Large hot provinces at the base of the mantle stabilise the palaeomagnetic field

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A crucial characteristic of Earth's magnetic field, known since 1600[1], is that it approximates a dipole (bar magnet) aligned with the planetary rotation axis. Previous studies have disagreed over the extent to which this situation has persisted through geological time[2-5] which is important for determining solar wind shielding and building palaeogeographic reconstructions. It has recently become possible to quantify this axial dipole dominance back in time using the equatorial variability of the palaeomagnetic field recorded in rocks[6]. Here we show that this aspect of palaeomagnetic field behaviour was extraordinarily stable across three periods spanning the last 3 billion years despite being an unstable characteristic in most numerical geodynamo simulations. We further demonstrate a means to reproduce this stability in new rapidly rotating and turbulent geodynamo simulations. Large, seismically-inferred, hot provinces in the lowermost mantle[7] suppress core flow beneath them[8], impose large-scale order on the magnetic field emerging from the core at low latitudes, and stabilise the palaeomagnetic behaviour of the simulations enabling them to reproduce the observations. Our combined modelling and observational results suggest that features akin to these large low velocity provinces in the lowermost mantle today may have stabilised Earth's magnetic field throughout much of geological time.

Main text

The extent to which the Earth's magnetic field resembles a bar magnet aligned with the rotation axis determines its navigational utility (making the local field point north or south) and impacts on the degree of planetary shielding provided from solar wind radiation[9]. Furthermore, the assumption that palaeomagnetic directions, when averaged over sufficient time, yield only the geocentric axial dipole (GAD) is crucial for palaeogeographic reconstructions that are relied on by investigations of geodynamics and palaeoclimate in the deep past[4]. Detailed descriptions of geomagnetic morphology have only been produced for the last 100 kyr[10]. Previous attempts[4,5,11-13] to confirm (or refute) the GAD-like nature of the field in deep time have been somewhat imprecise or prone to controversy[14,15]. The centrality of the concept of an axially dipole dominated magnetic field to the Earth sciences clearly motivates further investigations into both the morphology of the palaeomagnetic field itself, and of the potential mechanisms at work in the deep Earth to promote or confound axial dipole dominance.

It has long been considered likely that measures of the intrinsic variability of the palaeomagnetic field, referred to as palaeosecular variation (PSV), have the potential to provide quantitative morphological constraints[16] but this has only recently become possible in practise[6]. PSV is often defined by the angular dispersion of virtual geomagnetic poles (VGPs) retrieved from fast-cooled igneous units. VGP dispersion (S) has been shown to generally increase with the (palaeo)latitude of the sampling site[17] and is reasonably fit by the simple quadratic *Model G* [16]:

$$S^2 = a^2 + (b\lambda)^2 \qquad (1)$$

where a and b are constants (for a given epoch) that define the value of S at the equator and the rate of its increase with palaeolatitude (λ) respectively. It is the Model G a parameter (hereafter referred to as simply a) that has been shown to be a powerful predictor of axial dipole dominance via a power law[6].

For this study, we obtained a compilation of 6,178 VGPs for the time interval 0 to 0.32 Ga based on measurements made on igneous rocks that had already been deemed suitable for PSV analysis[18-22] and performed a new combined analysis (see *methods*). The results are best fit by Model G using parameters of $a = 9.9^{+1.0}_{-1.5}$ and $b = 0.26^{+0.02}_{-0.08}$ (figure 1a). This new estimate of a averaged over 320 million years overlaps entirely with the same parameters previously obtained from two much earlier time periods (0.5-1.5 Ga and 1.5-2.9 Ga; figure 1b and 1c)[23]. Allowing for the 95% uncertainties in the mean values, the full range in a across the three intervals since 2.9 Ga is 7.4° – 11.4°. The values of median instantaneous axial dipole dominance (AD/NAD_{median})[6] that this range predicts (5.7 to 56.0; figure 1d) agree well with that obtained directly from GGF100k[10], a continuous time-dependent field model built using datasets from the last 100,000 years and yielding AD/NAD_{median} = 20. This range of a further predicts inclination anomalies (deviations in palaeomagnetic inclination caused by departures from GAD) that peak in the range of ~2 to ~11° (see *methods*), similar to that seen in the recent field (Figure S1)[22] and which are unlikely to severely impact palaeogeographic reconstructions made assuming GAD.

