150 dipole decay times [*Valet and Meynadier*, 1993] together with the SINT-800 paleointensity stack over the last 800 kyr [*Guyodo and Valet*, 1999]. These are compared to the numerical results published by three groups: *Kida and Kitauchi* [1998], *Kageyama et al.* [1999], and *Glatzmaier et al.* [1999]. These results cover a long enough time interval to allow comparison with the Earth observations. In addition, a histogram of the dipole moment values in the corresponding time series is shown (normalized to unit area, except for Figure 1c and the red part of 1e (SINT-800) for which it was normalized to one half). Direct inspection of Figure 1 shows that the *Kida and Kitauchi* [1998] curve (Figure 1d) does not feature the alternance of polarities that characterizes chrons (i.e., fails to predict a symmetric bimodal histogram). The *Glatzmaier et al.* [1999] tomographic model (Figure 1b) has too much high frequency oscillations, i.e., secular variation and excursions (as noted by the authors), leading to a "flatter" histogram. The results of *Kageyama et al.* [1999] (Figure 1c) are closer to traditional ideas of what bimodal fluctuations should be like. However, there is a lack of lower dipole moment values, which is not observed in the case of Earth, and more importantly the large contribution of the octupole (not shown in Figure 1) is not at all Earth-like. Finally, only the results from the *Glatzmaier et al.* [1999] homogeneous model (Figure 1a) appear acceptable. Indeed, they look very much like a portion of the observed record near 90–120 τ_d, but, of course, the integration time (15 τ_d) is still too short to allow a robust comparison.

[91] Checking constraints concerning the duration of reversals would require comparison of plots of paleomagnetic directions with numerical results. Only *Glatzmaier et al.* [1999] provide such plots for reversals. Both reversals reported for the homogeneous model were appropriately fast; however, the second reversal in the tomographic model was apparently too long. It would be useful if other models provided such information. The nondipolar character of the field during reversals would also need documentation (such as field spectra during reversals, only provided by *Glatzmaier and Roberts* [1995a, 1995b]).

[92] Finally, when trying to constrain the g_n^m, h_n^m coefficients, it is clear that much relevant information is usually absent from most papers presenting numerical dynamos. These constraints are readily converted into constraints on the energy spectrum of the average field ($W_n = (n+1)\sum_{m=0}^{n}({g_n^m}^2 + {h_n^m}^2)$; [*Lowes*, 1974]) at the core-mantle boundary over a long enough period of time (above 10 τ_d). Spectra have been published for a few numerical models. Unfortunately, these spectra were often instantaneous or averaged over only a short period of time (the

