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The dynamo effect/L'effet dynamo

Geomagnetism and the dynamo: where do we stand?

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Abstract

We review recent developments both in the observation of the Earth's magnetic field (from the short, human life timescale, to the long, geological timescale) and in the modelling of its origin (using the numerical or the experimental approach). We attempt a confrontation of these results, coming from very different fields, and show how, when combined, they can yield a better understanding of the Earth's core dynamics. We assume prior knowledge of dynamo theory, but not of geophysics. ***To cite this article: E. Dormy, J.-L. Le Mouël, C. R. Physique 9 (2008).***

Résumé

Le point sur le géomagnétisme et la dynamo. Nous passons en revue les développements récents qu'ont connus à la fois l'observation du champ magnétique terrestre principal (de l'échelle du temps de la vie humaine à celle des temps géologiques) et la modélisation de ses sources, recourant au calcul numérique et à l'expérience. Nous rapprochons ces avancées et montrons comment, lorsqu'on les combine, elles peuvent permettre de mieux comprendre la dynamique du noyau fluide terrestre. Nous supposons une connaissance préalable de la théorie dynamo, mais pas du géomagnétisme. ***Pour citer cet article : E. Dormy, J.-L. Le Mouël, C. R. Physique 9 (2008).***

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1. Introduction

The so-called main magnetic field of the Earth has for origin a self-excited dynamo working in the metallic, liquid core of the planet, good conductor of heat and electricity. The core material is an alloy Fe–Ni with the addition, in the liquid phase, of a light element in the proportion of 10% (Si, S, O). The solid inner core radius is, today, 1200 km. The power requested to maintain the dynamo is of the order of 10^{11} watts. The Earth is cooling; the heat which escapes from the core at the core-mantle boundary (CMB) is of the order of 5×10^{12} watts, i.e. fifty times larger. Furthermore, as the core is cooling, the inner core increases by successive additions of layers which crystallized at the inner core boundary (ICB); the light elements of those layers do not enter the solid phase, they are released at the ICB, providing

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a gravitational energy of 10^{12} watts and buoyancy forces immediately usable for the convection (all these figures are orders of magnitude).

So, it is generally believed today that the motions in the core, which generate the magnetic field, are convective motions, thermal or compositional (also called chemical). Thermal and chemical forcings are characterized in dynamo modelling (Section 4) by the Rayleigh number. In numerical codes the value of these numbers is increased to pass the convective bifurcation, then the dynamo bifurcations. The corresponding values are, of course, passed in the Earth's case. Note, however, that another mechanism is still under consideration, despite ups and downs in its favor: the luni-solar precession might induce in the fluid core motions complex enough for the anti-dynamo theorems to be bypassed (which is, of course, not the case for the classical laminar motion associated with the precession). The historic experiments of Malkus [1], and more recent ones by Vanyo [2] indicate the existence of turbulent motions. Estimates of the energy dissipated by precession induced motions have varied along time, sometimes claimed to be large enough, sometimes too small (e.g. Loper [3]; Rochester [4]). Recent numerical computations (Tilgner [5]) suggest that these motions might provide enough power for dynamo action.

2. Geomagnetic observations

An excellent description of the distribution of the main field and of its temporal (secular) variation is available over the last two decades near the Earth's surface. The Oersted and Champ magnetic satellites, launched respectively in 1999 and 2000, still working, and Swarm, to be launched in 2010, provide and will provide a global figure of the field with an accuracy of the order of 1 nT, and a time resolution of a few weeks (Sabaka et al. [6], Maus et al. [7], Olsen et al. [8], Hulot [9]). Before the era of satellites, field models relied on data which were much more poorly distributed in time and space, and required long time averages (on the order of a few to one year) around the "epoch" attributed to the model. The array of ground magnetic observatories, not dense enough and rather poorly distributed, with, in particular, large gaps over oceans, but providing continuous measurements in time, played a major part in the description of the field. They are still nowadays of prime importance in geomagnetic studies. The first magnetic observatories were founded in 1832. More or less continuous observations of the direction of the field have been made since the middle of the XVIIth Century for the declination, since the middle of the XVIIIth Century for the inclination, in rare European sites, London and Paris being the most ancient. Discontinuous observations of the field – only of its direction before 1838, date of the first measurement of the intensity by Gauss – at different points of the surface of the Earth, made by travellers, explorers, navigators, are available. Although those points are highly irregularly distributed at the Earth's surface, planetary models have been drawn from these data; the oldest one is for the year 1590 (Jackson et al. [10]).