Our results provide evidence of remarkable stability in the palaeosecular variation of the magnetic field over the last 3 billion years and indicate that the assumption of uniformitarian GAD-like field behaviour over most of geological time is not unreasonable. That long-term average PSV has displayed such invariance is a somewhat surprising finding given that the secular evolution of Earth's core is expected to have radically affected the dynamo process responsible for generating this field during its long history[24-26]. To deduce possible mechanisms for producing this long-term stability, we investigated the capacity of numerical models to replicate our PSV observations using the outputs of a set of 88 geodynamo simulations (51 of which have been previously published[27,28 29]) each run with unique input parameters (Table S1; *Methods*). Some of these simulations were run with outer boundary conditions that specified a heterogeneous pattern of core-mantle heat flux (q) similar (see *Methods*) to that inferred from seismic tomography of the lowermost mantle[30]. This pattern is dominated by two equatorial and antipodal large low velocity provinces (LLVPs)[31] beneath Africa and the Pacific. Following common practice, we interpret variations in seismic velocity in terms of temperature such that the LLVPs are likely hotter than the average basal mantle and associated with the lowest values of q. The amplitude of the heat flow heterogeneity imposed on the outer boundary of the model, q^*, is defined by:

$$q^* = \frac{q_{max} - q_{min}}{q_{ave}}, \qquad (2)$$

where q_{max}, q_{min}, and q_{ave} are the maximum, minimum and average heat flux across the outer boundary respectively. Most simulations (56 or 64%) were run with q^*=0 signifying homogeneous boundary conditions whereas for the rest, $1 < q^* < 10$.

Most of the simulations (72 or 82%) produce Model G a parameter values outside of the bounds set by the palaeomagnetic records (figure 1d). The organisation of these simulations into "tracks"[27], whereby only the Rayleigh number (*Ra*; describing the ratio of buoyancy to the stabilising effects of rotation and thermal diffusion in the outer core) is varied, provides insight into why this is the case. There is a systematic tendency for values of *a* produced by the simulations to increase with increasing *Ra* such that, at best, they only reproduce the observed magnetic field behaviour at the surface in a narrow band of *Ra* values (figure 2). Because of the strong relationship between *a* and AD/NAD_{median}, most of the simulations are also far less dipolar than the palaeomagnetic field (figure 1d). *Ra* for Earth's dynamo has probably varied by more than a factor of 3 over geological time (see *methods*) and therefore those tracks passing through the palaeomagnetic bounds shown on figure 2 (or never entering it) do not represent Earth.

Figure 2c distinguishes simulations (denoted *LowE* from hereon) that were designed to mimic certain additional perceived characteristics of Earth's core dynamo: 1) a ratio of magnetic advection to magnetic diffusion timescales (the magnetic Reynolds number, *Rm*) of ~1000; 2) rapid rotation; 3) a balance of forces that is Quasi-Geostrophic (QG) at leading order and Magnetic-Archimedean-Coriolis (MAC) at first order[32,33]. With these properties, most of the simulated fields show excellent agreement with the palaeomagnetic and geomagnetic observations (figure 1d, figure 2c). Nevertheless, in the case where homogeneous boundary conditions were used (q^* = 0), the track follows the same tendency as non-*LowE* simulations (figure 2a,b) whereby the field becomes far too variable and multipolar as *Ra* is increased. The requisite stability only emerges when heat flow heterogeneity (q^* > 0) is also imposed at the outer boundary.

Other attributes of the palaeomagnetic field besides *a* and AD/NAD_{median} may be used to assess the capacity of the *LowE* dynamo simulations to reproduce palaeomagnetic observations. Here we developed a novel set of 5 criteria ($Q_{PM\text{-}100k}$) adapted from the established Q_{PM} criteria[27,28] but with thresholds derived (see *methods*) from the palaeomagnetic model GGF100k[10] rather than directly from palaeomagnetic records of the last 10 Myr. This was considered necessary because the short run times (50 – 140 kyr) of the *LowE* simulations, similar to GGF100k itself, does not allow them to capture the full range of variability exhibited by a 10 Myr span (incorporating many excursions, reversals, etc). The design of both Q_{PM} and $Q_{PM\text{-}100k}$ criteria dictates that compliance with observations produces misfits to individual criteria of 1 or less and total (summed) misfit values of 5 or less. By these metrics, those seven *LowE* simulations with heterogeneous boundary conditions outperform the two with homogeneous boundary conditions (Figure S2) and the majority of the non-*LowE* simulations (Table S1). Furthermore, the three *LowE* simulations run with the highest *Ra* and/or q* (LEDD003, 005 and 006) meet 4 or 5 of the criteria simultaneously and have extremely low total misfit values < 3.

Present-day LLVPs straddle the equator; therefore, if their influence on core dynamics is sufficiently important, one might expect a non-zonal (i.e. longitudinal) signature in the palaeomagnetic field observed at low latitudes, a region of particular importance in constraining Model G *a*. We compare the patterns of longitudinally varying VGP dispersion at low latitudes in GGF100k and those produced by our *LowE* simulations (figure 3). We find that the three simulations with lowest QPM_{100k} misfits (LEDD003, 005, 006) also produce nonzonal variability that is in good qualitative agreement with observations.