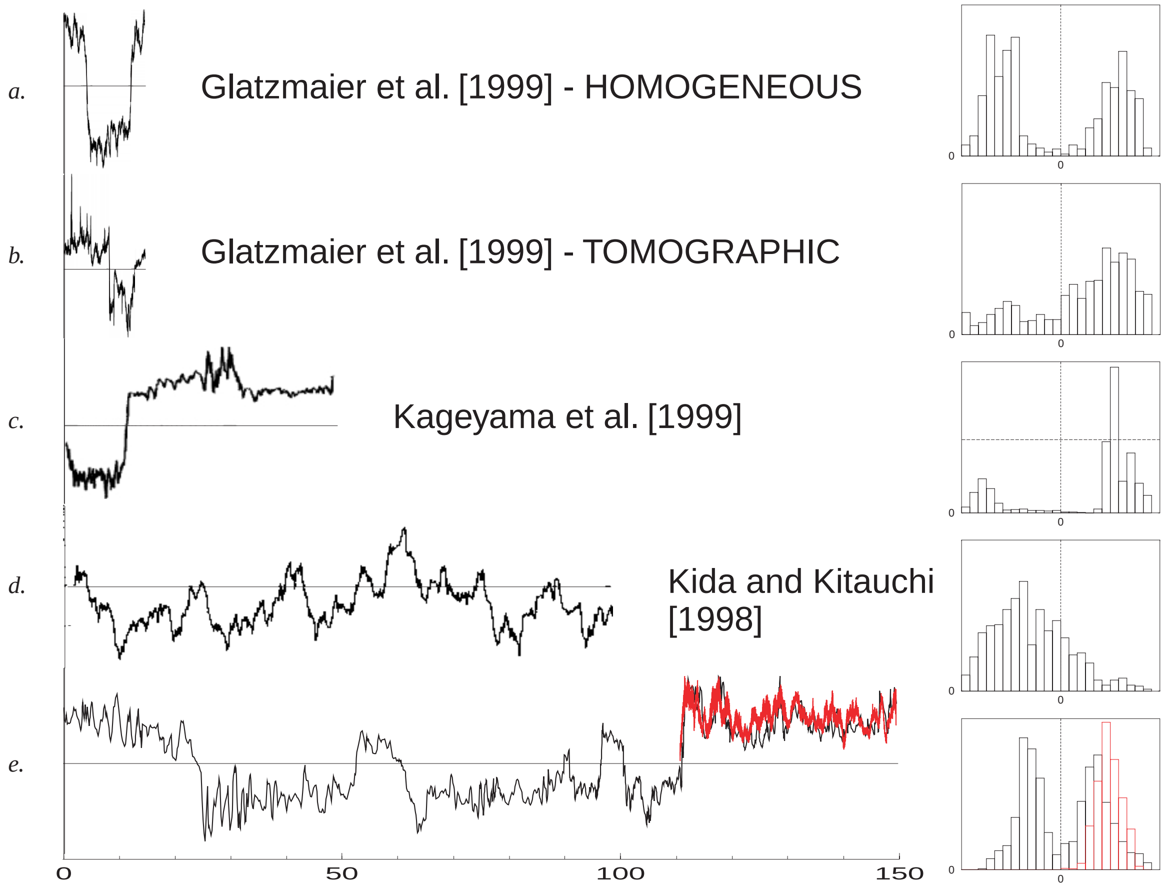

Figure 1. Time evolution of the dipole moment (relative values) for numerical dynamo models with integration times longer than 10 dipole diffusion times (base lines correspond to zero). The x coordinate has been scaled using dipole diffusion time τ_d as unit. (a, b) time evolution of the dipole moment as published by *Glatzmaier et al.* [1999]; (c) same for *Kageyama et al.* [1999]; and (d) same for Kida and Kitauchi, [1998]. It should be stressed that the length of time integration does not reflect the amount of modeling and computational work, since the parameter regime is very different from one model to the other (see Table 1). (e) Dipole moment measured for the Earth over the last 3 Myr [*Valet and Meynadier*, 1993] using the same convention for time (present is at the far right), the red graph corresponding to the SINT-800 paleointensity stack over the last 800 kyr [*Guyodo and Valet*, 1999]. The histograms on the right show the distribution of the dipole moments for each simulation and for the observations (see text).

only exception being the model of Glatzmaier and Roberts in the work of *Coe et al.* [2000]); also, these spectra have been plotted at different elevations (Earth surface, core-mantle boundary, or depth corresponding to a flat spectrum).

[93] We show all available spectra from those papers, reduced to a common elevation (outer bounding sphere, corresponding to the core-mantle boundary) on Figure 2. Spectra from Glatzmaier and Roberts (GR95-1995b; GR96b-1996b) correspond to instantaneous spectra (three spectra for GR95, one for GR96b). The spectra from Glatzmaier and Roberts (GR96a-1996a) correspond to a series of values over 300 year (the green bars correspond to the intervals