However, those few hundred years of observations are not enough for a comprehensive description of the dynamics of the geomagnetic field and dynamo, which contains much larger time constants. To go further backwards in time, one has to resort to indirect measurements of the fossil field, i.e. to the magnetization of baked clays (for the last millennia), volcanic and sedimentary rocks for longer recordings. The basic idea is quite simple: the direction of the magnetization of the rock sample is the direction of the magnetic field at the place of the sample and at the time when the volcanic rock (or baked clay) cooled down or the sediment was deposited. The intensity of the ancient field can also be retrieved using different techniques, most of them primarily due to E. Thellier [11–13]. The recovery and study of the field during historical times (a few millennia BC, relying mainly on the thermoremanent magnetization of baked clays and volcanic rocks) constitute the domain of "archeomagnetism". The recovery and study of the geomagnetic field during the immense geological times constitute the domain of "paleomagnetism". The time resolution of the observations obviously decreases as older ages are being considered.

3. Typical timescales of the geomagnetic field variation

The main geomagnetic field presents an extraordinarily large range of time constants, from a year or less (the screening of the slightly conducting mantle prevents possible shorter periods to be observed), up to tens of millions of years. We will rapidly review these, from short to long.

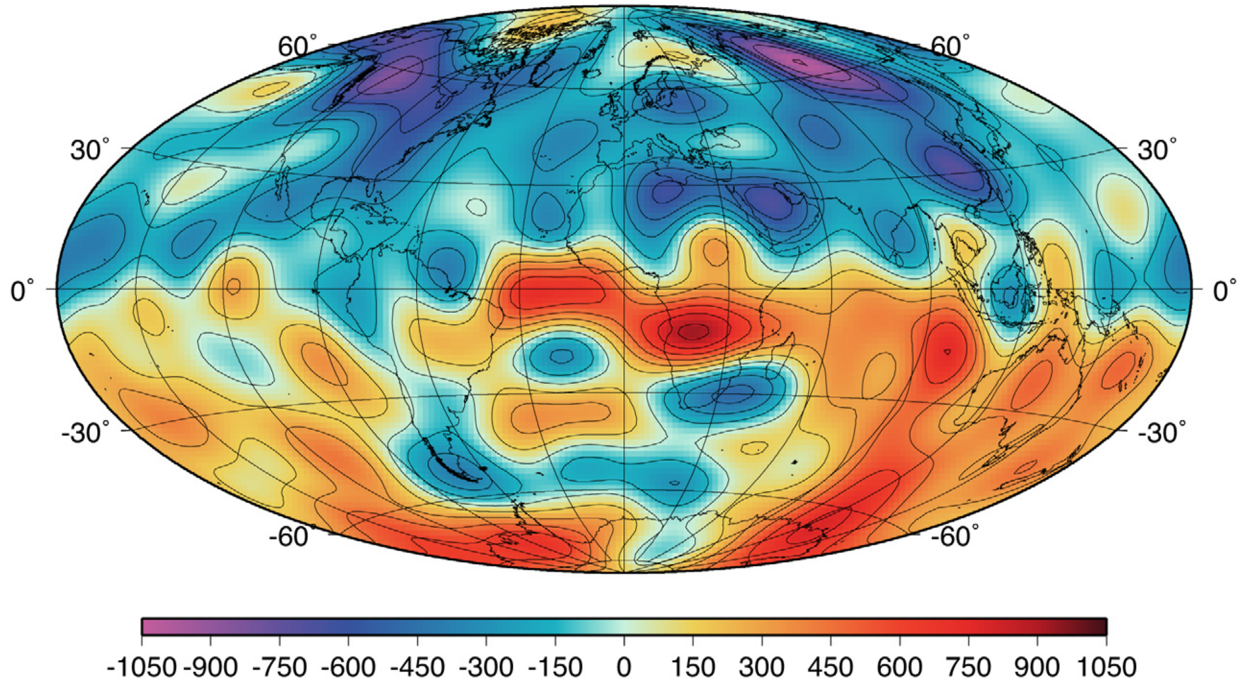

Fig. 1. Radial component of the Earth magnetic field (in mT) at the surface of the Earth's core in 2000 from Oersted initial field model (Olsen et al. [45]). Continents are represented for convenience, but the map represents the field some 3000 km below our feet.

