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Sustaining Earth's magnetic dynamo

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ABSTRACT

Earth's magnetic field is generated by fluid motions in the outer core. This geodynamo has operated for over 3.4 Gyrs. However, the mechanism that has sustained the geodynamo for over 75% of Earth's history remains debated. In this Review, we assess the mechanisms proposed to drive the geodynamo (convection, precession and tides) and their ability to match geomagnetic and paleomagnetic observations. Flows driven by precession are too weak to drive the geodynamo. Flows driven by tides could have been strong enough in the early Earth, prior to 1.5 Ga, when tidal deformation and Earth's spin rate were larger than today. Evidence that the thermal conductivity of Earth's core could be as high as $250 \text{ W m}^{-1} \text{ K}^{-1}$ questions the ability of convection to maintain the dynamo for >3.4 Gyrs. Yet, convection can supply enough power to sustain a long-lived geodynamo if the thermal conductivity is lower than $100 \text{ W m}^{-1} \text{ K}^{-1}$. Exsolution of light elements from the core increases this upper conductivity by 15% to 200%, based on the exsolution rates reported so far. Convection, possibly aided by the exsolution of light elements, remains the mechanism most likely to have sustained the geodynamo. The light-element exsolution rate, which remains poorly constrained, should be further investigated.

Key points:

- Numerical models of the geodynamo driven by thermo-chemical convection account for most of the observed properties of the present geodynamo.
- The thermal conductivity in Earth's core remains debated, with published values ranging between 20 and $250 \text{ W m}^{-1} \text{ K}^{-1}$. With a conductivity as high as $250 \text{ W m}^{-1} \text{ K}^{-1}$, motionless heat transport prevails in the core, hence convection cannot sustain Earth's magnetic dynamo for 3.4 Gyrs.
- Still, thermo-chemical convection due to the slow cooling of Earth supplies enough power to the geodynamo when the thermal conductivity is lower than $100 \text{ W m}^{-1} \text{ K}^{-1}$. The exsolution of light elements increases this limit only marginally or by up to a factor 3 depending on the exsolution rate.
- Flows driven by precession are too weak to drive the geodynamo.
- Flows driven by tides could have been strong enough prior to 1.5 Ga, when tidal deformation and Earth's spin rate were larger than today. This calls for further investigation of tidally-driven dynamos.

Website summary: The mechanisms that sustain Earth's long-lived geodynamo remain under scrutiny. This Review assesses the potential candidates – convection, precession and tides – revealing that convection, possibly helped by the exsolution of light elements, is the most likely scenario.

1 Introduction

The Earth's surface is immersed in a magnetic field that is mainly a dipole. The geomagnetic field deviates charged particles from the Sun, which may help to protect Earth's atmosphere from erosion by solar winds¹⁻³. Humans have used magnetic compasses to navigate the oceans and continents for thousands of years. Other animals, including sea turtles and salmon, are also suspected to use the magnetic field for navigation^{4,5}. Magnetotactic bacteria, which swim along magnetic field lines^{6,7}, rely on Earth's magnetic field to migrate up and down in the sediment column.

The magnetic field is hence an essential aspect of Earth. Yet, the origin of this field long remained a mystery. In 1919, it was hypothesised that the geomagnetic field originates from flow motions of conducting material inside the Earth through the

9 so-called *dynamo effect*⁸ (box 1). Over a century later, it is now widely accepted that the geodynamo operates in the outer core,
10 which is made of flowing liquid metal (figure 1a)⁹.

11 The most accepted scenario to drive turbulent core flows and the geodynamo is natural convection⁹ due to the slow cooling
12 of the Earth (figure 1b). Numerical simulations of a convection-driven geodynamo produce a magnetic field that resembles that
13 of the Earth¹⁰ and they are coming close to the conditions of Earth's core^{11–15}. Therefore, the behaviour of the geomagnetic
14 field at Earth's surface can be explained by the convective dynamics deep inside the core^{14,15}.

15 Magnetized rocks indicate that the geodynamo is at least 3.4 Gyrs old¹⁶. Whether convection can sustain the geodynamo
16 for such a long time critically depends on the thermal conductivity of the core¹⁷. However, the value of core conductivity
17 is highly debated, with published values^{18–26} ranging between 20 and 250 W m⁻¹K⁻¹. With high conductivity values of
18 100 – 250 W m⁻¹K⁻¹ found by experimental and numerical investigations^{23–26}, much of the core heat flow would escape by
19 conduction, leaving little to drive thermal convection, especially in the distant past when the core was fully molten^{17,27,28}. This
20 disconcerting result, coined the *new core paradox*¹⁷, motivated the search for other mechanisms to drive the geodynamo.

21 A possible mechanism is the exsolution of light elements from the core (figure 1c). Exsolution leaves behind a denser
22 liquid, which produces natural convection and helps to sustain a convective geodynamo^{29–31}. Tides and precession are also
23 alternative mechanisms^{32–35} that can produce turbulent flows in the outer core^{36–39}, and hence could drive a dynamo (figure 1de).
24 Advances in simulations of precession-driven dynamos since the 2000s^{40–43} have generated interest in a geodynamo powered
25 by orbital forcing⁴⁴. However, it remains debated whether the exsolution of light elements, precession or tides produce flows
26 that are strong enough to solve the new core paradox.

