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Modelling the palaeo-evolution of the geodynamo

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SUMMARY
Although it is known that the geodynamo has been operating for at least 3.2 Ga, it remains difficult to infer the intensity, dipolarity and stability (occurrence of reversals) of the Precambrian magnetic field of the Earth. In order to assist the interpretation of palaeomagnetic data, we produce models for the long-term evolution of the geodynamo by combining core thermodynamics with a systematic scaling analysis of numerical dynamo simulations. We update earlier dynamo scaling results by exploring a parameter space, which has been extended in order to account for core aspect ratios and buoyancy source distributions relevant to Earth in the Precambrian. Our analysis highlights the central role of the convective power, which is an output of core thermodynamics and the main input of our updated scalings. As the thermal evolution of the Earth’s core is not well known, two end-member models of heat flow evolution at the core–mantle boundary (CMB) are used, respectively, terminating at present heat flows of 11 TW (high-power scenario) and 3 TW (low power scenario). The resulting models predict that until the appearance of the inner core, a thermal dynamo driven only by secular cooling, and without any need for radioactive heating, can produce a dipole moment of strength comparable to that of the present field, thus precluding an interpretation of the oldest palaeomagnetic records as evidence of the inner core presence. The observed lack of strong long-term trends in palaeointensity data throughout the Earth’s history can be rationalized by the weakness of palaeointensity variations predicted by our models relatively to the data scatter. Specifically, the most significant internal magnetic field increase which we predict is associated to the sudden power increase resulting from inner core nucleation, but the dynamo becomes deeper-seated in the core, thus largely cancelling the increase at the core and Earth surface, and diminishing the prospect of observing this event in palaeointensity data. Our models additionally suggest that the geodynamo has lied close to the transition to polarity reversals throughout its history. In the Precambrian, we predict a dynamo with similar dipolarity and less frequent reversals than at present times, due to conditions of generally lower convective forcing. Quantifying the typical CMB heat flow variation needed for the geodynamo to cross the transition from a non-reversing state, we find that it is unlikely that such a variation may have caused superchrons in the last 0.5 Ga without shutting down dynamo action altogether.

Key words: Dynamo: theories and simulations; Palaeointensity; Palaeomagnetic secular variation; Reversals: process, timescale, magnetostratigraphy.

1 INTRODUCTION
Palaeomagnetic records can be used in order to shed light onto the past of the Earth’s dynamo. However, with increasing age, rock sequences are likely to be affected by weathering, alteration and metamorphism, thus destroying the pristine information on the early magnetic field. Although the situation constantly improves with newer samples and better techniques, especially regarding palaeointensity determinations (see Tauxe & Yamazaki 2007, for a review), uncertainties remain concerning the key characteristics of the geodynamo in the Precambrian. A first quantity of interest is the dipole moment. The oldest (3.2 Ga BP) reliable palaeointensity record to date (Tarduno et al. 2007) reveals that the virtual dipole moment (VDM) was possibly as large as its present value of $8 \times 10^{22}$ A m$^2$, although the consideration of the experimental cooling rate effect may lead to a twofold decrease in the determined VDM. As a
result, the debate concerning the long-term evolution of the VDM remains lively (Dunlop & Yu 2004), though there are suggestions (Macouin et al. 2004) of a long-term average monotonous increase from $3 \times 10^{22}$ A m$^{-2}$ at 1000–2000 Myr to $8 \times 10^{22}$ A m$^{-2}$ at present times. In addition, although it has been proposed (Stevenson et al. 1983; Hale 1987) that the power increase subsequent to inner core nucleation and onset of chemical convection in the core could cause a sudden increase in the dipole moment, it appears that until now, the palaeomagnetic data scatter has prevented a proper resolution of this feature. A second quantity of interest is the dipolarity of the field, which can be assessed (McFadden et al. 1991) through the analysis of latitude dependence in palaeosecular variation (PSV). Following this line, Smirnov & Tarduno (2004) proposed that the dipolarity should have been higher some 2.5 Ga ago. However, the existence of antipodal directions, which would be a clear evidence of high dipolarity (Dunlop & Yu 2004), is lacking in their data, and more generally, there are considerable issues with this method (Hulot & Gallet 1996), including the difficulty to separate the dipolarity effect from the secular variation effect in the PSV. The PSV can indeed also be used to study the intrinsic variability of the geomagnetic pole, as recently done by Biggin et al. (2008), who found evidence of a more stable dynamo 2.4–2.8 Ga ago. This led to the suggestion that a third quantity of interest, the reversal frequency of the geodynamo, was lower at that time than at present. The same conclusion was reached for the last 500 Myr (Eide & Torsvik 1996) through direct magnetostratigraphic analysis, which is obviously a more robust method when data is available, but some contrary indications of high reversal frequency have also been reported for the same period (Pavlov & Gallet 2001). It should be noted however that the strongest signal on the reversal frequency curve is related to the irregular occurrence of superchrons (see for instance, Pavlov & Gallet 2001) and that the strongest signal on the reversal frequency curve is related to the irregular occurrence of superchrons (see for instance, Pavlov & Gallet 2001). We study the Earth’s core at various stages of its existence, hence its rotation rate, aspect ratio $\chi = r_i/r_o$ and thermochemical buoyancy partitioning are variable over geological times, but can be assumed to be constant control parameters over timescales relevant for core dynamics. We define the deviation temperature field $T'$ and light element mass fraction field $\xi'$ with respect to the isentropic temperature and well-mixed mass fraction, and, within the Boussinesq approximation, both buoyancy effects are grouped into a codensity (or density anomaly) field (Braginsky & Roberts 1995) $C$ such that

$$C = a_\alpha T' + \Delta \rho \xi'. \quad (1)$$

Here $a_\alpha$ is the thermal expansion coefficient, $\rho$ is the fluid density and $\Delta \rho$ is the density difference between the light components that contribute to chemical convection and pure iron. The temperature and molar fraction fields are assumed to have the same diffusivity $\kappa$, due to turbulent mixing in the outer core. This allows us to write a single transport equation for the codensity $C$, which is solved numerically in a dimensionless form, together with the magnetic induction equation for the solenoidal magnetic field $\mathbf{B}$ in the magnetohydrodynamic approximation, and the Navier–Stokes and thermochemical transport equations for the incompressible velocity field $\mathbf{u}$, and pressure $P$