3.1. Time constants of secular variation

The so-called secular variation was discovered in 1635, by Gellibrand. It is described by direct measurements, briefly recapitulated above, and by archeomagnetic data. The figure of the geomagnetic field evolves continuously, with time constants varying with the degree of the harmonic component it is the sum of (this decomposition is a mathematical image): 200 years for the multipolar components, 500 years for the equatorial dipolar components, 2000 years for the axial dipolar component (these figures are only orders of magnitude; they are obtained through rather crude extrapolations). We just mentioned that the secular variation was continuous; this is not fully exact: the continuous evolution of the field is marked by short events which are acceleration jumps, completed in a year or less, distant from a few tens of years from one another, often of worldwide extension (e.g. Courtillot, Ducruix and Le Mouël [14]). These events, called magnetic jerks or magnetic impulses, are still poorly understood.

Archeomagnetic data allow us to measure in some places the long time constants of secular variation and compare them with the above estimates. At a given site, e.g. Paris, declination and inclination of the field exhibit strong variations of the order of a few tens of degrees, with time constants of the order of a thousand years (Gallet et al. [15]). Archeomagnetic data are not of worldwide extension (measurements are time consuming). Nevertheless, there is now a reasonable density of observations in Europe, Middle East and North America; Siberia and Asia are scarcely but rather uniformly covered. There are much less data available in the Southern Hemisphere. This allows, in some regions, spatio-temporal studies of the last thousands of years of the geomagnetic field.

3.2. The geomagnetic field in geological times: Chrons and reversals

Let us recall the most salient features, most important properties, of the geomagnetic field. At a given time, today for example, the dipolar component of the main field is dominant at the Earth's surface. The dominance of the dipole field is not only due to the geometrical attenuation; it still prevails at the core-mantle boundary (see Fig. 1). This is no longer true during reversals.

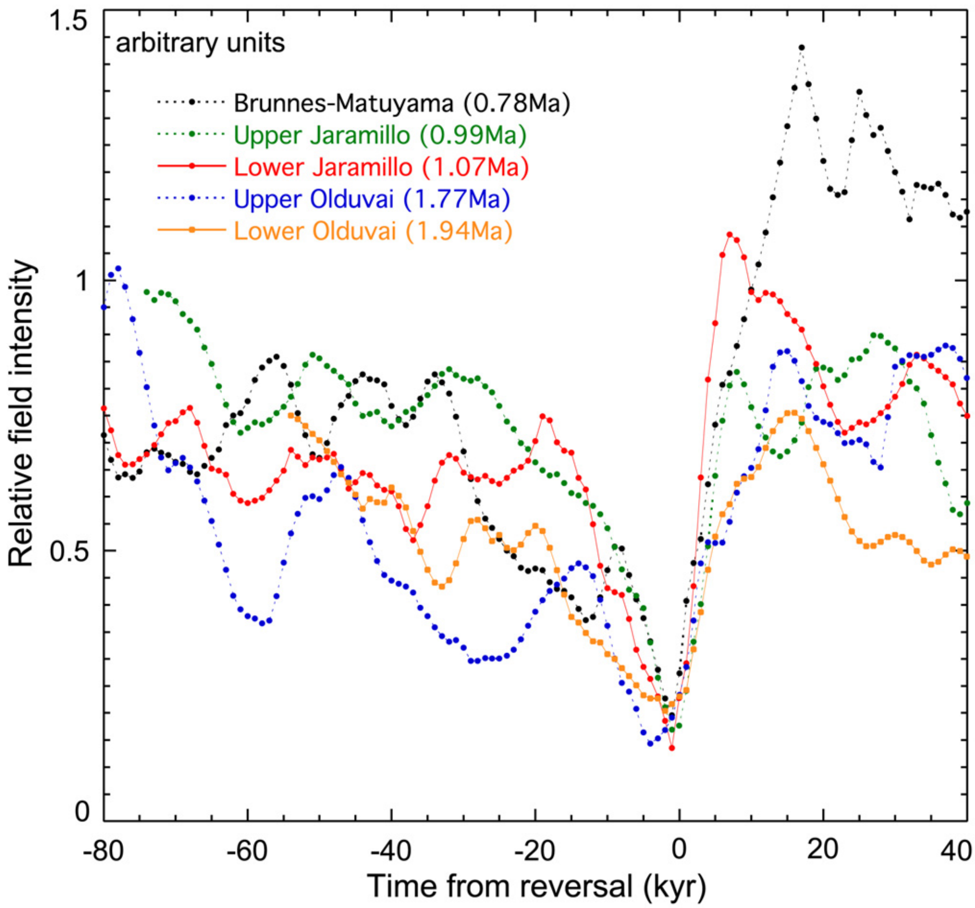

Fig. 2. Field intensity variation across reversals during the last 2 million years as reconstructed by Valet et al. [17].