We have established that the *LowE* simulations with heterogeneous boundary conditions, and particularly those with the highest values of *Ra* and q^*, are all capable of reproducing the palaeomagnetic field behaviour captured in GGF100k well. Furthermore, those aspects of field behaviour (*a* and AD/NAD_{median}) that appear to be representative of the very long term average, largely unchanged for 3 billion years (figure 1), are also stable to substantial changes in *Ra* and q^* in the *LowE* simulations. The mechanism by which the outer boundary heterogeneity stabilises the PSV and morphology of the *LowE* simulations is the imposition of large-scale order on both the flow and the magnetic field in the uppermost core (figure 4). Specifically, radial core flow and radial magnetic field are heavily suppressed immediately beneath regions of low core-mantle heat flux associated with the two hot, equatorial large low velocity provinces (LLVPs) seen in seismic tomographic models. The effect in some of these models is to produce regional stratification[8] in lenses a few hundreds of km thick at the top of the core. These regions contribute near-zero radial magnetic field (figure 4) and thereby maintain a similar (small) degree of low-latitude palaeosecular variation in regions beneath areas of low core-mantle heat flux and (large) axial dipole dominance regardless of typical flow speeds and flow scale, which vary with the Rayleigh number. The result is to effectively broaden the region of parameter space where our *LowE* dynamo simulations can produce solutions that satisfy the low-latitude PSV and axial dipole dominance constraints required from the observations (figure 1). Heterogeneity imposed at the outer boundary can reorganise flow throughout the core in high Ekman simulations, but this influence tends to become localised towards the top of the core at low Ekman numbers[34]. Striking a balance between unaltered free convection in the bulk of the core and modified flow from mantle control near the top may explain why we find that the mantle-imposed boundary heterogeneity enables the simulations to satisfy the observational constraints in our *LowE* simulations but not in simulations run with less Earth-like parameters (figure 2).

On the basis of the stability it imparts to the *LowE* simulations (figure 2), we propose strong thermal heterogeneity in the lowermost mantle as a key source of the stability in palaeomagnetic secular variation and field morphology reported here and elsewhere[4]. It is noted, however, that the geodynamo has experienced significant variations, not only in Ra, but also in the ratio of thermal to chemical buoyancy, bottom to internal driving, and shell geometry. While we cannot rule out that such factors help to stabilise a in the real Earth, they do not in our simulations (figure 2). Changes in the large-scale pattern of CMB heat flow, e.g. between the degree 2-dominated present-day planform and degree 1 hemispheric pattern[35], may also affect a. However, our work on non-magnetic convection suggests that these patterns yield similar flow and heat transport properties[8,36]. In terms of bulk dynamics, our simulations are in QG-MAC balance, which is the dynamical regime that prevails in recent high-resolution simulations and seems likely to exist in Earth's core[32]. Since the behaviour in our simulations relies on outer boundary heat flow heterogeneity, which is assigned a geophysically realistic pattern and amplitude[8], and not on viscosity or inertia, which become progressively weaker as Earth's core conditions are approached, we believe we have identified robust behaviour that could potentially operate in Earth's core.

Our results suggest that the regional suppression of core flow immediately beneath LLVPs provides a powerful means of stabilising low latitude palaeosecular variation and thereby maintaining an Earth-like axial dipole dominance in the magnetic fields output by numerical dynamo simulations. Given the observed stability of these aspects of field behaviour evident from palaeomagnetic records, the corollary is that this phenomenon may have been affecting Earth's core for much of geological time. This is plausible since strong heterogeneity in core-mantle heat flux is unlikely to be a recent development. In today's lowermost mantle, LLVPs are bounded by slab graveyards associated with overlying subduction occurring since at least the break-up of Pangaea (~ 0.32 Ga)[37]. The origin, mobility, and dynamic role of these LLVPs are highly controversial topics[37-39]. It does however seem credible that some form of strong large-scale thermal heterogeneity has existed in the lowermost mantle since the onset of whole mantle convection, potentially several billion years ago[40]. The present study requires this. Future analyses may help constrain the specific pattern of this heterogeneity through determining configurations of core-mantle heat flux that allow dynamo simulations to reproduce palaeomagnetic behaviour.