$$\frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} + 2e_z \times \mathbf{u} + \nabla P = \frac{\mathbf{r}}{r_o} \frac{r}{r_o} C - \frac{\mathbf{r}}{r_o} \times \nabla \times \mathbf{B} + E \nabla^2 \mathbf{u} \quad (2)$$

$$\frac{\partial \mathbf{B}}{\partial t} = \nabla \times (\mathbf{u} \times \mathbf{B}) + \frac{E}{P_m} \nabla^2 \mathbf{B} \quad (3)$$

$$\frac{\partial C}{\partial t} + \mathbf{u} \cdot \nabla C = \frac{E}{P_r} \nabla^2 C + S_{r/\xi} \quad (4)$$

$$\nabla \cdot \mathbf{u} = 0 \quad (5)$$

$$\nabla \cdot \mathbf{B} = 0. \quad (6)$$

Here $\mathbf{r}$ is the radius vector. Time is scaled with the inverse of the rotation rate $\Omega^{-1}$. Length is scaled with the shell gap $D = r_o - r_i$. Velocity is scaled with $\Omega D$. Magnetic induction is scaled by $(\rho \mu)^{1/2} \Omega D$, where $\rho$ is the fluid density and $\mu$ the magnetic permeability of the fluid. The Ekman number $E$, magnetic Prandtl and Prandtl numbers $P_m$ and $P_r$ are defined as

$$E = \frac{v}{\Omega D^2} \quad (7)$$

$$P_m = \frac{v}{\lambda} \quad (8)$$

$$P_r = \frac{v}{\kappa}. \quad (9)$$

In these respects, this paper takes advantage of advances made in both fields of numerical dynamo modelling and core thermodynamics to update a previous analytical treatment (Olson 1981). In Section 2, we introduce the numerical dynamo model, which we use for our systematic parameter space study. The numerical results are presented in Section 3.1. Then we turn to the investigation of various Earth cooling histories, and how they relate to the dynamo power (Section 3.2). Finally, we produce our time evolution model for palaeomagnetic observables (Section 3.3) and discuss the results in light of the palaeomagnetic observations and geophysical constraints (Section 4).
Here \( \nu, \lambda \) are, respectively, the viscous and magnetic diffusivities of the fluid. In this study, the codensity boundary conditions relevant to thermochemical convection are treated in the following way: at the inner-core boundary, the release of latent heat and light elements correspond to a positive mass anomaly flux \( F_i \) (expressed in kilograms per second), which we consider uniform and imposed on the long term by global core thermodynamics. The dimensional form of \( F_i \) writes

\[
F_i = - \int_{S_i} \kappa \nabla C \cdot \text{d}S.
\]

(10)

where \( S_i \) is the inner boundary surface. Similarly, at the outer boundary, we consider that the mantle imposes a uniform mass anomaly flux (which in fact corresponds to a heat flux without chemical contributions). The dimensional form of \( F_o \) writes

\[
F_o = - \int_{S_o} \kappa \nabla C \cdot \text{d}S.
\]

(11)

Here \( S_o \) is the outer boundary surface. Note that \( F_o \) is defined with respect to the advective heat flux carried out at the outer boundary (this is the reference state of the Boussinesq system). Hence, \( F_o \) can be either positive or negative, but in any case the total mass anomaly flux \( F = F_i + F_o \) must be positive for convection to occur. The codensity is scaled with \( F/4 \pi \Omega D^2 \). The Rayleigh number based on mass anomaly flux, \( RaQ_o \), which appears in (2) is therefore defined as

\[
RaQ_o = \frac{g_o F}{4 \pi \Omega D^2}.
\]

(12)

Here \( g_o \) is gravity at radius \( r = r_o \). The present formulation is slightly different from, but equivalent to that in Christensen & Aubert (2006), where a Rayleigh number based on advected buoyancy flux \( RaQ_o' \) was introduced. For sufficiently supercritical convection the conversion from their formalism to ours can be achieved through

\[
RaQ \approx \frac{RaQ_o}{D^3} RaQ_o'.
\]

(13)

The reason for introducing this change of formulation is that \( RaQ_o' \) is singular in the case where the inner core is not present.

In Section 3.1, we carry out a systematic analysis of the Boussinesq system and let the mass anomaly fluxes \( F_i \) and \( F_o \) be independent. The two associated control parameters are the Rayleigh number \( RaQ \) and the fraction of inner-boundary originated buoyancy \( f_i = F_i/F_o \). However, we emphasize that in the Earth's core, \( F_i \) and \( F_o \) are related through a consideration of the complete core thermodynamics, such as done in Sections 3.2 and 3.3. A general situation describing an Earth system which is slowly cooling on geological timescales is \( F_i \neq F_o \), in which case the basic state over which the Boussinesq system is considered has a decreasing temperature and increasing light element mass fraction, while the Boussinesq system itself is statistically stationary. This can be accounted for within our framework by adding a volumetric correction term \( S_{r/3} \) in (4), such that the mass anomaly budget of the spherical shell vanishes (Braginsky & Roberts 1995; Kutzner & Christensen 2002). The correction term then has the following dimensionless expression:

\[
S_{r/3} = \frac{1 - 2 f_i}{r_o^3 - r_i^3}.
\]

(14)

The other boundary conditions at both boundaries are of rigid type for velocity, and insulating for the magnetic field (the effect on inner-core conductivity on the long-term behaviour of dynamo simulations has been found to be negligible, as shown by Wicht 2002). The numerical implementation PARODY-JA is used in this study (see Aubert et al. 2008, for details). The numerical scheme is of finite-differencing type in the radial direction with up to 120 grid points, and uses a spherical harmonic decomposition in the lateral directions up to degree and order 106. No particular symmetry along longitude was assumed.