Taken in average over a few thousands of years, the magnetic field is, outside the core, the field of an axial centered dipole (located at the Earth's center, aligned with the rotation axis). These are of two prominent results of paleomagnetism, from which we infer two major consequences:

– the planet rotation, through the Coriolis force, is an essential ingredient of the geodynamo;
– the container, i.e. the convective mantle, does not impose a durable lack of axial symmetry to the geodynamo field.

During geological times the dipolar field – unavoidable time and space averaging smooths out multipolar components; an exception is again to be made during reversals – presents long time intervals (chrons) during which its polarity is constant, i.e. the dipole is oriented N.S. (direct polarity) or S.N. (inverse polarity), interrupted by reversals which are rapid transitions from a polarity to the other. The duration of chrons during the last 80 millions years, varies from 100 000 to 10^6 years (Fig. 3). Its average over the last 250 millions of years is $\sim 1.5 \times 10^6$ years, excluding superchrons.

The sequence of periods of constant polarity, or chrons, marked by the reversals, includes indeed chrons of exceptional duration, or superchrons (Fig. 3) whose durations are of a few tens of million of years. Are they extreme events of a standing statistics, i.e., occurring without a change in the boundary conditions or forcing of the dynamo mechanism, or do they require such a change? It is quite possible to imagine scenarios in which the heat flow at the CMB varies, with time constants of the order of 10^7 years, through delayed exchanges between the upper layers of the core and the D″ layer at the base of the mantle. The scenarios, of course not firmly buttressed, would have huge

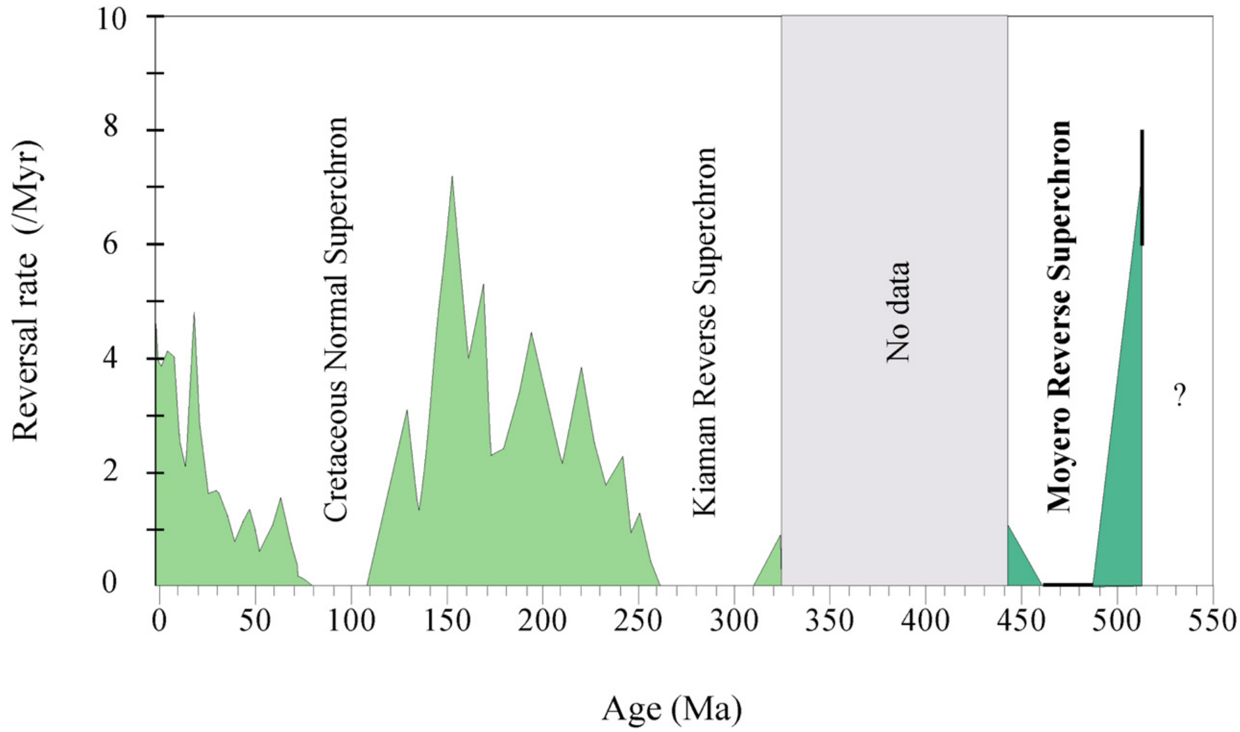

Fig. 3. Variation in geomagnetic reversal frequency over some 5×10^8 years as reconstructed from rock magnetization (Pavlov and Gallet [46]).

geodynamic implications (e.g. Courtilllot and Olson [16]). Note, however, that the rather regular variation, in geologic times, of the frequency of the reversals (Fig. 3) requests a progressive change of the forcing, often included in the scenarios mentioned. However, the whole question is difficult; one hesitates on the effects of the invoked change, even on the sign of this effect: e.g., if the heat flow escaping from the core increases, does the reversal frequency increase or decrease?