Methods

Palaeosecular variation analyses

Palaeosecular variation (PSV) analyses are performed using virtual geomagnetic poles (VGPs) each associated with the geocentric dipole that best fits a time-instantaneous magnetic field direction obtained at a single location. A total of *N* VGPs from a time series at that location are grouped and the angular dispersion (*S*) calculated using equation M1:

$$S = \left[\frac{1}{N-1}\sum_{i=1}^{N} \Delta_i{}^2\right]^{1/2} \qquad \text{(M1)}$$

Where Δi is the angular distance of the ith VGP from the mean VGP position, calculated using Fisher[41] statistics and corrected for within-site dispersion as per a well-described process[19,42]. All final values of *S* used in this study were calculated following the application of a variable cutoff[43]. This involves excluding outlier VGPs defined as those with $\Delta i > 1.8S + 5°$. Once this is done, *S* and values of Δi are recalculated and further outliers, if any, are also excluded. More iterations are performed, as necessary, until no more outliers are detected.

For the interval 0-0.32 Ga, data were taken at both the site and locality levels from four recent PSV studies spanning the intervals 0-10 Ma[18], 10-25 Ma[20], 73-200 Ma[19], and 200-320 Ma[21] and leaving empty the interval 25-73 Ma for a future study.

For the squares and the diamonds shown in figure 1a (northern and southern hemisphere bins respectively), the large quantity of data available for the 0-0.32 Ga interval allowed a binning approach to be followed. Here, site-level directional data (Supplementary Dataset 1) from the same locality were rotated and relocated together. The rotation was performed about a vertical axis and was of a magnitude and sense as required to produce a mean declination of zero (after first flipping any reversed polarity directions). The relocation involved determining their palaeolatitude such that their new mean VGP was coincident with the palaeogeographic pole. Through this treatment, all 6,178 site-level data were corrected for inferred tectonic motions affecting them since their magnetisations were imparted and were therefore able to be pooled and treated collectively independent of age. Data were placed in 10 palaeolatitudinal bins of 15° width between latitudes ±75° (no data fell in bins with latitude > 75° N or S) and the mean palaeolatitude taken. All data within each palaeolatitudinal bin were then treated as coming from a single locality for the purposes of the PSV analysis as described above (i.e. application of equation M1 and the variable cutoff).

In the case of the circles shown in figure 1, values of *S* and the palaeolatitude were taken directly from the published literature at the locality level (Supplementary Dataset 1). Model G parameters were then calculated using equation (1) and a least squares fit (to the bins for 0-0.32 Ga and to the published datasets for the two older intervals). The 95% uncertainties were obtained using 10,000 bootstraps.

Where the output of dynamo simulations were subject to PSV analysis, this was done using the approach of Sprain et al.[28]. This consisted of downsampling the models (truncated to degree 10) at the modified locations of PSV10[22], randomly selecting the timesteps and extracting VGPs as required to match the number of palaeomagnetic sites in the original datasets. This and the subsequent process of extracting Model G parameters was repeated 10,000 times. The median parameters and 2.5% and 97.5% percentiles were then used for the calculated parameter and its 95% uncertainty bounds.

Estimates of AD/NAD_{median} were calculated according the procedure outlined by Biggin et al.[6]. A time series of *AD/NAD* was calculated at Earth's surface using the Lowes power [44] for the magnetic field energy (*W*):

$$AD/NAD = W_0^1/(W - W_0^1) \qquad \text{(M2)}$$

where

$$W = \sum_{n=1}^{n_{max}} \sum_{m=0}^{n} W_n^m \quad \text{(M3)}$$

and

$$W_n^m = (n+1)[(g_n^m)^2 + (h_n^m)^2] \quad \text{(M4)}$$

where g_n^m and h_n^m are the Gauss coefficients with degree *n* and order *m* of the spherical harmonic expansion.

Inclination anomaly analysis

Inclination anomaly (ΔI) is defined as:

$$\Delta I = I_{obs} - I_{gad} \quad \text{(M5)}$$

where I_{obs} is the observed inclination and I_{gad} is the inclination expected at the given latitude (λ) under the geocentric axial dipole (GAD) assumption whereby $\tan I_{gad} = 2 \tan \lambda$. The calculation of ΔI thus requires an estimate of latitude that is independent of palaeomagnetism which renders it difficult to estimate in deep time[4]. Here we approach the problem indirectly making use of a demonstrated relationship between the value of *a* and the maximum absolute value of ΔI exhibited simultaneously by numerical dynamo simulations (Figure S1).