Table 1 gives details on the 43 models which we have integrated for this study. The parameter range is \( E = 3 \times 10^{-5} \) to \( 3 \times 10^{-4} \) for the Ekman number, \( RaQ = 10^{-6} \) to \( 10^{-3} \) for the Rayleigh number, \( Pm = 1 \) to 10 for the magnetic Prandtl number, and the Prandtl number \( Pr \) is set to 1 in all simulations. It is kept in mind that due to computational limitations, and just like all accessible numerical dynamo simulations (Christensen & Aubert 2006), our simulations operate in a parametric regime still very far from that of the Earth’s core, where \( E \approx 3 \times 10^{-13}, RaQ \approx 10^{-13} \) and \( Pm = 10^{-6} \). Scaling analyses attempt to overcome this intrinsic limitation by identifying robust trends supported by reasonable underlying physical considerations. Most of the scalings used here bear little residual influence of the various diffusivities, which gives credence to their applicability to core conditions, since diffusivities represent the main reason why models and Earth are distant in parameter space. The models presented here explore the new parameter space axes relative to \( \chi \) and \( f_i \). The aspect ratio goes down to \( \chi = 0.01 \), where the inner core is practically non-existent in the simulation. This situation is meant to simulate the core before or at the inner core nucleation time. The geophysically relevant buoyancy driving mode is therefore secular cooling \( (f_i = 0) \), which, in our formalism, corresponds to no buoyancy at the inner boundary, a positive volumetric source term, and a fixed heat flow at the outer boundary. Several models have been computed at \( \chi = 0.05 \), where \( f_i \) has been set to 0.5 and 1. These cases correspond to a system where the inner core has just nucleated, and the buoyancy driving is therefore a mix of secular cooling and chemical convection. Finally, models with \( \chi = 0.35 \) were also needed for comparison with present Earth. Core thermodynamics predict that the main buoyancy source of the present-day geodynamo (in terms of available power) is chemical convection (Lister & Buffett 1995). For that reason, cases with \( f_i = 0, \chi = 0.35 \) are left out of the present study. However, the uncertainties pertaining to the determination of core–mantle boundary (CMB) heat flow (Lay et al. 2008) and core adiabat leave some room for the determination of \( f_i \). Here we use the published runs from Christensen & Tilgner (2004), Christensen & Aubert (2006) and Olson & Christensen (2006), which have been performed with fixed temperature boundary conditions. In our formalism, this is roughly equivalent to \( f_i = 0.5 \) (see Fig. 1). In addition, we have included models with \( f_i = 1 \), and also with \( f_i = 2, 10 \), describing situation of dominant chemical convection where the CMB heat flow is, respectively, just adiabatic, or below the adiabat. In the latter case, a stably stratified layer exists at the top of the outer core.

We define several outputs, which are all averaged over times much longer than core flow timescales (but shorter than geological timescales for Earth’s mantle variations), and over the full volume of the spherical shell or the surface of the outer boundary. Most of these bear the same definitions as in Christensen & Aubert (2006): the root-mean-squared velocity inside the shell \( Ro \), rms magnetic field amplitude inside the shell \( Lo, \) rms harmonic degree in the velocity field \( T, \) ohmic dissipation fraction of the convective power \( f_{\text{ohm}}, \) ratio \( b_{\text{diff}} \) of the mean field strength inside the shell \( Lo \) to the dipole strength on the outer boundary \( B_{\text{dy}} \), and ratio \( f_{\text{dy}} \) of \( B_{\text{dy}} \) to the rms amplitude of the magnetic field at the outer boundary truncated at spherical harmonic degree 12. Parameters with definitions which are
proportional to the sum of the inner- and outer-bounding originated supercritical convection), the total dissipation $\Phi$ of the dynamo is proportional to the sum of the inner- and outer-boundary originated mass anomaly fluxes

$$\Phi = \Phi_i + \Phi_o = F_i(\psi_i - \overline{\psi}_i) + F_o(\overline{\psi} - \psi_o).$$  \tag{15}$$

Here $\psi$ is the gravitational potential such that the gravity vector is $\mathbf{g} = -\nabla \psi$, and $\psi_i, \overline{\psi}_i, \overline{\psi}$ are, respectively, the inner boundary, outer boundary, and mean values of the gravitational potential. The physical meaning of (15) is that the dissipation results from taking mass anomaly at a given gravitational potential, and redistributing it at the mean gravitational potential, which is where the good mixing assumption enters.

In the present context of radial gravity, the gravitational potential is expressed as $\psi = r r_g \sigma/2r_o + c s t$. The expression for $\overline{\psi}$ is (Buffett et al. 1996)

$$\overline{\psi} = \frac{3g_o}{10r_o} \left( \frac{r_o^3 - r_i^3}{r_o^3 - r_i^3} \right).$$ \tag{16}
We first present (Fig. 2) images from two models with a small inner core ($\chi = 0.05$, see Table 1 for other parameters). Model C is driven by inner-boundary originated buoyancy (chemical convection, $f_i = 1$), and model T is driven by volumetric heating and outer-boundary originated buoyancy (secular cooling, $f_i = 0$). Convection sets up where the thermochemical gradients are most unstable: one single convection cell near the inner boundary for C, five cells extending from mid-shell to the outer boundary for T. DMFI visualization (Aubert et al. 2008) reveals that the magnetic field is generated according to the classical macroscopic $\alpha^2$ mechanism (Olson et al. 1999), although model C additionally has an enhanced toroidal field production by zonal flow near the outer boundary. Both models are dipole-dominated, with magnetic dipoles of similar relative strengths (C and T, respectively, have $f_{\text{dip}} = 0.47, 0.49$). They are typically less dipolar than similar models with $\chi = 0.35$ (see Fig. 6).

In model T, the lower dipolarity can be explained by shallow convection columns which enhance magnetic flux expulsion and thus enrich the outer boundary power spectrum in multipolar content. In model C, the small inner core surface over which the buoyancy is distributed favours the occurrence of magnetic upwellings (Aubert et al. 2008) which reduce the dipolarity by frequently disrupting the magnetic dipole. In both models, the absence of the inner core favours a global axisymmetric poloidal circulation, which, at a given instant in time, concentrates magnetic field at one pole and disperses it at the outer pole (in the snapshots presented in Fig. 2, flux concentration is occurring at the south pole for model C and the north pole for model T). It should finally be noted that in thermal models, the absence of buoyancy at the inner-core boundary decouples the inner core from the convection and dynamo processes, with two consequences: first, thermal models with $\chi = 0.05$ or 0.01 yield almost the same results, which are presumably those which would be obtained from a case where $r_i = 0$. Second, the relationship

2.3 Statistics of least-squares fits

In Section 3.1, we obtain power laws of the form $y = ax^b$ from numerical data. The problem reduces to getting a linear least squares fit of the form $\ln y = \ln a + b \ln x$. In this fit, the slope $b$ is usually fairly well constrained by the large number of data points available. The standard best-squares error $\sigma_b$ on $b$ is therefore discussed only as we check the internal consistency of the scalings. For all other purposes, the vertical variance $\sigma^2$ of the errors $e_i = \ln \hat{y}_i - \ln \overline{y}_i$ is used, and as a standard practice, we present the best-fitting laws $y = ax^b$ together with their $3\sigma$ lines $y = (a/e^{\sigma})x^b$ and $y = ae^{\sigma}x^b$, which theoretically enclose 99.7 per cent of the data if the distribution of errors $e_i$ is normal.