The intensity of the dipole field reveals quite a remarkable dissymmetric evolution in a neighbourhood of a reversal, illustrated by Fig. 2, in which the origin of time is taken as the time of the reversal, for each curve. The decay of the magnetic field is announced some 80 000 years in advance, while, after the reversal, its intensity is rapidly re-established at a higher level than that which prevailed before the reversal (Valet et al. [17]).

4. During reversals and excursions

We come now to the reversals. Their duration is a few thousand years. The axial dipole moment decreases, and becomes weak compared to what it was before the event. Then, at every site, the temporal variations of the field, which is dominantly multipolar, are large, rapid, erratic. It is the transitional state. After a while the dipole increases again and gets stronger than it used to be before the event. If the new polarity is opposed to the previous one, it is a reversal; if the dipole resumes its previous polarity, the event is an excursion.

This is the basic scenario. But the reversals may be more complex and display some kind of oscillatory behavior (Valet and Herrero-Bervera [18]; Narteau et al. [19]). The dipole makes a first attempt to establish itself in its new polarity, fails, goes back to the transitional state characterized by a strong variability of the field direction. Eventually, it succeeds in retrieving a strong intensity, with the overshoot pointed out above (Fig. 2). It is very interesting, and fundamental for the understanding the dynamo, to study the reversals: it is during these events that we see the magnetic field being constructed. But the requested field work to reach a fair description of the event is long and difficult; only a dozen of reversal studies exist to this day.

Table 1
Typical diffusivity parameters of the geodynamo

	Diffusivities	Diffusive timescale	Ratios
Kinematic viscosity	$\nu \sim 10^{-6}$	$L^2/\nu \sim 10^{19}$ s	$Pr = \nu/\kappa \sim 10^{-1}$
Thermal diffusivity	$\kappa \sim 10^{-5}$	$L^2/\kappa \sim 10^{18}$ s	$q = \kappa/\eta \sim 10^{-5}$
Magnetic diffusivity	$\eta \sim 1$	$L^2/\eta \sim 10\,000$ years	$Pm = \nu/\eta \sim 10^{-6}$

5. Geodynamo modelling

5.1. Direct numerical simulations

For many years the study of the Earth's magnetic field was restricted to observation and theoretical studies. The situation changed dramatically since 1995 with the rapid development of many numerical models able to produce self-excited dynamo fields with properties similar in many respects to the Earth's magnetic field. Due to computational limitations, these models always rely on a parameters regime which is very remote from the geophysically relevant one. In fact some of the most relevant parameters such as the *Ekman number* (characterizing the ratio of the length of the day to the viscous decay time) or the magnetic *Prandtl number* (measuring the ratio of the ohmic to viscous decay time) are off of their geophysical values by factor in excess of a million. Despite these limitations, numerical models have been able to produce dominantly dipolar fields with occasional polarity reversals! This, very surprising, result could suggest a strong robustness of the main characteristics of dynamo fields to model alteration.

There are several ways to present the situation. Dynamo MHD equations reveal characteristic times directly related to transport parameters whose values for the Earth's core are given in Table 1, where L is a typical lengthscale (here the radius of the core), ν the kinematic diffusivity, κ the thermal diffusivity, η the magnetic diffusivity.

Let us also give the value of the Ekman number:

$$E = \nu/\Omega L^2 \sim 10^{-15} \quad (L = 3400\ \text{km},\ \Omega = 2\pi/T,\ T = 1\ \text{day})$$

One of the difficulties comes from the weak molecular viscosity of the liquid iron in the core, which is close to that of water (at usual temperature and pressure). Viscous forces are important only in boundary layers, CMB and ICB, or thin shear layers in the body of the flow. On the contrary, in numerical models, the viscous force is essential to the stability of the computation. As a consequence, in terms of Ekman number, numerical computations can be performed only for E values $> 10^{-6}$, i.e. 10^9 larger than the value relevant to the Earth. In the same way, it is very difficult to maintain in the numerical simulations a dynamo action with a magnetic diffusion as efficient as in the real core. Practically, it is required that magnetic and thermal diffusivities be of comparable magnitude (and close to the kinematic viscosity), in such a way that, in those convective dynamo models, the spatial spectral properties of the magnetic and flow field be comparable. Computations are then performed with order 1 value of the Roberts number $q = \kappa/\eta$. As seen in Table 1, the value of q is 10^{-5} for the Earth's core. That, of course, is not without consequences. For example, the classical asymptotic regime, the strong field regime introduced by P.H. Roberts, which should be relevant for the terrestrial dynamo, cannot be attained by numerical dynamos (this regime, in which Laplace–Lorentz force and Coriolis force equilibrate, is also called the magnetostrophic regime).