The maximum $|\Delta I|$ is calculated for each dynamo simulation along *a* according to the process outlined by Sprain et al.[28]. The two parameters are positively correlated as might be expected given the already proven relationship between *a* and AD/NAD_{median}. It is important, however, to note that AD/NAD_{median} reflects the average of many instantaneous field morphologies whereas ΔI reflects the morphology of the field that fits the time-average of many measurements. Over the Earth-like range of a $\lesssim 15°$, the relationship between *a* and the maximum $|\Delta I|$ is reasonably well fit (adjusted $R^2 = 0.71$) by a linear regression. At higher values of *a*, the relationship bifurcates with the majority of simulations following a shallower track while several exhibit very high maximum $|\Delta I|$ (Figure S1a). Here we focus on the well-defined relationship below this bifurcation and use the 95% prediction bounds to infer that values of maximum $|\Delta I|$ appropriate for the Earth, based on the range of *a* derived in figure 1, are in the range of 1.9-10.7°.

Numerical geodynamo simulations

The simulation setup and solution method is the same as our previous work, where detailed descriptions can be found[45-49]. Briefly, an incompressible Boussinesq fluid is rotated with angular frequency Ω in a spherical shell with inner to outer core radius ratio of 0.35. We use no-slip mechanical boundary conditions, an electrically insulating mantle, and a fixed flux outer boundary condition for the buoyancy source in all simulations. The inner core boundary is either insulating or conducting, in which case the solid and liquid are assumed to have same electrical conductivity. Fluid motion is driven either thermally or chemically (Table S1) using the codensity approach whereby a single variable accounts for density perturbations due to both thermal and compositional sources. Thermal cases are driven by internal heating, the configuration relevant to the pre inner core dynamo, or bottom heating, which represents latent heat release at the inner core boundary. Chemical cases have zero flux at the outer boundary (representing no mass exchange with the mantle), a fixed radial flux of light element at the inner boundary, and a uniform internal mass sink that achieves mass conservation of light element.

Some simulations employ lateral variations in the outer boundary heat flux, derived from an expectation that LLVPs are anomalously hot, even if chemical effects are important in setting their anomalous seismic properties. In all these cases, a pattern was chosen to approximate that observed by seismic tomography of the lowermost mantle. Three specific patterns were chosen (referred to as Tomo, RY20 and Z10 in Table S1). The first is itself a tomographic model[30]; the second is a simplified recumbent Y20 pattern[50] that approximates the tomographic planform, and the third is the output of a geodynamical mantle circulation model[39] for the present day after it had been run using 450 Myr of plate reconstructions at the surface. All patterns are characterised by two antipodal and equatorial negative flux anomalies beneath Africa and the Pacific representing the LLVPs. The magnitude of the heterogeneity is varied using the q^* parameter as explained in the main text. Other dimensionless parameters defining the models are the Ekman number (E), Prandtl number (Pr) the magnetic Prandtl number (Pm) and the Rayleigh number (Ra):

$$E = \frac{\nu}{2\Omega d^2} \tag{M6}$$

$$Pr = \frac{\nu}{\kappa} \tag{M7}$$

$$Pm = \frac{\nu}{\eta} \tag{M8}$$

$$Ra = \frac{g\Delta C d}{2\Omega\kappa} \tag{M9}$$

Here, ν, η, and κ are the momentum, magnetic and thermal/compositional diffusivities respectively, *d* is the shell thickness, *g* is gravity at the outer boundary and ΔC is a codensity scale that depends on the convective driving mode[27].

For the purpose of assessing their PSV, morphology and agreement with palaeomagnetic observations, all output magnetic fields were truncated to degree and order 10 and upward continued to Earth's surface. As a check, we experimented with truncating to degree and order 4 but found this only made a difference to AD/NAD_{median} of ≤ 5%.

Parameter selection

For this study we have selected 82 simulations that span a wide range of physical conditions as described in our previous work[27,28]. We have also run 6 new dynamo simulations driven by bottom heating, which we now discuss in detail. The parameters are $Pr = 0.2$, $E = 10^{-5}$, $Pm = 1$, $Ra = 2000, 6000$, $q^* = 0, 2.3, 5$ and a radius ratio of 0.35. The chosen value of Pr is favoured over the value $Pr = 1$ usually adopted in geodynamo simulations because it is closer to the estimates of $\mathrm{Pr} \approx 0.05$ obtained by ab initio calculations of properties of iron alloys[51]. With the chosen value of E the $Ra = 2000$ simulations lie in the rapidly rotating regime of spherical shell rotating thermal convection obtained for $Pr = 1$[52], which is characterised by a strong influence of rotation and vigorous convection, while the $Ra = 6000$ simulations lie towards the lower end of the transitional regime, where the dynamics smoothly deviate from rapidly rotating and towards non-rotating behaviour. Geophysical estimates[53] of E and the supercriticality Ra/Ra_c, where Ra_c is the critical Rayleigh number for onset of non-magnetic convection, suggest that Earth's core lies near the intersection of rapidly rotating and transitional regimes, as in the simulations. The value of Pm is selected in order to achieve strong-field dynamo action at these conditions.