3 RESULTS

3.1 Results from numerical dynamos

We present (Fig. 2) images from two models with a small inner core ($\chi = 0.05$, see Table 1 for other parameters). Model C is driven by inner-boundary originated buoyancy (chemical convection, $f_i = 1$), and model T is driven by volumetric heating and outer-boundary originated buoyancy (secular cooling, $f_i = 0$). Convection sets up where the thermochemical gradients are most unstable: one single convection cell near the inner boundary for C, five cells extending from mid-shell to the outer boundary for T. DMFI visualization (Aubert et al. 2008) reveals that the magnetic field is generated according to the classical macroscopic $\alpha^2$ mechanism (Olson et al. 1999), although model C additionally has an enhanced toroidal field production by zonal flow near the outer boundary. Both models are dipole-dominated, with magnetic dipoles of similar relative strengths (C and T, respectively, have $f_{\text{dip}} = 0.47, 0.49$). They are typically less dipolar than similar models with $\chi = 0.35$ (see Fig. 6).

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For the scaling relationships to be presented in the next section, the use of $p$ instead of $Ra_Q$ as a basic scaling parameter is motivated by several important reasons. First, the theory underlying these scalings (Christensen & Aubert 2006) relies on convective power arguments, while $Ra_Q$ is merely an approximate proxy for convective power. In our present study, where we vary the buoyancy partition $f_i$ and the aspect ratio $\chi$, the use of $Ra_Q$ as a basic scaling parameter does not capture the geometrical and buoyancy distribution effects contained in (18), thus resulting in a large scatter that the use of $p$ corrects for. Another advantage of using $p$ instead of $Ra_Q$ is that $p$ is a geophysical parameter which is constrained by thermodynamic studies of the Earth’s core (e.g. Lister 2003, see Section 3.2). Finally, expressing the scalings with $p$ leads to a useful internal consistency relationship to be detailed in Section 3.3.

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Figure 1. Test of the perfect mixing theoretical $p/Ra_Q$ relationship with numerical data. Symbols represent the numerical data, and lines represent the theoretical prediction of eqs (18) and (17). Red colour: Christensen & Aubert (2006) data points. Our theory applies to their choice of fixed temperature boundary conditions provided $f_i$ and $1 - f_i$ are replaced by 1 in eq. (18). White symbol filling and light grey line: cases of secular cooling with $f_i = 0$. Medium-grey symbol filling and line: cases of thermochemical convection with $f_i = 0.5$. Dark-grey symbol filling and black line: cases of purely chemical convection with $f_i = 1$. Blue symbols and lines: cases where the shell is stably stratified at the outer boundary (subadiabatic system) with $f_i = 2$. Green symbols: same as blue, for $f_i = 10$. For symbol shape definitions see Fig. 3.

Expanding (15) with the help of (16), and making use of the fact that the conservation of energy, when averaged over times long relatively to core flow timescales, yields $(p\Omega^2D^2)$ $pF = \Phi$ (recall that $p$ is a dimensionless quantity), we obtain after some algebra the following proportionality relationship between the power per unit volume $p$ and the total mass anomaly flux $F$

$$p = \gamma Ra_Q$$

with

$$\gamma = \frac{3(r_o-r_i)^2}{2(r_o^2-r_i^2)} \frac{f_i}{\gamma} \left[ \frac{3}{5} \frac{\chi}{\gamma} \frac{f_i}{\gamma} \chi - \frac{r_i^2}{5} \right] + (1-f_i) \left[ \frac{3(r_o-r_i)^2}{2(r_o^2-r_i^2)} \frac{f_i}{\gamma} \chi - \frac{r_i^2}{5} \right].$$

Eqs (18) and (17) are the generalizations to arbitrary buoyancy distributions of the relationships obtained in appendix A of Christensen & Aubert (2006). They are tested versus our numerical data in Fig. 1. The agreement between theoretical and numerical values of $p/Ra_Q$ becomes good as the supercriticality of convection increases. Cases with a strong stable density stratification (i.e. $f_i \gg 1$) need a stronger level of mixing to approach the theoretical line. This is simply a consequence of the fact that convection does not fill the entire shell in these cases. For a given level of convection supercriticality, the mass anomaly, while indeed produced at the inner boundary, fails to be redistributed at the mean gravitational potential of the shell to a greater extent.

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between reverse magnetic flux patch locations and inner core size (Stanley et al. 2007) does not hold if secular cooling drives the dynamo.

Fig. 3 presents a plot of the Rossby number $Ro$ (or dimensionless rms flow velocity) versus the dimensionless convective power $p$. The best fit and $3\sigma$ lines have equations

$$Ro = (0.69, 1.31, 2.49)p^{0.42}. \quad (19)$$

When cast into a $p - Ro$ space, the Christensen & Aubert (2006) scaling is virtually unchanged by the addition of the new data points.
Dynamos with a dipole fraction $f$ of dimensionless rms magnetic field $B_{\text{rms}}$, which obeys a scatter which is also unchanged. In the dipole-dominated regime with variable inner core size and buoyancy distribution, with a present aspect ratio ($\chi < 0.05$) additional models (triangles and open stars) have $Pm = 3$. Our data set is filtered to exclude the non-dynamo run.

Figure 5. Magnetic dissipation time $\tau_{\text{diss}}$, normalized by the standard magnetic diffusion time $\tau_{\text{mag}}$, as a function of the magnetic Reynolds number $Rm$. Symbols as in Fig. 3, except the red crosses which represent the Christensen & Tilgner (2004) models. Our data set is filtered to exclude the non-dynamo run.