27 In this review, we compare the candidate driving mechanisms for the geodynamo - convection, precession and tides
28 (figure 1bcde). We first summarise the main properties of Earth's dynamo that are deduced from geomagnetic and paleomagnetic
29 observations. For each driving mechanism, we discuss the latest simulations and evaluate their compliance with the present-day
30 geomagnetic field. We then examine whether the power or kinetic energy produced by each mechanism over geological time
31 can sustain the long-lasting magnetic field observed in paleomagnetic data. In the future, better understanding of the early
32 geodynamo requires further investigation of the flow and magnetic fields produced by convection or tides in a fully liquid core,
33 with no inner-core. Additional constraints on the rate at which light elements have exsolved from the core throughout Earth's
34 history are also needed.

35 2 Geomagnetic and paleomagnetic observations

36 Our understanding of Earth's dynamo rests on the analysis of the present and past geomagnetic field. The direct observation
37 of the geomagnetic field since the seventeenth century places important constraints on the working of the geodynamo on
38 interannual to secular time scales. Paleomagnetic studies provide additional information on the mechanisms driving the
39 geodynamo on millennial to geological time scales.

40 2.1 From the Age of Sail to the present day

41 Nowadays, satellites and ground-based observatories provide us with a global and time-dependent map of the Earth's magnetic
42 field⁴⁵. The main field, of about 50 μT at the Earth's surface⁴⁶, is generated in the Earth's core⁴⁷. This field is dominated by a
43 dipole slightly tilted with respect to the axis of rotation of the Earth, with a dipole moment of about 8×10^{22} A m².

44 To trace the variations of the geomagnetic field, one has to recourse to the archives of magnetic observatories, some of
45 which date back to the seventeenth century. Additional information can be obtained from mariners' logbooks, in which the
46 direction of the magnetic north pole was reported during voyages⁴⁸. This information is available from the end of the sixteenth
47 century onward. The rate of change of the geomagnetic field, called the *geomagnetic secular variation*⁴⁹, displays 6 yrs period
48 oscillations during the last century⁵⁰. These oscillations reflect the propagation of hydromagnetic waves, called *Alfvén waves*,
49 in the outer core. As their propagation speed is proportional to the magnetic field, the field strength deep in the core can be
50 estimated to ~ 4 mT⁵¹. This value is about 10 times the field strength at the core-mantle boundary (CMB), and it corresponds
51 to a magnetic energy⁵² of $\sim 10^{21}$ J inside the core.

52 Beyond waves, flows inside the core transport magnetic field lines. This process explains most of the decadal to secular
53 fluctuations of the geomagnetic signal⁵³. These variations suggest a large-scale velocity of $\sim 3 \times 10^{-4}$ m s⁻¹ below the core
54 surface. Assuming that this figure is representative of the flow in the bulk of the outer core yields a kinetic energy of 8×10^{16} J,
55 about 10^4 times smaller than the magnetic energy⁵¹.

56 2.2 Over geological time

57 Prior to the sixteenth century, observations of the geodynamo rely on the remnant magnetization carried by rocks or archeological
58 recorders^{54,55}. In the laboratory, it is possible to recover the direction or amplitude of the geomagnetic field that reigned when
59 the magnetization was acquired. These indirect measurements allow geologists to determine the motion of continents over
60 hundreds of millions of years⁵⁶. They also open a window on the behaviour of the geodynamo over several billions of years.

61 One of the most fundamental question for paleomagnetic studies is the age of the geodynamo. Magnetic inclusions in
62 extremely old minerals from South Africa provide robust evidence that the geodynamo was active 3.4 Gyrs ago¹⁶. Analyses of
63 samples from Greenland suggest that the geodynamo had started 3.7 Gyrs ago⁵⁷. An even older dynamo record from 4.2 Gyrs
64 ago has been proposed⁵⁸ but this finding remains controversial⁵⁹.

65 The geodynamo has evolved on multiple time scales (figure 2). Over the last millennia, the dipole strength varied between
66 0.7 and 1.4 times its present-day value (figure 2a). Earth's magnetic field has frequently reversed its polarity (figure 2b,c). The
67 frequency of reversals is highly variable on time scales of a hundred million years^{54,60}. During the last few million years, the
68 field reversed 4 times per million year, but further in the past, there were periods of several million years with no reversal
69 (figure 2c). Thanks to the increase in quality and number of paleointensity measurements, it is now possible to search for
70 long-term trends in the paleointensity signal over the last 3 billion years⁶¹⁻⁶³ (figure 2d). Paleomagnetic data show that the
71 geodynamo produced a strong, seemingly dipole-dominated field, with a moment of $5 \pm 2 \times 10^{22}$ A m² during more than 75%
72 of Earth's history^{61,62,64,65}.

73 3 Driving mechanisms

74 In this section, the possible driving mechanisms for the geodynamo are introduced. Their ability to produce a field resembling
75 the present-day geomagnetic field are assessed in the light of numerical simulations.

76 3.1 Cooling and inner-core growth

77 The Earth is cooling down. At present, a heat flow of 46 ± 3 TW escapes at the solid Earth's surface⁶⁶, while ~ 18 TW are
78 produced by the disintegration of radioelements within the mantle and crust^{67,68} and less than 2 TW by radioelements within
79 the core²⁷.

80 The cooling of the Earth is the most accepted mechanism to drive the present-day geodynamo^{52,69-71}. As the core slowly
81 cools down, the solid inner core grows, releasing latent heat and light elements⁷² at the inner-core boundary (figure 1b). The
82 light fluid rises to the CMB, thereby generating vigorous plumes in the outer core⁷³ (figure 1a). Meanwhile, as heat leaves the
83 core, colder fluid forms at the CMB and sinks downwards (figure 1b).

84 Earth's rotation strongly influences these convective motions, which take the form of swirling columns aligned with the
85 rotation axis⁷⁴⁻⁷⁸ (figure 1a). These swirling flows are prone to dynamo action⁹ (box 1). The dipolar geomagnetic field, and its
86 close alignment with the rotation axis, are well explained by these columnar motions⁷⁹ (figure 1a).