Previous work has suggested that the ratio E/Pm and the magnetic Reynolds number Rm are important for determining the morphological semblance between simulated fields and the modern geomagnetic field. Our simulations yield values of $Rm \sim 1000 - 2000$, comparable to estimates for Earth's core[54], and $\frac{E}{Pm} = 10^{-5}$. These values place our simulations in the wedge-shaped region of $(E/Pm) - Rm$ parameter space that has been suggested to yield simulations that share prominent morphological characteristics with Earth's modern field[55].

From a dynamical perspective the balance of forces that is thought to arise in Earth's core involves a Quasi-Geostrophic (QG) balance at leading order and a Magnetic-Archimedean-Coriolis (MAC) balance at first order[32]. This so-called QG-MAC balance arises in simulations where the magnetic to kinetic energy ratio M exceeds 1[56]. Our $Ra = 2000$ simulations produce $M \approx 3$ while our $Ra = 6000$ simulations produce $M \approx 1$ (except the multipolar solution with $q^* = 0$ where $M < 1$) and hence are in QG-MAC balance.

Recent work has shown that it is possible to construct a uni-dimensional path in parameter space that links the currently accessible conditions in dynamo simulations to conditions approaching those in Earth's core[32]. Starting from a direct numerical simulation that is rapidly rotating ($E \sim 10^{-5}$) and in QG-MAC balance, progression along the path towards Earth's core conditions (using large eddy simulations at the most demanding parameter combinations) revealed that the spatial structure of the solution, the magnetic Reynolds number, and the scale at which MAC balance is achieved remain largely invariant while inertial and viscous effects became increasingly subdominant. These results were obtained for suites of simulations with and without outer boundary forcing. Our simulation parameters are close to those near to the start of the "path" simulations and hence we expect that the large-scale behaviour we see, which is dominated by Magnetic, Coriolis and buoyancy effects, will also remain invariant were we to progress to more extreme conditions. However, such an effort is impractical for our purpose, which relies on long simulations that capture variations on paleomagnetic timescales. Indeed, our suite of 6 simulations required 14.6 million CPU hours on national high-performance computing resources.

The values of q^* adopted in this study are consistent with estimates for the present-day Earth. Some mantle convection simulations suggest that q^* is relatively invariant under steady state convection[57] and so in the absence of further information we assume that our chosen range of q^* are representative of variations in the past. For $q^* > 2$ convection becomes suppressed in regions at the top of the core that are spatially related to the imposed heat flow pattern. According to our previous analysis of non-magnetic convection, the thickness of these regions depends very weakly on E, Pr and Ra, while the strength of stratification (when it exists) is independent of inertia and viscosity[58]. We therefore expect their properties to be maintained at more extreme simulation conditions.

We have adopted a simple driving mechanism that represents latent heat release at the inner core boundary. The main advantage of this choice is it allows us to leverage the significant insights into rapidly rotating boundary-forced spherical shell convection from our previous non-magnetic studies [8,36,58]. One alternative is to adopt a codensity approximation that combines thermal and chemical buoyancy. This, however, assumes that the temperature and composition have the same diffusivities and satisfy the same boundary conditions, which is almost certainly not the case in Earth's core (e.g. [51]). A better approach is to solve thermal and compositional fields separately[59], but at the expense of introducing further dimensionless parameters (Rayleigh and Prandtl numbers) that cannot be set to Earth-like values. Tassin et al. [59] show that morphological agreement between simulations and Earth's modern field can be obtained for any ratio of thermal to chemical driving, while Figure 2 of the main text shows that the behaviour of *a* is similar between purely thermal and purely chemical cases without boundary heterogeneity, suggesting that the treatment of thermo-chemical buoyancy is not the primary explanation for the disparity in *a* behaviour. Ideally, one would conduct a suite of simulations that represent distinct epochs in the history of the core, each with distinct thermochemical driving properties [26,60]. However, the evolution of thermal and chemical buoyancy in the core is poorly constrained and so the results would lack generality. Instead, we have exploited the wealth of existing simulations with generic buoyancy sources and distributions, which allows us to isolate individual effects on *a*. Finally, we note that the dynamics at high q^* near the top of the core are dominated by the boundary-forced flows[34,61] and are therefore probably less influenced by the nature of the homogeneous buoyancy source. Given these considerations and the vast computational expense of conducting QG-MAC simulations under rapid rotation, we argue that considering a single driving mechanism is a reasonable simplification.