For this last scaling, the newer runs cause a significantly larger scatter than that obtained by Christensen & Tilgner (2004). We attribute this to the rather small size (26 models) and limited parameter space extent of the data set used in that study, and conclude that the aspect ratio and buoyancy distribution have little influence on this scaling.

For an extrapolation of the three scalings presented above to Earth’s core conditions, the scaling prefactors need to be determined. Here we derive a useful internal consistency relationship tying these prefactors. We first define the dimensional values $B_{\text{rms}}$ and $U_{\text{rms}}$ for $Ro$ and $Rm$, expand them using (19) and (20) and also recall the definition of $\tau_{\text{diss}}$

$$B_{\text{rms}} = Lo(p\mu)^{1/2}D = c_1 f_{\text{shm}}^{0.5}p^{0.34}(p\mu)^{1/2}D, \quad (22)$$

$$\tau_{\text{diss}} = c_2 Rm^{-1}D^2/\lambda \quad \text{with} \quad Rm = U_{\text{rms}}D/\lambda, \quad (23)$$

$$U_{\text{rms}} = (D^2)p^{0.42}D. \quad (24)$$

Here the $c_i$ coefficients are scaling prefactors, to be chosen within the $3\sigma$ range of each scaling. In the case of the geodynamo, where the magnetic diffusivity is much larger than the viscous diffusivity, and the magnetic energy much larger than the kinetic energy, we expect most of the convective power $\rho\Omega^2 D^2 pV$ to be dissipated through ohmic losses (i.e. $f_{\text{shm}} \approx 1$), hence

$$\rho\Omega^2 D^2 p = \frac{B_{\text{rms}}^2}{2\mu} \frac{1}{\tau_{\text{diss}}}. \quad (25)$$

After expanding all quantities with the use of their scaling laws, the dimensionless version of (25) writes

$$1 = \frac{c_1^2 c_3}{2\sigma} p^{0.1}. \quad (26)$$

This reveals a good consistency of the scalings: for instance, the dependency in $\lambda$ vanishes as it should, thanks to the scaling exponent $-1$ in (23). However consistency is not perfect, as witnessed by the residual power 0.1 at which $p$ appears. As in the next section, $p$ will be assumed to vary over geological time, this will prevent (26) to be exactly satisfied with time-independent values for $c_{1-3}$. This problem can be related to the standard error in the least-squares determination of each exponent. We obtained standard exponent
errors of $\sigma_{b1} = 0.0169, \sigma_{b2} = 0.0639$ and $\sigma_{b3} = 0.0178$ for the $Lo$, $\tau_{dus}$ and $Ro$ scalings, respectively. The error on the exponent of the ohmic dissipation, in the right-hand side of (26), is therefore

\[ \sigma_b = 2\sigma_{b1} + 0.42\sigma_{b2} + \sigma_{b3} = 0.078. \]  

(27)

As $\sigma_b$ is comparable with the residual power 0.1 present in (26), we conclude that this inconsistency reflects the inherent error introduced by data scatter in the least-squares fitting procedure used to derive the three scalings. This error remains small however, and can be compensated (as will be done in Section 3.3) by choosing a set of $c_{i-1}$ that minimizes the deviations caused by $p^{b,1}$ (typically 20 per cent).

The control parameter $Ro_I = RoI/\pi$ was identified in Christensen & Aubert (2006) as the main parameter governing the dipolarity and stability versus reversals of numerical dynamo models. Here we follow the same approach (Fig. 6). In the dipole-dominated regime ($f_{dip} > 0.35$), Fig. 6 shows again that the absence of the inner core generally decreases the dipolarity, as already explained in Fig. 2. In contrast, the influence of buoyancy distribution on dipolarity is not clear-cut and not monotonic. The critical Rossby number for reversals $Ro_{Ic}$ is located in a narrow range $Ro_{Ic} = 0.04–0.12$. The frequency of polarity reversals increases with increasing departures of $Ro_I$ from $Ro_{Ic}$.

We then analyse (Fig. 7) the ratio $b_{dip}$, which determines the relative strength of the internal field and the dipole field at the outer boundary. Since, as shown by Fig. 6, the geodynamo is likely to have had a dipolarity lower or equal to the present-day value throughout its history, we restrict the data to the range $0.4 < f_{dip} < 0.7$. In this range, we find little variability of $b_{dip}$ with the convective power (or $Ro_I$), whereas the main source of variability comes from $f_I$ and $\chi$. This can be attributed to depth variations of the dynamo region, the dipole seen at the surface being reduced in the case of a deeper dynamo. Indeed chemical models (deeper dynamos) have a markedly larger $b_{dip}$ value than thermal models (shallow dynamos, see Fig. 2). The highest $b_{dip}$ values are obtained for models with $f_I > 1$, where the extent of the dynamo region is restricted even further by a stably stratified layer at the top of the shell (see for instance, Christensen 2006). In Section 3.3, a model will be needed for $b_{dip}$ in order to estimate the dipole moment at any point in the geological history. We adopt the simplest possible linear dependency

\[ b_{dip} = 7.3(1 - \chi)(1 + f_I). \]  

(28)

The last scaling which we check (Fig. 8) is the relationship (Olson & Christensen 2006) between the local Rossby number $Ro_I$ and the dynamo control parameters $p$, $Ek$, $Pm$ and $Pr$. As there does not currently exist a physical rationale to exclude some control parameters, the powers in this last scaling were obtained by an approach of empirical scatter minimization. We confirm (Fig. 8a) the scaling relationship obtained by Olson & Christensen (2006)

\[ Ro_I = (0.34, 0.68, 1.35)p^{0.48}E^{-0.32}Pr^{0.19}Pm^{-0.19}. \]  

(29)

Unlike in the previous scalings, there appears an additional dependence on the aspect ratio $\chi$, which we resolve by adopting the following relationship, yielding a smaller data scatter (Fig. 8b):

\[ \frac{Ro_I}{(1 + \chi)} = (0.33, 0.54, 0.89)p^{0.48}E^{-0.32}Pr^{0.19}Pm^{-0.19}. \]  

(30)

Note that factoring out a $(1 + \chi)$ dependence also slightly reduces the scatter in the critical $Ro_{Ic}$ values obtained from Fig. 6. We therefore subsequently replace the criterion for reversals $Ro_{Ic} = 0.04–0.12$ with

\[ \frac{Ro_{Ic}}{(1 + \chi)} = (0.33, 0.54, 0.89)p^{0.48}E^{-0.32}Pr^{0.19}Pm^{-0.19} \approx 0.04–0.1. \]  

(31)

The aspect ratio thus disappears from our condition for the onset of reversals.