87 The first self-consistent numerical dynamo driven by convection was obtained in 1995^{80,81}. Simulations soon could account
88 for several features of the observed geomagnetic field^{10,82} (figure 3a), including its dipolar dominance, patches of enhanced
89 magnetic field at high latitudes (figures 3b and 3c) and polarity reversals (table in box 2). With the increase in computing
90 power, a large number of simulations could be run. Parametric exploration led to the derivation of scaling laws that relate the
91 power from convection to the magnetic and flow intensity^{83,84}. These laws predict a magnetic field intensity of ~ 1 mT inside
92 the core, in line with the estimates from geomagnetic observations (section 2.1).

93 In turbulent geodynamo simulations^{11,12,15}, the intensity of the magnetic field inside the core is about 10 times stronger
94 than at the CMB, as inferred for the Earth⁵¹. The hierarchy of dominant forces is now the same as in Earth's core^{12,15,85}.
95 Simulations also produce fast waves interacting with the slower convective motions^{14,15,86}. Thanks to results from these
96 simulations, abrupt changes in the surface field (termed geomagnetic jerks) are now interpreted as the arrival of waves that are
97 excited by convective plumes in the core^{14,15}. Convective dynamos generate a time evolving field that explains most of the
98 observed secular variations (table in box 2).

99 3.2 The exsolution of light elements

100 In addition to iron and nickel, the outer core contains lighter elements, including Si, O, S, H or Mg⁷². The partitioning of some
101 light elements in iron increases with temperature^{30,87-89}. The Earth's core experienced high temperatures > 5000 K during
102 its formation, when large impacts brought new metal and silicates to the growing planet^{90,91}. After each collision, the newly
103 brought metal sank into the mantle, where it incorporated light elements, and then merged with the Earth's core^{30,72}.

104 Subsequently, on geological time, the core cooled down. For each light element with a given initial concentration, there
105 is a critical temperature below which exsolution starts²⁹⁻³¹, leaving an iron-rich liquid that sinks into the deeper core. These
106 convective motions can contribute to the geodynamo not only today but also in the distant past, prior to the growth of the inner
107 core²⁹⁻³¹ (figure 1c).

108 Which light element exsolves from the core is still debated. Some high-pressure experiments and molecular dynamics
109 simulations suggest that magnesium oxides MgO would be the first exsolved species^{30,31,92,93} while others favour silicon
110 oxides SiO₂^{89,94}. One study⁹⁵ found that the liquid Fe-Si-O system separates into immiscible liquid alloys as the core cools
111 down, instead of exsolving SiO₂ oxides, but the robustness of these results has been questioned⁹².

Of paramount importance is the depth at which exsolution starts. Were it at the bottom, a dense iron-rich liquid would accumulate there, generating a stable stratification but no convection. Several studies^{88,93} find that the exsolution of MgO starts at the top of the core, and hence can drive convection. For SiO₂, the uncertainties on the solubility curve are still too large to determine the depth at which its exsolution starts⁸⁹.

The exsolution of light elements at the top of the core can drive the geodynamo through convection. Thus, the results from dynamo simulations driven by thermo-chemical convection (section 3.1) to some extent apply to exsolution. In a dynamo simulation, exsolution can be modelled by volumetric buoyancy sources that are balanced by a buoyancy flux at the CMB. However, only few^{96–98} have investigated this configuration, which is also appropriate for thermal convection driven by the slow cooling of the core prior to the growth of an inner core.

3.3 Precession and tides

Precession and tides can also trigger flow motions and dynamo action. Precession, a slow variation of the orientation of Earth's rotation axis (figure 1d), forces the fluid core to rotate along a different axis than the mantle^{99–101}. At present, precession induces a diurnal differential motion of about 60 m at the CMB and a velocity $u_p \sim 4 \text{ mm s}^{-1}$. Because the core is not perfectly spherical, precession also triggers a weak secondary flow¹⁰² of amplitude $u_p f_p$, where $f_p \approx 1/400$ is the CMB ellipticity¹⁰³. Tides produce a diurnal CMB deformation of ellipticity $f_t \approx 10^{-7}$ moving around the core at speed $u_t = 250 \text{ m s}^{-1}$ (figure 1e). These flows do not produce magnetic fields by themselves (box 1)^{41,104}. Still, instabilities near the boundary or in the bulk core can lead to turbulent flows and dynamo action^{40,43,105}.

Simulations of dynamos driven by precession^{40,43,105} or tides¹⁰⁶ have yet to provide scaling laws for the intensity of the magnetic field. Furthermore, the magnetic field is always weak and small-scale^{43,105,106} (figure 3d), and hence very different from the dipolar geomagnetic field (figure 3a). In ellipsoids, simulations of large-scale magnetic fields were initially reported¹⁰⁷, but their validity was subsequently questioned¹⁰⁸. Currently, it is not known whether orbital forcings are able to produce a large-scale, dipolar magnetic field with an amplitude compatible with that of the geomagnetic field. Lowering viscosity in future simulations might answer this question.

While low viscosity is hard to reach in simulations, it is a feature of laboratory experiments that use liquid sodium (see Supplementary Fig.1)^{109–112}. No experimental dynamos driven by precession or tides have yet been published. However, all eyes are on the DRESHDYN dynamo experiment, which is currently being built and will be driven by precession¹¹³.

4 Sustaining a convective geodynamo

Paleomagnetic observations indicate an active geodynamo during the last 3.4 Gyrs. In this section, the energetics of the core are used to assess whether convection can power such a long-lived dynamo.