Estimation of possible Ra variation through geological time

In Figure 2 of the main text, our "tracks" of simulations, in which only the value of Ra is varied, may be loosely interpreted as representing the temporal evolution of the core dynamo. To justify this interpretation, we note that the Prandtl numbers Pr and Pm are determined by the viscous, thermal and magnetic diffusion coefficients, which have probably varied only weakly over geological time as the core cooled. The shell thickness, d, has varied from its current value of $d = 2264$ km to $d = 3480$ km prior to inner core formation. The rotation rate of the Earth, Ω, has decreased through time, such that the length-of-day was about 18 hours 2.45 billion years ago[62]. These effects combine to increase the Ekman number by a factor of 3 over the span of geological time we consider. We nevertheless consider the primary change in control parameters over time is through Ra as discussed below.

For Earth's core there are two Rayleigh numbers: the thermal Rayleigh number Ra and the compositional Rayleigh number Ra_C. The equivalent definition of Ra to that used in our thermal simulations is

$$Ra = \frac{\alpha g\left(\frac{dT'}{dr}\right)d^2}{2\,\Omega\kappa} \qquad \text{(M10)}$$

Where dT'/dr is the difference between the CMB temperature gradient $\frac{dT}{dr} = -Q_{cmb}/(4\pi r_c^2 k)$ and the adiabatic gradient $dT_a/dr = \alpha g T/c_p$. Here Q_{cmb} is the (time-dependent) CMB heat flow, k is the thermal conductivity and $r_c = 3480$ km is the CMB radius. We assume that the material properties have constant values[54]: $\alpha = 2 \times 10^{-5}$ /K, $g = 10.6$ m^2/s, $\kappa = 10^{-6}$ m^2/s, $c_p = 860$ J/kg/K. The core thermal conductivity has recently been reviewed in detail [63], who concluded that recent ab initio and experimental works favour a value of k in the range $70 - 90$ W/m/K and we initially take the upper end of this range. Estimating Ra requires the time variation in rotation rate Ω, and shell thickness, d, which we have discussed above, and dT'/dr.

Estimating dT'/dr requires knowledge of Q_{cmb} and the CMB temperature T. Neither can be observationally constrained in the past so we must defer to models. Some mantle convection simulations predict temporal variations in Q_{cmb} of over an order of magnitude[64], which yields a large change in Ra because dT_a/dr only varies with the core temperature, which is predicted by all thermal history models to have changed by less than a factor of 2 over Earth's history [54,65,66]. However, the key constraint is provided by Figure 2 of the main text, which shows that all of our dynamo simulations with homogeneous boundary conditions remain in the Earth-like range of Model G *a* over a narrow range of Ra. If the value of Ra/Ra_c at which the simulation track enters the Earth-like range for Model G *a* is Ra_{in} then by $Ra = 2Ra_{in}$ all homogeneous dynamos have left the Earth-like region. We therefore seek the smallest plausible changes in Ra over Earth's history, acknowledging that the real changes were probably larger and hence present an even greater difficulty for the simulations in Fig 2a and b.

The thermal history model of Driscoll and Bercovici[67] couples core and mantle evolution. It predicts a drop in Q_{cmb} from 60 TW to 14 TW over the last 4 Gyrs and a drop in T from 5000 K to 4000 K over the same period, which corresponds to a factor ~200 fall in Ra. The model of Davies et al.[54] is designed to produce a lower bound on core cooling because the dissipation available to the dynamo is set to zero before inner core formation. These models predict a change in CMB heat flow of about 5 TW of

4.5 Gyrs and a 1000 K temperature drop, yielding a factor 14 change in Ra. With the lower k values these ratios drop to 23 and 3.5 respectively. Therefore, we conclude that the change in thermal Rayleigh number over the course of Earth history is at least a factor of 3 and could be at least at order of magnitude larger.

The compositional Rayleigh number Ra_C is harder to estimate than Ra because it depends on the rate at which light elements are expelled from the solid inner core on freezing. The present value of Ra_C has been estimated as 10^{38} (ref-[53]), which is enormously supercritical. However, Ra_C must trend to zero at inner core nucleation and hence varies even more than Ra over Earth's history. This argument ignores the possible precipitation of light elements at the top of the core [68], but the dynamics of this scenario are yet to be investigated in detail and hence no constraints are available.