### 3.2 Core thermodynamics and cooling models

The previous section shows how time-average properties of the past geodynamo can be estimated from the convective power. The assumed time scale for this time average is long (say a million years) compared to core flow timescales, but short compared to any geological evolution timescale of the Earth. The geological evolution...
of dynamo properties can be obtained from our scaling laws if we have a long-term evolution model for the convective power, which we now derive from core thermodynamics (see the recent studies from Labrosse 2003; Lister 2003, and references therein). The fundamental inputs which are needed is an history of core-mantle heat flow, which is assumed constant over time. In what follows, we use the simple parametrization of Lister (2003) which holds if the inner core volume is small relative to the outer core volume ($\chi^3 \ll 1$).

The entropy budget of the system, when time-averaged over timescales much longer than core flow timescales, but shorter than geological timescales, gives an expression for the total dynamo dissipation $\Phi$ as a function of the fundamental inputs listed above. The part $\Phi_i$ of the dissipation that originates from mass anomaly flux at the outer boundary is then

$$\Phi_i = (Q_{\text{cub}} - Q_o)(\epsilon_L + \epsilon_B).$$

where $\epsilon_L$ and $\epsilon_B$ are, respectively, the thermodynamic efficiencies of latent heat and light element release, which are given by

$$\epsilon_L = \frac{3\mathcal{L}[1 - \mathcal{F}(\phi)]\chi}{2 + 3(\mathcal{L} + \mathcal{B} - \mathcal{C})\chi}$$

$$\epsilon_B = \frac{3\mathcal{B}\chi}{2 + 3(\mathcal{L} + \mathcal{B} - \mathcal{C})\chi}.$$  

The values of the thermodynamic parameters $\mathcal{L}$, $\mathcal{B}$, $\mathcal{C}$, $\mathcal{F}(\phi)$ and $\phi$ are given in Table 2. The part $\Phi_o$ of the dissipation that originates from mass anomaly flux at the outer boundary is

$$\Phi_o = (Q_{\text{cub}} - Q_o)\epsilon_S,$$

where $\epsilon_S$ is the thermodynamic efficiency of thermal convection, given by

$$\epsilon_S = e^{\delta}\Phi(\phi) - 1.$$  

The dimensionless, total volumetric power $p$ can be obtained through the time average conservation of energy $\rho \Omega^2 D^2 p V = \Phi_i + \Phi_o$, where $V$ is the shell volume. Once the dissipations $\Phi_i$ and $\Phi_o$ are known, the associated mass anomaly fluxes $F_i$ and $F_o$ can be retrieved from eq. (15), to finally determine $f_i = F_i/(F_i + F_o)$. Note that for strongly subadiabatic cases, eq. (15) may

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Meaning</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_{\text{cub}}(t)$</td>
<td>Earth’s cooling model</td>
<td>Variable</td>
<td>See Section 3.2</td>
</tr>
<tr>
<td>$Q_o$</td>
<td>Adiabatic heat flow at the CMB</td>
<td>6 TW</td>
<td>See Section 3.2</td>
</tr>
<tr>
<td>$Q_r$</td>
<td>Radiogenenic heating in the outer core</td>
<td>0 W</td>
<td>See Section 3.2</td>
</tr>
<tr>
<td>$\mathcal{M}$</td>
<td>Heat capacity for solidification</td>
<td>$9.2 \times 10^{29}$ J</td>
<td>Lister (2003)</td>
</tr>
<tr>
<td>$\mathcal{L}$</td>
<td>Latent heat effect</td>
<td>2.1</td>
<td>Lister (2003)</td>
</tr>
<tr>
<td>$\mathcal{B}$</td>
<td>Buoyancy effect</td>
<td>0.86</td>
<td>Lister (2003)</td>
</tr>
<tr>
<td>$\mathcal{C}$</td>
<td>Compositional effect</td>
<td>$-0.8$</td>
<td>Lister (2003)</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Adiabatic decay parameter</td>
<td>0.256</td>
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</tr>
<tr>
<td>$\mathcal{F}(\phi)$</td>
<td>Adiabatic decay integral</td>
<td>0.8595</td>
<td>Lister (2003)</td>
</tr>
<tr>
<td>$r_o$</td>
<td>Outer core radius</td>
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<td></td>
</tr>
<tr>
<td>$\Omega$</td>
<td>Earth’s rotation rate</td>
<td>variable</td>
<td>See Section 3.3</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Outer core magnetic diffusivity</td>
<td>$1.3 \text{ m}^2 \text{ s}^{-1}$</td>
<td>Secco &amp; Shloessin (1989)</td>
</tr>
<tr>
<td>$\nu$</td>
<td>Outer core viscosity</td>
<td>$10^{-6} \text{ m}^2 \text{ s}^{-1}$</td>
<td>de Wijs et al. (1998)</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>Outer core thermochemical diffusivity (assumed to be the same as thermal diffusivity)</td>
<td>$5 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$</td>
<td>Labrosse et al. (2007), Lay et al. (2008)</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Outer core density</td>
<td>$10^4 \text{ kg m}^{-3}$</td>
<td>Dziwonski &amp; Anderson (1981)</td>
</tr>
</tbody>
</table>
not hold because of incomplete mixing, which in this case might lead to an underestimation of $F_i$. It should be mentioned here that breaking the total dissipation $\Phi$ into boundary-originated terms, and equating these two terms separately to their equivalent convective fluxes is not a trivial operation, since dissipation is a global non-linear quantity. The derivation presented in appendix shows that this is legitimate if the accuracy of the Boussinesq approximation is tolerable.