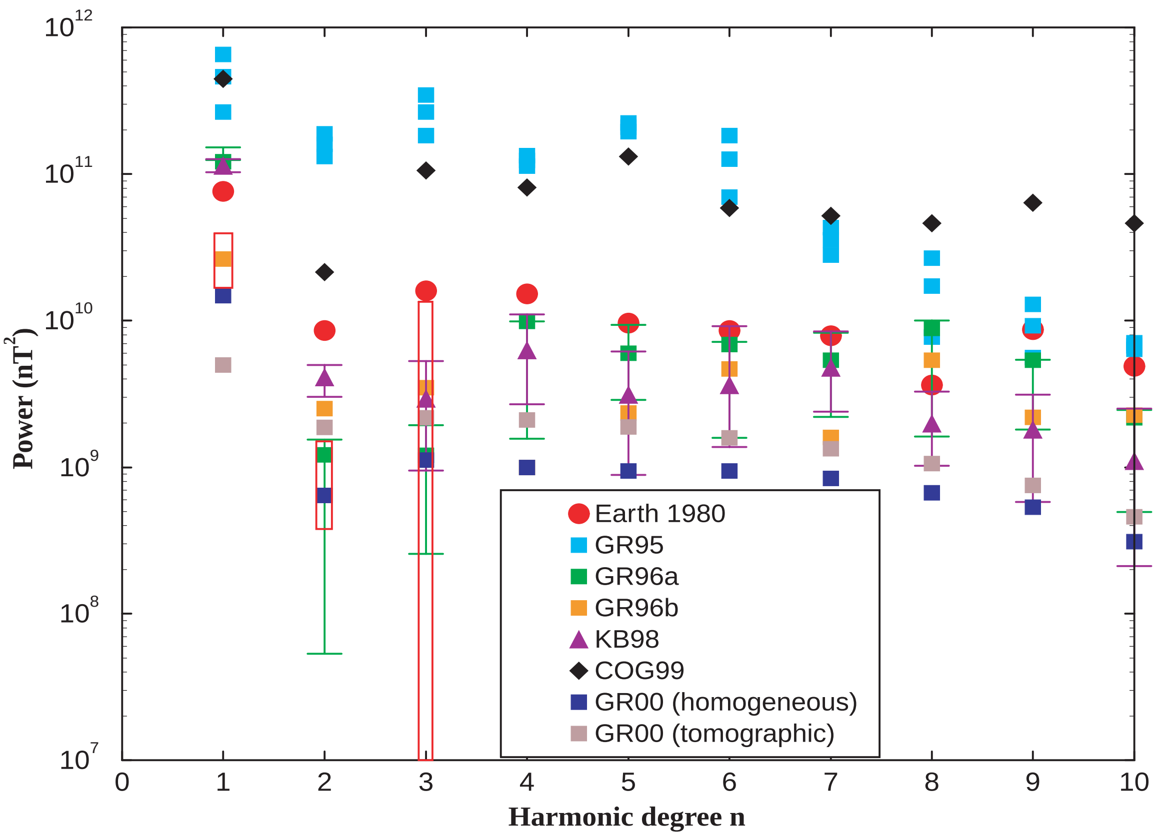

Figure 2. Published power spectra (w_n=$(n+1)\sum_{m=1}^{n}(g_n^{m2}+h_n^{m2})$ [after *Lowes*, 1974] for numerical dynamo models. Spectra have been converted to a common unit and reduced to a common depth (core-mantle boundary). Red dots represent the 1980 instantaneous spectrum for the Earth; red boxes represent the intervals for the average field as inferred from paleomagnetic data (see text). Symbols for numerical models are given in the inset; codes and references are given in the text (Note that not all model spectra are averaged over the same length of time; see text).

over which these range). Kuang and Bloxham (KB98-1998) display a spectrum averaged over 1500 years shown with its variance (bars). They have, unfortunately, not published a spectrum averaged over the total duration ($\sim$2 τ_d) of their simulation. COG98 is a time average (probably over 3 τ_d) spectrum from *Christensen et al.* [1998]. Two spectra from *Coe et al.* [2000] were averaged over 15 τ_d. They correspond to the Glatzmaier and Roberts model (GR00; one with homogeneous thermal boundary conditions, the other with an imposed heat flux inspired by tomographic seismological studies). The 1980 field model (constructed with MAGSAT data and observatories) is shown as red dots. This spectrum is used for comparison by most authors (with the exception of *Kuang and Bloxham* [1998], who used the *Bloxham and Jackson* [1992] model for comparison). It is clear from section 2 that the field spectrum of the Earth in 1980 or over 300 years cannot be used as a valid constraint. On the other hand, the first three components of the spectrum are well constrained when averaged over more than 10 τ_d (using the knowledge of the corresponding g_n^m, h_n^m coefficients; see Figure 2). The corresponding bounds are also indicated, for information, on Figure 2. Clearly, one should not use these bounds to test all spectra displayed on Figure 2, as they do not correspond to fields averaged over more than 10 τ_d. The only exceptions are the two spectra presented by *Coe et al.* [2000] that correspond to the Glatzmaier and Roberts model.

[94] Figure 2 allows us to conclude that models such as GR95 and COG99, though averaged over only a limited duration, are far more energetic at all degrees than the actual geomagnetic field. On the other hand, the "homogeneous" model (GR00) of *Coe et al.* [2000], averaged over 15 τ_d, is rather close to the best information we have about the paleomagnetic field spectrum, whereas the "tomographic" model has a much too weak dipole. The GR00 homogeneous model [*Glatzmaier et al.*, 1999; *Coe et al.*, 2000] appears to fit rather well the contraints summarized in Figure 1 and 2, and it is the only one so far to come close to such success (the Kuang and Bloxham KB98, though possibly promising, is averaged over only a fraction of τ_d, which is too short for a robust conclusion). However, we recall that the GR00 model suffers, like the other models, from unrealistic values of the Ekman and magnetic Prandtl numbers. Puzzling features include a large Elsasser number ($\sim$100), contrary to expected (but not demonstrated) Earth values which are on the order of one. Also, magnetic induction and fluid motion is localized mostly within the inner part of the cylinder tangent to the solid inner core, whereas models based on secular variation have most of the motion outside of this cylinder. Finally, the value of the mean dipole term in the spectrum could be ill resolved, since the authors state that their high-resolution tests produce field intensities that can be greater than Earth values [*Glatzmaier et al.*, 1999].

[95] We may conclude this review with some perspectives for forthcoming studies. We suggest that future numerical models wanting to test for Earth-like behavior should clearly present quantities averaged over the full integration time, which should (if possible) be at least 10 τ_d. From the observational side it would, of course, be most useful to establish the validity of some observations in category B, to propose additional constraints for dynamo models. Indeed, *Carlut et al.* [2000] propose that the mean values derived from paleomagnetic data over half a million years and more could be attained for timescales only slightly in excess of archeomagnetic values, of the order of 10 kyr, i.e., not 10 but only one half a dipole diffusion time.

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