Derivation of $Q_{PM\text{-}100k}$ criteria

The structure and manner of application of the $Q_{PM\text{-}100k}$ criteria exactly mimics that of the published Q_{PM} criteria[28] and will only be covered briefly here. Both sets of criteria were developed to test the adherence of the magnetic field behaviours output by numerical geodynamo simulations to palaeomagnetic observations. The only difference is the time interval over which the comparison is made. The original Q_{PM} criteria are based on observations made over the interval 0-10 Ma and are therefore most suitable for long-running dynamo simulations featuring multiple reversals and excursions and associated large variations in dipole moment. Our new $Q_{PM\text{-}100k}$ criteria are rather based on the interval 0-100 ka and, specifically, the palaeomagnetic field model GGF100k[10]. This was considered appropriate because the computationally expensive nature of the *LowE* models does not allow them to run sufficiently long to capture the full range of variability of the geomagnetic field on a multi-million year timescale.

The criteria themselves (Table S2) are based on five parameters that can be obtained directly from palaeomagnetic datasets. The first three of these were Model G *a* and *b* parameters and the maximum inclination anomaly all taken directly from the PSV10 compilation in the case of Q_{PM} and a PSV10-like resampling in the case of $Q_{PM\text{-}100k}$. For maximum inclination anomaly and Model G *b*, 95% confidence limits (Table S2) were calculated using a bootstrapping technique. The uncertainty in estimating Model G *a* for GGF100k is substantially lower (<0.1°) than observed from PSV datasets derived directly from paleomagnetic records, likely due to the regularization and smoothing employed in constructing GGF100k. For that reason, the uncertainty in estimating Model G *a* from PSV10 (±2°) is retained instead. The fourth criterion (*V%*) reflects the normalised variance of virtual dipole moment measurements. For the original Q_{PM} criterion, the lower and upper limits were taken from records from the PINT database pertaining to the intervals 0-1 Ma and 1-10 Ma respectively. The last 100 kyr displays somewhat lower variance (Table S2); the lower limit for the $Q_{PM\text{-}100k}$ V% criterion was obtained from the 95% bootstrap uncertainty while the *V%* defined from the entire Brunhes chron (used as the lower limit in the original Q_{PM} V% criterion) was assigned as the upper limit for $Q_{PM\text{-}100k}$. The final criterion is based on the proportion of time that the field was in a transitional state (τ_t) such that the geomagnetic pole was between -45 and +45° latitude. This was bounded in the original Q_{PM} criterion by assuming 10 excursions and 5 reversals per Myr each with a duration of between 2.5 and 10 kyr. The model GGF100k contains no timesteps when the field was transitional by this definition, due to data selection and the regularization procedure employed smoothing out excursional signals (cf. LSMOD.2, ref [69]), but we conservatively set the upper bound to allow 5 kyr of excursions. We note that LSMOD.2 was specifically constructed to highlight excursional features spanning 30-50 ka and yields a τ_t of ~3.2 kyr when scaled to a duration of 100 kyr.

Compliance of each dynamo simulation with $Q_{PM\text{-}100k}$ is based on values of normalised misfit $\Delta Q^{i}_{PM-100k}$:

$$\Delta Q^{i}_{PM-100k} = \frac{|m^{i}_{Earth} - m^{i}_{Sim}|}{\sigma^{i}_{Earth} + \sigma^{i}_{Sim}} \qquad \text{(M11)}$$

where m^{i}_{Earth} and m^{i}_{Sim} are the median values of the *i*th parameter derived from Earth (Table S2) and the simulation respectively. Similarly, σ^{i}_{Earth} and σ^{i}_{Sim} are the appropriate upper or lower uncertainty bounds for Earth and the simulation respectively. Thus, if the difference between the measured and expected results for a criterion is smaller than the relevant uncertainties then $\Delta Q^{i}_{PM-100k}$ is less than 1 and the simulation can be said to have passed that criterion. In the case of τ_t, the lower uncertainty bound for $\sigma^{\tau_t}_{Earth}$ is set to 0, such that simulations with a $\tau_t \leq 2.5$ kyr have a $\Delta Q^{\tau_t}_{PM-100k} = 0$.

Median and uncertainty values for simulation parameters (i.e. m^{i}_{Sim} and σ^{i}_{Sim}) were calculated following an identical approach to that used previously[28]. This entailed each simulation being downsampled using a Monte Carlo approach according to the distribution of sites represented in PSV10[22]. This was repeated 10,000 times such that 95% confidence intervals could be calculated.

The individual misfit values are given alongside the summed $Q_{PM\text{-}100k}$ score (total number of criteria passed) in Table S1.

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Author Contributions

AB derived the concept, performed analyses and wrote the paper. JM, CD, and RB ran the simulations, performed analyses and wrote the paper.

Data availability Statement

All palaeomagnetic datasets are included as Supplementary Dataset 1. Other datasets analysed during the current study are available from the corresponding author on reasonable request.

Code availability Statement

Code used for the analysis during the current study are available from the corresponding author on reasonable request. Correspondence and requests for materials should be addressed to: biggin@liverpool.ac.uk.

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