Once the inner core is present, the evolution of the Earth core aspect ratio $\chi$ is constrained by the heat capacity for solidification $\mathcal{M}$ (value in Table 2) through the equation

$$\mathcal{M} \frac{d}{dt} \left[ x^2 + (L + B - C)x \right] = -(Q_{\text{cmb}} - Q_r). \quad (37)$$

Eq. (37) can be integrated backwards in time from present (Labrosse et al. 2001) until the inner core age $a$ (Here time is measured before present, therefore $a > 0$). The amount of radioactive heating $Q_r$ in planetary cores is debated. Experiments of potassium partitioning between iron and silicates suggest that present maximum potassium concentration amounts to values ranging from 30 p.p.m (Hirao et al. 2006) to 60–130 p.p.m (Rama Murthy et al. 2006), from geochemical constraints and from the suggested present crystallization of a basal magma ocean in the lower mantle. As indicated by Fig. 9(a), it yields a typical dynamo power of 2.7 TW at present. The second, low power scenario is motivated by the fact that the scaling of ohmic dissipation in numerical dynamos (Christensen & Tilgner 2004) favours a low present dissipation of about 0.2–0.5 TW. Using the thermodynamic analysis presented above, this implies that the top of the Earth’s core is presently subadiabatic ($Q_{\text{cmb}} < Q_e$, see Fig. 9b). A variety of idealized, constant rate cooling histories can be built, which cross the adiabat at an age $b$. Plausible models are such that $b < a$ (Labrosse et al. 1997), because if $b > a$ then convection stops in the Earth’s core between the adiabat crossing and the nucleation of the inner core. This is not acceptable since a conducting core would not cool fast enough to subsequently nucleate an inner core before present. Our second scenario is built according to this constraint, taking an initial CMB heat flow of $Q_{\text{cmb}} = 11$ TW, and a present value $Q_{\text{cmb}} = 3$ TW, corresponding to a present dynamo power of about 0.3 TW.

3.3 Time evolution models for palaeomagnetic observables

We now combine the dynamo scaling study from Section 3.1 with the thermodynamic elements from Section 3.2 in order to evaluate how the main properties of the geodynamo evolve over time. At any point in time, the power $P$ can be accessed from the analysis of the previous section. The Prandtl and magnetic Prandtl numbers are set according to the diffusivity values listed in Table 2. In order to determine the Ekman number $E$, the rotation rate of the Earth is needed. We use the length-of-day model (LOD) of Varga

Figure 9. High- (a) and Low- (b) power models for: heat flow from the core to the mantle $Q_{\text{cmb}}$, adiabatic heat flow $Q_r$, inner- and outer- boundary originated dissipations $\Phi_i, \Phi_o$, and total power $P = \Phi_i + \Phi_o$ as functions of the time before present. The greyed area represents the uncertainty range for the adiabatic heat flow. The inner core nucleates at $a \approx 1.8$ Ga before present in the low power scenario, and $a \approx 0.8$ Ga before present in the high-power scenario.
Figure 10. (a) Dynamo power and ohmic dissipation (respectively, left- and right-hand side of the internal consistency relationship (25), both multiplied by the shell volume $V$), using the prefactor set $(c_1, c_2, c_3) = (1.65, 0.11, 1.31)$ for both scenarios. (b) rms core velocity $U_{\text{rms}}$, with indications of the equivalent magnetic Reynolds numbers $Rm$ at present, and a rough delineation of the dynamo onset which would correspond to $Rm \approx 40$ (Christensen & Aubert 2006). (c) rms core magnetic field $B_{\text{rms}}$. (d) Local magnetic Reynolds number $R_{\text{ol}}$, (corrected by $1 + \chi$). For this last scaling the central value 0.54 from (30) is used, and the 3σ uncertainty range is propagated to the location of the critical value $R_{\text{olc}}$ for reversals obtained from (31) (shaded zone, lighter shade of grey means higher likeliness for reversals/higher reversal frequency). In (b)-(d), the dashes represent the epoch with no available palaeomagnetic samples.

The determination of absolute values for dynamo properties is subject to a considerable amount of uncertainty, which is discussed in detail in the next section. For that reason, we focus on the trends, or absence thereof, rather than the absolute values. Here we therefore present time evolution models obtained with scaling prefactors from (22) and (24) as close to their central values as possible, while still satisfying the constraint (26) of internal consistency. Using the central values for the prefactors $c_i$ as a starting point, we obtain $p^{0.1} c_1^2 c_3 / 2 c_2 \approx 0.2$ throughout time. We therefore need to adjust the prefactors $c_i$ within the 3σ error range, increasing $c_1$ and decreasing $c_2$. In order to keep $c_{1,3}$ close to their central values, we first decrease $c_2$ from 0.26 to its minimal acceptable value 0.11. There subsequently remains some discrepancy in (26), which we can cancel out by increasing $c_1$ from 1.17 to 1.65 while keeping $c_3 = 1.31$. Our predictions for $U_{\text{rms}}$ (or the magnetic Reynolds number $Rm$), $B_{\text{rms}}$ and the local Rossby number $R_{\text{ol}}$ are reported in Figs 10(b), (c) and (d). We note that the model for $B_{\text{rms}}$ implicitly assumes that $f_{\text{atom}} \approx 1$ throughout time, and that the dynamo has been dipole-dominated throughout Earth’s history. This last point is reasonable since our models show that $R_{\text{ol}}$, the parameter controlling the breakdown of dipolarity (Fig. 6) has been below its present-day value throughout Earth’s history (Fig. 10d).

An illustrative indication of how the internal properties evolution previously computed may translate to surface observables can be obtained by computing the true dipole moment

$$M = \frac{4\pi r_0^3}{\sqrt{2\mu}} B_{\text{rms}}.$$  

For the determination of $b_{\text{dip}}$, we use the simple model (28), the time evolution of which is presented in Fig. 11(a). In order to put the results in perspective with the considerable scatter in palaeointensity data, the resulting true dipole moment time-series (Fig. 11b) are presented together with virtual dipole moment values (VDM) from the IAGA palaeointensity database (Perrin & Schnepp 2004; Biggin et al. 2009).