4.1 Thermal conductivity

Mantle convection sets the heat flow that escapes from the core^{52,66}. A substantial portion of this heat is transported by thermal conduction in the core and does not participate to core convection. The higher the conducted heat flow, the lower the power available for the geodynamo.

The conducted heat flow, often called *isentropic* heat flow, reads at the CMB

$$Q_{\text{is}} = -k \frac{\partial T}{\partial r} \mathcal{A}, \quad (1)$$

where the temperature gradient is

$$\frac{\partial T}{\partial r} = -\frac{\alpha g T}{C_p} \quad (2)$$

in an isentropic, well-mixed core¹¹⁴, k is the thermal conductivity, \mathcal{A} the area of the CMB, α the coefficient of thermal expansion, C_p the specific heat capacity, and g the acceleration of gravity at the CMB.

The isentropic heat flow is therefore proportional to the thermal conductivity k , whose range of published values broadened since the early 2010s^{23–25,115}. Direct experimental measurements suggest low values^{18–20} of k between 20 and 46 $\text{W m}^{-1} \text{K}^{-1}$, in line with older estimates^{21,22}, while other studies find larger values of k ^{23–26,116} between 90 and 250 $\text{W m}^{-1} \text{K}^{-1}$. This scatter is partly due to the relation between electrical and thermal conductivity which remains to be clarified at high pressure and temperature¹¹⁷. Adding light elements in the core lowers the thermal conductivity²⁰, further increasing its uncertainty.

Here, the impact of two end-member values on the geodynamo are examined: a high-conductivity $k \simeq 100 \text{ W m}^{-1} \text{K}^{-1}$, which results in a conducted heat flow $Q_{\text{is}} \simeq 15 \text{ TW}$, and a low-conductivity $k \simeq 40 \text{ W m}^{-1} \text{K}^{-1}$, which results in $Q_{\text{is}} \simeq 6 \text{ TW}$.

4.2 Convection can power the current geodynamo

In the Earth's core, the magnetic field generates electric currents, which dissipate energy by ohmic heating. Theory, numerical models and experiments suggest that the power lost today by ohmic dissipation is on the order of 1 TW^{9,118,119}. The driving mechanism of the geodynamo must therefore supply enough power to balance this ohmic dissipation.

From the energy and entropy budgets of the core, one can estimate the power originating from convection^{27,28,120–123}. As shown in this section, a high-conductivity value does not prevent a convective dynamo at the present day. Regardless of the exact value of the thermal conductivity, the outer core still solidifies into an inner core, driving convection and dynamo action.

At the CMB, thermal convection is possible when the core heat flow Q_{cmb} exceeds the conducted heat flow Q_{is} . The mass anomaly flux that drives convection at this boundary is^{120,121}

$$F_o = \frac{\alpha}{C_p}(Q_{\text{cmb}} - Q_{\text{is}}). \quad (3)$$

At the base of the outer core, the slow solidification of the inner core drives convection with a mass anomaly flux^{120,121}

$$F_i = 4\pi r_i^2 \dot{r}_i \left(\Delta\rho + \frac{\alpha\rho L}{C_p} \right), \quad (4)$$

where ρ is the mean density of the outer core, $\Delta\rho$ the density deficit due to the release of light elements, r_i the inner-core radius and L is the latent heat released by the freezing of the inner core. Equation (4) assumes an isentropic heat flow through the inner-core boundary¹²⁰. The two terms in equation (4) describe the flux of light elements due to inner-core growth and the flux of latent heat. Both are proportional to the inner-core growth rate \dot{r}_i . The growth rate can be estimated from the energy balance of the core. After the onset of inner-core growth, this balance reads¹²⁴

$$Q_{\text{cmb}} = \dot{r}_i P(r_i) + Q_r, \quad (5)$$

where the function P gathers the contributions from latent heat release at the inner-core boundary, gravitational energy release, and cooling of the core, which are all functions of r_i ¹²⁴, and Q_r is the radiogenic heat production term. The heat flow through the CMB, Q_{cmb} , must be balanced by those contributions internal to the core. At the present day, assuming $Q_r = 0$, the growth rate is $\dot{r}_i \simeq 600 \pm 250 \text{ km Gyr}^{-1}$ ²⁷ and F_i is in the range $1 \times 10^5 - 3 \times 10^5 \text{ kg s}^{-1}$. Differences in the function P between studies propagate as an uncertainty of 30% on the age of the inner core.

The convective power Φ_{conv} available for the dynamo is approximated by^{120,121}

$$\Phi_{\text{conv}} = F_i(\bar{\psi} - \psi_i) + F_o(\psi_o - \bar{\psi}), \quad (6)$$

where ψ_i and ψ_o are the gravitational potential at the inner-core and core-mantle boundaries, and $\bar{\psi}$ is the mass-averaged gravitational potential in the outer core¹²⁰. Equation (6) shows that the convective power originates from taking mass anomalies at the gravitational potential of the inner or outer boundary, and redistributing it throughout the outer core at the mean gravitational potential.

Expressions (3)-(6) contain several approximations^{120,121}, including the assumption of a well-mixed core. The errors associated with these approximations are comparable to that of the Boussinesq approximation, which is $\sim 10\% - 15\%$. More precise expressions^{27,28,125–127} lead to conclusions similar to those drawn below.