Nevertheless, from the first computations of Glatzmaier and Roberts, it has been noticed that numerical dynamos do present Earth like features: dipolar structure, secular variation, sometimes even reversals. Direct three-dimensional models of the geodynamo have achieved dramatic improvement over the last few years. Models now cover a wider region of the parameter space and their dynamics is better understood. Their relevance to the geodynamo problem however still remains questionable (Dormy et al. [20]).

Because of the strong computational resources necessary to perform a direct integration of the full three-dimensional MHD equations, only a limited subset of the parameters space is presently available, despite the use of some of the world largest computers. The first essential result obtained since the review of Dormy et al. [20] is the "phase diagram" produced by Kutzner and Christensen [21]. In this pioneering work, the authors cover a range of *Ekman numbers* from 10^{-3} to 10^{-4} and a range of *magnetic Prandtl numbers*, *Pm*, from 5 to 0.5 (the Prandtl number is here set to unity and *Pm* is therefore strictly equivalent to the *Roberts number*, q). Through this systematic parameter space survey, they were able to improve the work initiated by Christensen et al. [22], which had demonstrated that at fixed E, dynamo action could only be obtained if Pm exceeds a critical value, Pm_c, which decreases as E decreases. This suggests that there might be a continuous path in the space of models from the presently available models towards a model relevant for the Earth (i.e. at low E and low Pm). Kutzner and Christensen established that as the *Rayleigh number*, Ra, is increased from zero (all other parameters being fixed), if the flow can act as a dynamo (i.e. if Pm is large enough in the above sense), it first produces a field dominated by a strong axial dipole; but if Ra is further increased, reversing dynamos are produced. These are, however, found to be weakly dipolar (and thus not very "Earth-like").

This parameter range has been later extended by Christensen and Aubert [23], who confirm this general behavior with values of E as small as 3×10^{-6} and values of Pm as small as 0.05. They confirm an "approximate relation" (introduced in Christensen et al. [22]) of the form $Pm_c \sim E^{3/4}$. If it can be applied to the Earth's core, such a scaling law would then yield a very small estimate of Pm_c, as small as 10^{-11}. Such an estimated threshold being lower than the typical Pm relevant to the Earth's core dynamics (10^{-5} or 10^{-6}), provides another hint of a possible connection between numerical dynamos and a dynamo model, yet to be obtained, in the relevant parameter régime for the core of the Earth.

This scaling, $Pm_c \sim E^{3/4}$, has, as far a we are aware, not yet been interpreted. We propose here an interpretation in term of exponential amplification by a time dependent shear (Dormy and Gérard-Varet [24]). We know (Morin [25], Morin and Dormy, in preparation) that the minimum value of Pm for dynamo action does not correspond to an instability of the hydrodynamic solution to a magnetic perturbation, but instead to a detached dynamo branch. However, as the full phase diagram appears to scale in a similar manner with E, we can assume that the critical value of Pm above which the purely hydrodynamic flow becomes unstable decreases with E in the same manner (i.e. scales as $E^{3/4}$ according to Christensen and Aubert [23]). From previous theoretical studies of the onset of convection in a rotating sphere (Roberts [26], Busse [27]), it is expected that the pulsation of the thermal Rossby wave at onset scales as $\omega_c \sim E^{-2/3}$. These scalings are all expressed using the viscous timescale as unit of time. Following Dormy and Gérard-Varet [24], we note that dynamo action through a rotating shear requires $\tau_\beta < \tau_\eta$ (where $\tau_\eta = L^2/\eta$ and τ_β is the typical timescale of rotation or oscillation of the shear, here a rotation). We then get (with this unit of time) the necessary condition for this dynamo mechanism (based on timescale separation) to occur, $Pm > E^{2/3}$ (because Pm measures τ_η in the τ_ν timescale). The above scaling is in excellent agreement with the numerical data points provided by Christensen and Aubert, and this argument provides, to our knowledge, the first theoretical explanation for the decay of the critical magnetic Prandtl number for dynamo action when the Ekman number is decreased.