4 DISCUSSION

Fig. 11(b) shows that throughout the Earth’s history, the dipole moments predicted by our models agrees with the observed palaeointensities to better than an order of magnitude, a fact which can be
seen as a success for the underlying theories of convective dynamos and core thermodynamics. However, it must be kept in mind that a more specific analysis, such as discriminating between the high- and low-power scenarios based on comparison with palaeointensity data, is necessarily plagued by a considerable amount of uncertainty. First, we have used central prefactor values for our scaling predictions, bearing a typical uncertainty of a factor 2 above and below the central line. Similarly, the conversion from deep to surface properties is also suffering from a similar factor 2 uncertainty (Fig. 7) and from departures from the perfect mixing theory (though in our models, $f_i$ does not greatly exceed 1, thus limiting the impact of incomplete mixing). The determination of thermodynamic properties in the upper part of Table 2 is also uncertain to a typical factor 2 above and below the central line. Similarly, the palaeointensity data against which we compare our models has a large scatter which could result from either time variability of the field or intrinsic determination difficulties to generate a magnetic field with strength comparable to that of the present field: both cooling scenarios predict a dipole moment of $5 \times 10^{22}$ Am$^2$ 3 Ga ago, on par with present-day values, and compatible with the findings of Tarduno et al. (2007). Based on this result, the suggestion (Dunlop 2007) that old palaeomagnetic samples such as those analysed by Tarduno et al. (2007) provide evidence that the inner core was already present before 3.2 Ga ago should be discarded. While the inner-core nucleation itself certainly has a significant impact on the internal magnetic field amplitude, the signature is largely attenuated in the dipole moment time-series (Fig. 11a), due to the evolution of the conversion factor $b_{dp}$. Indeed, as the inner core nucleates, the onset of chemical convection brings additional thermodynamic efficiency in the system, extracting more power from a given CMB heat flow. At the same time, however, the location of the dynamo shifts from mid-shell to the shell centre, thus masking the field amplitude increase seen from the surface. The prospect of observing the signature of inner core nucleation through a sudden increase of the palaeointensity at the Earth’s surface, following the suggestion of Hale (1987), is therefore reduced.

We now discuss the predicted evolution for the dipolarity and stability of the geodynamo. Our results suggest that for a given convective forcing (more precisely, for a given local Rossby number), the absence of an inner core reduces the dipolarity. All forcing conditions being equal, the earlier geodynamo could therefore have been less dipolar than at present. However, inferring the dipolarity of the past geodynamo implies to weigh the effects of the geometry at given forcing, and of the forcing variations, a smaller $R_{ci}$ implying a possibly more stable and more dipolar dynamo (see Fig. 6). As, in both models, $R_{ci}$ has been generally lower than its present value throughout Earth’s history, and also less than an order of magnitude away from the critical value $R_{ci}$ for reversals (Fig. 10d), it is possible that 3–2.5 Ga ago, the effects related to the absence of the inner core and to a lower $R_{ci}$ might have compensated to yield
A striking feature of the recent palaeomagnetic record is the large dispersion of virtual dipole moments (see Fig. 11b), with, for instance, fluctuations from about 2 \times 10^{22} to 15 \times 10^{22} A m^2 in the last 500 Myr, occurring over very short (million year) timescales. These are unlikely to be explained by magnetohydrodynamic processes taking place in dynamo models, where the magnetic field has typically smaller (and faster) fluctuations (see for instance Olson 2008), with epochs of low dipole moment representing only rare events associated with reversals. Moreover, these are also too short to represent a response of the dynamo to changing mantle conditions, and, even if this was the case, the CMB heat flow would have to come very close to the minimal 2 TW required to get a dynamo (Fig. 12) in order to cause such large magnetic field fluctuations. Refinements in the knowledge of CMB heat flow variations, in dynamo theory, and in variability analyses of palaeomagnetic samples, will be needed in order to conclude regarding the physical nature of these variations.

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 Modelling the palaeo-evolution of the geodynamo

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APPENDIX

Here we follow the analysis presented by Lister (2003) to derive an approximate equivalence between dissipation at each boundary and its corresponding mass anomaly flux. Eq. (24) in Lister (2003) gives the original definitions of boundary-originated dissipations $\Phi_i$ and $\Phi_o$ as they stand in (32) and (35):

$$\Phi_i = \int f_c \cdot g \, dV + Q^*_{ICB}[1 - \frac{\bar{T}}{T_o(r_i)}]$$  \hspace{1cm} (A1)

$$\Phi_o = Q^*_{CMB}[1 - \frac{\bar{T}}{T_o(r_i)}].$$  \hspace{1cm} (A2)

Here $f_c$ is the convective compositional mass anomaly flux, $T_o$ is the adiabatic temperature gradient, $\bar{T}$ is the mass-averaged temperature in the shell, $Q^*_{ICB}$ and $Q^*_{CMB}$ are the superadiabatic heat flows at inner-core boundary and CMB. The last term in eq. (24) of Lister (2003) is neglected because the average temperature of dissipation is assumed to be $\bar{T}$. Denoting the compositional mass anomaly flux at the inner core boundary as $F_{ci}$, and assuming, as in Lister (2003) that compositional diffusion can be neglected and that the convective (not total) compositional flux is well mixed throughout the shell, we have $\nabla \cdot f_c = F_{ci}/V$ and an integration by parts of the first term in (A1) yields

$$\int f_c \cdot g \, dV \approx F_{ci}(\psi_i - \overline{\psi}).$$  \hspace{1cm} (A3)

Integrating the definition of the adiabatic temperature gradient and neglecting the radial variations of $\alpha$ and the heat capacity $C_p$, we have

$$\frac{\alpha}{C_p} (\psi(r) - \psi_i) = \ln \left[ \frac{T_o(r)}{T_o(r_i)} \right].$$  \hspace{1cm} (A4)

Under the assumption of moderate departure of $T_o$ from the mass-averaged temperature $\bar{T}$ and of moderate density variations, the mass average of the previous equation yields

$$\frac{\alpha}{C_p} (\psi_i - \overline{\psi}) \approx [1 - \frac{T}{T_o(r_i)}]$$  \hspace{1cm} (A5)

and we have therefore

$$Q^*_{ICB}[1 - \frac{\bar{T}}{T_o(r_i)}] \approx F_{ci}(\psi_i - \overline{\psi}),$$  \hspace{1cm} (A6)

where $F_{ci}$ is the thermal mass anomaly flux at the inner boundary. From here it is apparent that $\Phi_i$ as it stands in (32) corresponds to $\Phi_i$ as used in (15), the same result holding for $\Phi_o$. As discussed in Lister (2003), the accuracy of the various approximations used in this derivation reduce to that of the Boussinesq approximation.