The inner-core growth rate \dot{r}_i , and consequently the convective forcing F_i at the inner-core boundary, do not depend on the thermal conductivity (equations 5 and 4). When the inner core is growing, the convective power Φ_{conv} (equation 6) therefore remains large regardless of the value of k . Even with a high thermal conductivity $k > 100 \text{ W m}^{-1} \text{ K}^{-1}$, the convective power is larger than 1 TW at present (dotted lines at time 0 Gyr in Figure 4a,b). With a core heat flow $Q_{\text{cmb}} = 10 \text{ TW}$, the available convective power is 1.5 TW with $k = 100 \text{ W m}^{-1} \text{ K}^{-1}$ and 2.5 TW with $k = 40 \text{ W m}^{-1} \text{ K}^{-1}$ (time 0 Gyr in figure 4b). The power raises up to 3–4 TW when assuming a large heat flow of 15 TW escaping from the core (time 0 Gyr in figure 4a).

4.3 The new core paradox

Trouble comes prior to the onset of inner-core growth. At the present day, the contribution F_i (equations 4) from the solidification of the inner core represents more than 70% of the mass anomaly flux that drives convection. Before the nucleation of the inner core, the major power contributor to the current geodynamo was therefore missing. The only power source that remained was thermal convection due to the slow cooling of the core. The core heat flow must then exceed the conducted heat flow to drive convection, as shown by expressions (3) and (6). However, the core heat flow ranges within 6–17 TW^{128–132} while the conducted heat flow is larger than 15 TW when the thermal conductivity is larger than $100 \text{ W m}^{-1} \text{ K}^{-1}$. The heat budget of the core is hence very tight when $k \gtrsim 100 \text{ W m}^{-1} \text{ K}^{-1}$. This serious problem was termed the *new core paradox*¹⁷ as it comes after

187 the Higgins-Kennedy core paradox raised and solved in the 1970s^{133,134}. The growth of the inner core likely started less than 1
 188 Gyr ago^{27,28,125,132}. The new core paradox therefore impacts more than 2.4 Gyrs, hence 70% of the observed history of Earth's
 189 magnetic field.

190 The convective power Φ_{conv} must necessarily be positive to sustain a convective dynamo. More stringent requirements are
 191 obtained from paleointensity measurements⁶¹ (figure 2d) combined with scaling laws from dynamo simulations⁸⁴ (details in
 192 Supplementary Note 1). These require the ohmic dissipation, and hence the convective power Φ_{conv} , to be in the range 0.1 – 10
 193 TW over Earth's history (shaded region in figure 4a,b). Another condition is that the convective velocity U must be large
 194 enough so that the magnetic Reynolds number $Rm = UD/\eta$ (box 1) exceeds about 100⁸³, where η is the magnetic diffusivity
 195 of the core and D the outer core thickness.

196 Prescribing the time-evolution of the core heat flow $Q_{\text{cmb}}(t)$ and the radiogenic heating term $Q_r(t)$, one can integrate
 197 equations (3)-(6) backward in time to estimate the convective power in the past. Here, it is assumed that $Q_{\text{cmb}}(t)$ increases
 198 linearly in the past, and that $Q_r = 0$ at all times. With the evolution of the convective power, dynamo scaling laws^{12,135} are
 199 used to estimate the convective velocity U , and hence the magnetic Reynolds number Rm . Figure 4 shows the convective power
 200 and the magnetic Reynolds number from this evolutionary model.

201 Assuming a low thermal conductivity $k = 40 \text{ W m}^{-1} \text{ K}^{-1}$, convection can easily provide a power larger than 0.1 TW and a
 202 magnetic Reynolds number $Rm > 100$ during the last 4 billion years (dashed lines in figure 4).

203 In contrast, with a high conductivity of $100 \text{ W m}^{-1} \text{ K}^{-1}$, the convective power and Rm match the above constraints only
 204 when the present-day core heat flow $Q_{\text{cmb}} \gtrsim 15 \text{ TW}$ (dotted lines in figure 4). This value is close to the upper estimates of the
 205 present-day Q_{cmb} ^{130,136}. A thermal conductivity larger than $100 \text{ W m}^{-1} \text{ K}^{-1}$ therefore precludes an ancient dynamo driven by
 206 thermal convection. These results prompt the study of auxiliary sources of power.

207 4.4 The exsolution boost

Exsolution of light elements can contribute to core convection and the geodynamo^{88,89,137}. When exsolution takes place near
 the CMB, exsolution releases a denser liquid with a mass anomaly flux^{31,89}

$$F_{\text{ex}} \simeq -\alpha_c M_c \frac{dC}{dT} \frac{dT_{\text{cmb}}}{dt}, \quad (7)$$

208 where $\alpha_c = -(\partial\rho/\partial C)/\rho$ is the chemical expansion coefficient for a given light element, C the concentration of this element
 209 in the liquid core measured by the total mass of the light element divided by the outer core mass M_c , and T_{cmb} is the CMB
 210 temperature. While the value of the chemical expansion coefficient α_c is close to one for both MgO and SiO₂^{89,137,138}, values
 211 of the exsolution rate dC/dT are debated. For MgO, high-pressure experiments^{88,139} find rates of $0.2 - 0.6 \times 10^{-5} \text{ K}^{-1}$ while
 212 molecular dynamics simulations⁹³ suggest rates 10 times larger. For SiO₂, experiments⁸⁹ suggest a high exsolution rate of
 213 $4 \times 10^{-5} \text{ K}^{-1}$.

The exsolution rate of light elements in the core affects the power available for the dynamo, expressed as

$$\Phi_{\text{conv}} = (F_o + F_{\text{ex}})(\psi_o - \bar{\psi}) \quad (8)$$

214 before the birth of the inner core. Using the exsolution rates from experiments^{88,89,139}, MgO-exsolution generates an added
 215 power of about 0.5 TW while SiO₂-exsolution yields an added power of about 3 TW (Supplementary Figure 2). When it is
 216 coupled with thermal convection, exsolution substantially helps driving the dynamo^{88,138}.