Let us also quote an interesting, yet puzzling, result of the Christensen and Aubert [23] study. Using a large dataset from numerical dynamos, for different parameter values, they derive a power law which relates the intensity of the produced field to the thermal forcing. In other words, the field intensity in these models is controlled by the available power, and does not result from a magnetospheric equilibrium. This raises interesting issues, especially as the scaling they propose is also compatible with the typical order of magnitudes for the geodynamo.

We finally want to mention briefly two other developments. First a set of impressive simulations performed by Takahashi et al. [28] on the *Earth Simulator*. They were able to decrease the Ekman number down to 4×10^{-6} and obtain a numerical dynamo with reversals. They have made an attempt to test whether their computations could achieve the *Taylor State* (i.e. the magnetostrophic branch). They were able to show that, when decreasing the Ekman number from 4×10^{-5} to 4×10^{-6}, their dynamo model evolved from a state in which the viscous torque equilibrated the magnetic torque to a state (which they called "quasi-Taylor-state") in which the viscous torque is significantly lower than the magnetic torque. The magnetic torque is thus large and dominant. The way to the real Taylor state, however, still appears rather long, as Taylor state is characterized by a vanishing magnetic torque! Recent models produced by Peter Olson (Olson [29]), appear to produce reversing dynamos, yet dominated by a strong dipole. This was obtained using an elaborate heating mode (combining thermal and compositional convection) and relying or rather large values of the magnetic Prandtl number (much larger than unity).

5.2. *Simplified models*

Despite significant progress, it is fair to say, as pointed out above, that we have not yet reached a full understanding of the mechanism of the Earth's and planetary dynamos.

There is, therefore, still room for simplified models, containing a small number of parameters and degrees of freedom, which can "capture" and illuminate salient features of the terrestrial dynamo, such as its time constants, without aiming at a realistic representation of its mechanism. The first simplified dynamo models are the disk dynamos of Bullard [30] and Rikitake [31] (see Hide [32], and Nozière's article in this issue [33]). We briefly mention a few followers:

1. *The relaxation model*. The Nozières dynamo [34] is presented in this issue by his author himself [33]. It has not received all the attention it deserves and must be ranked among the founding papers on simplified models. Further developments on a similar path, but fully dynamical, relying on the quasi-geostrophic approximation, are emerging (Lebrat and Dormy [35]).
2. *The multiple scale dynamo*. In a few papers (Le Mouël et al. [36]; Blanter et al. [37]; Narteau and Le Mouël [38]) another simplified dynamo model was developed, called a multiscale dynamo. As in Nozières's dynamo, the interaction of two magnetic modes and a convective flow is considered. But the flow is more complicated and intends to be more geophysical, although abstract and schematic, than Nozières's constant geometry flow. The magnetic field is the sum of a poloidal mode S and a toroidal mode T, T being generated from S by a differential rotation flow in the core, S from T through some kind of generalized α effect associated with a turbulent component of the fluid flow in the liquid core, characterized by a global helicity parameter $H(t)$ (a classical loop in dynamo theory). The abstract turbulent flow is made of cyclones of different scales, which have either a positive or a negative helicity; the resulting global helicity at time t is $H(t)$. The dynamics of the multiscale flow includes a transfer of helicity from smaller to larger scales (inverse cascade) and a transfer of helicity from larger to smaller scales (direct cascade). Both are necessary for the system to produce large scale symmetry breakings, i.e. positive or negative $H(t)$ with sudden sign changes, whereas as many positive helical cyclones as negative ones are continuously produced at small scale. Furthermore the magnetic field reacts on the flow, providing the necessary saturation mechanism. The model produces chrons of different lengths and reversals of different kinds (Section 4). S and T are never constant during a chron, but present a strong secular variation of the dipole. Slow and in average monotonous changes in the population of cyclones trigger at a time – unpredictable due to the stochastic ingredients of the model – big excursions of $H(t)$ towards negative values. A reversal or an excursion is then possible. Other dynamo models, partly based on similar ingredients, are due to Hoyng et al. [39,40]; Ryan and Sarson [41].

5.3. *Experimental models*

Since the year 2000, a new scientific approach is available to investigate fluid dynamos: liquid sodium experiments. The first two successful fluid dynamo experiments were driven with rather constrained flows, and yielded fluctuating, but non-reversing fields. In September 2006, the VKS experiment was successful in obtaining a self-induced dynamo field in an unconstrained volume of conducting liquid sodium (see Aumaître et al. [42] this issue).