217 However, to solve the new core paradox, exsolution must also loosen the restrictions on the thermal conductivity. Thermal
 218 convection alone can drive a dynamo with no inner core only when the thermal conductivity $k \lesssim 100 \text{ W m}^{-1} \text{ K}^{-1}$. With the
 219 help of MgO-exsolution at a rate of $0.5 \times 10^{-5} \text{ K}^{-1}$, the conductivity range widens only moderately to $k \lesssim 130 \text{ W m}^{-1} \text{ K}^{-1}$
 220 (Supplementary Figure 3b). MgO-exsolution can drive a dynamo when the core is thermally stratified ($F_o < 0$ in (8)), but only
 221 for a narrow range of core heat flow¹³⁹ (Supplementary Figure 2 and 3a). The heat budget of the core therefore remains tight
 222 when using the low exsolution rate proposed for MgO^{139,140}.

223 In contrast, with the larger exsolution rate of $4 \times 10^{-5} \text{ K}^{-1}$ proposed for SiO₂⁸⁹, a thermal conductivity up to $k \simeq$
 224 $350 \text{ W m}^{-1} \text{ K}^{-1}$ and a core heat flow as low as $Q_{\text{cmb}} = 5 \text{ TW}$ are compatible with a long-lived geodynamo (Supplementary
 225 Figures 2 & 3ab). Such a high exsolution rate would solve the new core paradox.

226 4.5 Coupling with Earth's mantle

The heat flow Q_{cmb} escaping from the core enters the heat budget of the mantle, which reads^{141,142}:

$$Q_{\text{surf}} = Q_{\text{cmb}} + H_m - C_m \frac{dT_m}{dt}, \quad (9)$$

227 where Q_{surf} is the heat flow at Earth's surface, H_m is the internal heat produced by radiogenic elements in the mantle, C_m is the
228 heat capacity of the mantle, and T_m its average temperature. The thermal evolution of the core is therefore coupled with that of
229 the mantle^{126,127}.

230 At present, $Q_{\text{surf}} \simeq 46$ TW while geochemical models⁶⁷ imply $H_m \simeq 18$ TW. The difference must be balanced by the mantle
231 cooling and the heat Q_{cmb} escaping from the core in equation (9). For $Q_{\text{cmb}} < 10$ TW, thermal evolution models¹⁴² predict
232 a cooling rate of more than 200 K Gyr^{-1} , three times larger than estimated from petrological observations^{143,144}, implying
233 massive melting of the mantle as recently as 2 Gyrs ago. This conundrum is known as the mantle thermal catastrophe.

234 The large Q_{cmb} values needed to sustain the geodynamo at ancient times therefore help in avoiding the mantle thermal
235 catastrophe^{126,127}.

236 However, a high core heat flow also implies a fast cooling rate of the core, and hence can lead to a core thermal catastrophe
237 with a temperature exceeding 5500K below the core-mantle boundary 3 Gyrs ago²⁷. With such high core temperatures, the
238 lower mantle was molten, forming a basal magma ocean¹⁴⁵. This scenario of a hot early core and a basal magma ocean is
239 acceptable as long as the upper mantle remains solid for the last 3.5 Gyrs.

240 Some proposed that convection in the basal magma ocean could generate Earth's magnetic field in the early Earth¹⁴⁶⁻¹⁴⁸.
241 Using an electrical conductivity of $2 \times 10^4 \text{ S m}^{-1}$ ¹⁴⁷ and the same dynamo scaling as in figure 4cd, with a convective power of
242 ~ 1 TW in a 300 km thick ocean¹⁴⁸, yields a magnetic Reynolds number $Rm \approx 12$. Such a value of the magnetic Reynolds
243 number is too small to drive dynamo action (box 1). In addition, it remains uncertain whether the basal ocean convects or is
244 chemically stratified¹⁴⁹.

245 5 Sustaining a mechanical geodynamo

246 The new core paradox motivates the assessment of precession or tides as alternative driving mechanisms^{44,101,103}. In this
247 section, the orbital history of the Earth-Moon system is combined with the latest results on the flows driven by precession and
248 tides to determine whether these mechanisms could have powered the ancient geodynamo.

249 5.1 An elusive power estimate

250 Precession and tides tap into the rotational and gravitational energy reservoir stored into the spinning Earth and its orbiting
251 Moon, of which 10^{29} J are left today¹⁰¹. In contrast with the convective case, the orbital power that is converted into turbulent
252 flows and available to the dynamo is much harder to estimate.

253 Still, an upper bound can be obtained from orbital observations, which are the recession of the Moon and the increase in
254 Earth's length-of-day. The total dissipation of the Earth-Moon system Φ_{orb} varies from 1 TW to 15 TW during Earth's history,
255 with an average value of 2 to 3 TW (Supplementary Figure 4). Current models assume that this power is dissipated in ocean
256 tides through Earth's history¹⁵⁰. Yet, a fraction of this dissipation could have occurred in the Earth's core. Today, for a total
257 tidal dissipation of 2.2 TW about 0.1 TW is dissipated in the core and mantle¹⁵¹. The fraction dissipated in the core could have
258 been larger in the past when the Moon was closer to the Earth. The total dissipated power Φ_{orb} can be seen as the analogue for
259 the total heat flow out of the Earth, from which only a small fraction might contribute to the geodynamo. This upper bound is
260 larger than the minimum power of 0.1 TW needed to drive the ancient geodynamo (section 4.3). Thus, the orbital history of the
261 Earth-Moon system leaves room for an orbitally-driven dynamo in the past, but not with an ample margin.