This experiment was primarily guided by physical motivations. It was not directly designed as a model for the Earth's core. The geometry of the experiment is cylindrical, and not spherical, the forcing is mechanical rather than thermal and it is not built on a rotating table to include the effects of the *Coriolis force*, thought to be important in the Earth's core dynamics. With such major discrepancies, it may seem hopeless to attempt any comparison of the experimental data with the geomagnetic measurements. On the other hand, prompted by the apparent success of some numerical models, which also differ very significantly from the actual Earth (see above), cautious comparisons may be attempted. Comparing the VKS experiment with the Earth's core obviously prevents any detailed comparison (because of the differences in forcing and in geometry), but can help identify robust physical mechanisms. One may argue that an experimental dynamo must be affected by global rotation (as the core of the Earth) before it can be compared with the geomagnetic field. While the VKS-experiment is not placed on a rotating table, the mean flow can be, in some cases, dominated by a rapid global rotation, if the disks are not exactly counter rotating.

We will restrict our attention here to a set of experimental data published by the VKS-team (Berhanu et al. [43]; and also this issue [42]). It was observed in the VKS experiment that reversals only occur when the driving disks are not in exact counter rotation. When the rotation of the two counter-rotating disk is respectively fixed at 16 Hz and 22 Hz, series of seemingly chaotic reversals are reported. The genuine similarity of this time sequence with geomagnetic records was pointed by the authors, who also identified failed reversals, or excursions. Forgetting about boundary layers on the sidewall, and assuming that the difference in rotation between the propellers yields an average bulk angular velocity $\Omega \sim 20$ rad s^{-1}, one can then estimate the relevant parameters for the VKS experiment and compare

Table 2
Typical parameter regimes for the VKS-experiment, the Earth's liquid core and numerical models of the geodynamo

	VKS-experiment	Earth's core	Numerical models
Reynolds number Re (UL/ν)	$\sim 10^7$	$\sim 10^8$	~ 100
Magnetic Reynolds number Rm (UL/η)	~ 100	~ 100	~ 100
Magnetic Prandtl number Pm (ν/η)	$\sim 10^{-5}$	$\sim 10^{-6}$	~ 1
Ekman number E $(\nu/\Omega L^2)$	$\sim 10^{-6}$	$\sim 10^{-14}$	$\sim 10^{-5}$
Rossby number Ro $(U/\Omega L)$	~ 1	$\sim 10^{-6}$	$\sim 10^{-3}$

these with the geophysical estimates (see Table 2). It is then striking that the experimental conditions are in many respects closer to the core's dynamic than current numerical models.

Other quantities such as the magnetic Prandtl number (note that, the driving being mechanical, the Roberts number is here irrelevant and the magnetic Prandtl number provides the relevant diffusivities ratio) are indeed much closer to geophysical estimates in the experiment than in numerical simulations available so far (see Table 2). The occurrence of reversals in the "rotating regime" only provides an exciting observation for geophysicists. The dynamo threshold was not observed to decrease with this "global rotation" in the experiment. This observation is consistent with the above interpretation in terms of "timescale separation". Indeed, Rossby waves are absent in the cylinder with flat ends.

Let us finally stress that not only reversals, but excursions were also observed in the experimental data (see this issue). Polarity reversals are extremely reproductible in the experiment (meaning that the typical intensity variation during a reversal follows a well defined curve). It is characterized by an overshoot of the intensity immediately following the reversal, an ingredient which is present in paleomagnetic reconstructions (see Fig. 2). Let us stress that smaller experiments on magneto-convection and spherical Couette flow have also been performed in a geometrical configuration closer to the Earth's one and including the effect of global rotation (Nataf et al. [44]). Such a configuration (although providing us with interesting results) has however so far not led to dynamo action.

6. Conclusion

Despite impressive progress, numerical dynamos are not yet comprehensive representations of the Earth's dynamo. Nevertheless, their contribution is of first importance to help the understanding of the various ingredients of the geodynamo. As a result, there is still room for simplified models, containing a small number of degrees of freedom, which can also shed light on important features of the terrestrial dynamo. Finally, recent experiments open a new and exciting field of investigation. They will provide behaviors to be checked or predicted by numerical codes. The temporal series of chrons and reversals produced by the VKS dynamo can already be fruitfully compared with the corresponding series for the Earth's field and with actual observations of the polarity time scale.

All elements are present to offer combined approaches relying on observations, direct numerical simulations, simplified models, and now experiments, to yield progress in our understanding of the origin of the Earth's magnetic field.

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