262 Whether the flow excited by precession or tides can convert part of this power into the geomagnetic field is an unsettled
263 question. The power drawn by laminar^{34,35} and turbulent¹⁰⁵ flows due to the strong shear at the boundaries is at least two orders
264 of magnitude too low to feed the ohmic dissipation of the geodynamo. These conclusions do not change when including the
265 dissipation in the boundary layer near the inner core, or by considering a CMB topography < 10 km (details in Supplementary
266 Note 2).

267 Flow instabilities and turbulence in the bulk core could drain notably more power from Earth's orbital evolution, with a
268 theoretical upper bound for the dissipation¹⁵² of 10^9 TW. This value implies that the total energy of 10^{29} J contained in the
269 Earth-Moon would be dissipated in a few years. This upper bound is therefore too large to be useful.

270 The above arguments demonstrate that turbulence in the bulk core is a necessary condition to sustain an orbitally-driven
271 geodynamo. In the following section, we examine whether turbulent bulk flows can be triggered by precession and tides in
272 Earth's core and whether they are strong enough to produce a dynamo.

273 5.2 Turbulence in the bulk core

Bulk turbulence requires the laminar flow to be unstable, which happens when the strain rate is larger than the viscous damping
rate, that is

$$\xi u f > K \sqrt{v \Omega}, \quad (10)$$

274 where ν is the kinematic viscosity, Ω the Earth's rotation rate, $K \gtrsim 2.62$ a damping coefficient^{36,153}, ξ a numerical prefactor
275 ($\xi \lesssim 9/16$ for tides¹⁵⁴, and $\xi \lesssim 5\sqrt{15}/32$ for precession¹⁵³). For precession, $u = u_p$ is the differential velocity between the
276 outer core and the mantle and $f = f_p$ the polar ellipticity. For tides, $u = u_t$ is the speed of the tidal bulge at the CMB as it
277 rotates around the core and $f = f_i$ the diurnal tidal ellipticity.

278 While a convective geodynamo highly depends on the value of core thermal conductivity (section 4.3), condition (10) shows
279 that core viscosity is the key parameter for a dynamo driven by precession or tides. The current Earth's core is marginally
280 stable to bulk instabilities for both precession^{103,153,155} and tides¹⁵⁶. Whether turbulent bulk flows can develop is therefore
281 very sensitive to the value of core viscosity, estimates of which range from¹⁵⁷ 3×10^{-7} to 5×10^{-6} m²/s.

282 Using the most accepted scenario for the evolution of orbital parameters over time^{158,159} yields a past obliquity lower than
283 the current 23.5° value (orange curve in figure 5a). With this scenario, precession meets condition (10) only for a low viscosity
284 of $\nu \sim 10^{-7}$ m² s⁻¹ and prior to 3 Gyrs ago (figure 5a). Even in the high-obliquity scenario, a viscosity of $\nu = 10^{-6}$ m² s⁻¹ is
285 not low enough for instabilities to grow (Supplementary Figures 5). Core-filling turbulence driven by precession is therefore
286 unlikely in Earth's core.

287 Because the Earth and Moon were closer in the past, the diurnal tidal ellipticity increases when going back in time
288 (orange curve in figure 5b). With this evolution, tides can easily meet condition (10). With a viscosity of 10^{-6} m² s⁻¹ or
289 3×10^{-6} m² s⁻¹, tidal instabilities occur prior to 1.5 Gyrs or 3 Gyrs ago, respectively.

290 The growth of instabilities is not the only condition to power a dynamo. In addition, the vigour of the bulk flow, as measured
291 by the magnetic Reynolds number Rm , must be high enough (box 1). Results obtained for tides, both experimentally¹⁶⁰
292 and numerically¹⁶¹, suggest that inertial instabilities can sustain a turbulent flow of the order uf . From this scaling for the
293 vigour of bulk flows, and criterion (10) for their emergence, the evolution of orbital parameters over time^{158,159} (detailed in
294 Supplementary Note 2) allows to estimate the magnetic Reynolds number Rm for precession-driven and tidal flows (figure 5
295 and Supplementary Figure 5). Bulk flows driven by precession yield $Rm \lesssim 60$ at all times (fig. 5a), which is hardly enough for
296 a dynamo. In contrast, $Rm \gtrsim 100$ for the tidal flow in the core, reaching $Rm \simeq 700$ for the early Earth (figure 5b).

297 Thus, even in the unlikely event that precession triggers bulk instabilities, the resulting flows are not strong enough to
298 generate a magnetic field. In contrast, tides can excite strong flows filling the entire core. These could drive a dynamo, especially
299 in the early stages of Earth's history.

300 6 Implications

301 The long-term evolution of the geodynamo is inherently connected with the thermal history of the core and mantle, the evolution
302 of stratified layers in the core and the observed paleomagnetic field.

303 6.1 Thermal history of the core and mantle

304 With a high or moderate thermal conductivity, the current core heat flow Q_{cmb} must exceed ~ 10 TW to drive a convective
305 geodynamo. Such a high core heat flow is compatible with global mantle convection models¹³² and could avert the mantle
306 thermal catastrophe^{126,127} (equation (9)).

307 Only a few investigations couple the thermal evolution of the core and mantle. They solve simultaneously the energy budget
308 for both layers^{126,127,137}. When the lower mantle is about 5 times more viscous than the upper mantle, these models produce a
309 current CMB heat flow of about 13 TW, which reaches up to 40 – 80 TW in the distant past. These values allow for a long-lived
310 convective dynamo while avoiding the mantle thermal catastrophe^{126,127}. A current CMB heat flow larger than 13 TW implies
311 that the inner core is less than 700 million year old and that the lower mantle was molten before 2 ± 1 Gyr ago^{27,28}.

312 It has been hypothesized that the core is cooling too slowly to power a convective dynamo⁴⁴. This scenario relies on the
313 early mantle be fully molten. A fully-molten mantle would cool down the core in less than 100 Myrs and hence would leave no
314 heat to drive the geodynamo by convection during the subsequent billion years¹⁶². However, it is plausible that only a fraction
315 of Earth's mantle was molten during the last giant impact that formed the Earth⁹¹. With a partially-solid, hence viscous, mantle
316 above it, the core likely retained enough heat to power a convective dynamo.

317 6.2 Stratification in the early core

318 During the accretion of the Earth, giant impacts brought metal into contact with liquid silicates at very high temperature,
319 facilitating the dissolution of light elements into the core^{30,31,72,87,163}. The metal added by each giant impact was enriched in
320 light elements¹⁶⁴ and hence formed stratified layers at the top of the core¹⁶⁵. The mixing during a giant impact was too small to
321 destroy this stratification^{166,167}. The early core was therefore likely stratified in composition over hundreds of kilometers, with
322 a stratification strength, as measured by the buoyancy frequency N , 2 to 10 times the Earth's rotation rate Ω ^{165–167}.

323 Such a strong stratification would prevent the generation of magnetic fields by tides¹⁶⁸. Similarly, convection would have a
324 hard time to overcome this stratification unless the latter is localised at the top or bottom of the core.

Geodynamo simulations indicate that, at the present day, the stratified layer atop Earth's core must be thinner than about 100–300 km with a stratification not stronger than $N \sim \Omega$ to be compatible with geomagnetic observations^{169–171}. A mechanism therefore destroyed or partially mixed the primordial stratification. Yet, numerical simulations¹⁶⁷ suggest that thermal convection cannot erode more than 10 km of primordial stratification. Tidal flows or chemical convection due to the exsolution of light elements are possible mixing mechanisms that deserve further studies.

6.3 Paleomagnetic field

The driving mechanism of the geodynamo controls the strength and morphology of the paleomagnetic field. Paleomagnetic data hence provide clues about the dynamo mechanism and its evolution through time^{61,63,64,70}.

The nucleation of the inner core was a major transition for the dynamo mechanism. In the presence of a growing inner core, a light fluid is released at the inner-core boundary and drives convection (section 4.2). This driving mechanism was missing prior to the growth of an inner core (section 4.3), irrespective of whether tides or convection drove the ancient dynamo. The nucleation of the inner-core therefore coincided with a strong increase in the power available to the dynamo (section 4.3).

Theory has long predicted that this increase in power left footprints in the paleointensity record^{70,141}. However, paleointensity does not vary substantially over Earth's history⁶¹ (figure 2d). Dynamo simulations linked with thermal evolution of the core allow to investigate the signature of inner-core growth on a convective geodynamo^{97,98}. Such simulations suggest that the nucleation of the inner core caused an increase in magnetic field strength inside the core, but no resolvable change in the field intensity at Earth's surface⁹⁷. This result reconciles the absence of long-term trends in the paleointensity record with theoretical predictions. It also suggests that long-term paleointensity trends are unlikely to constrain the age of the inner core.

Still, these evolutionary dynamo simulations indicate that, during short time intervals, the magnetic field can be weak and multipolar prior to inner-core growth^{97,98}. A multipolar field could have caused short-lasting paleomagnetic anomalies. Such anomalies have been reported in the paleomagnetic record^{172,173} for ages ~ 375 Ma and ~ 580 Ma. Further paleomagnetic investigations are needed to know whether these anomalies could be a signature of the absence of an inner core.

A geodynamo driven by convection is therefore compatible with paleomagnetic data. However, only a few dynamo simulations with no inner core have been published^{84,96,97}. More simulations, especially in a turbulent regime, are needed to better understand the ancient geodynamo and its paleomagnetic signature.

7 Summary and future directions

Unlike mechanical dynamos, a geodynamo driven by thermo-chemical convection accounts for most of the properties of the geomagnetic field (table in box 2). Convection also produces enough power to generate a magnetic field over the last 3.4 Gyrs, as inferred from paleomagnetic data. Convection therefore remains the most likely driving mechanism for the geodynamo.

Yet, convective dynamo models almost all assume that composition and temperature have the same diffusivity, while light elements released at the inner-core boundary diffuse several orders of magnitude slower than temperature. The effect of double diffusion, when temperature and composition diffuse at different rates, on the geodynamo has been little studied^{174–178}. Further examination of double diffusion could open new avenues for exploring the dynamics of Earth's core.

Whether the exsolution of light elements solves the new core paradox strongly depends on the value of the exsolution rate, which is debated^{88,89,93,139}. More precise estimates of the exsolution rate from high-pressure experiments and calculations are therefore needed. In addition, whether MgO^{88,139}, or SiO₂⁸⁹, or other components^{94,138} exsolve, and when their exsolution started, is an unsettled issue. It remains unknown whether the early core contained sufficient amounts of magnesium for the exsolution mechanism to start prior to 3 Gyrs ago and power the early geodynamo^{94,138,179}. Similarly, SiO₂ might not start exsolving at temperatures larger than 4000 K¹⁷⁹, which are expected in the ancient core. Oxides are extracted from the core at a rate set by mantle convection¹³⁸. This interaction with the mantle notably affects the nature of the exsolved species and the time at which exsolution starts¹³⁸, and therefore deserves further investigation.

In the Earth's core, precession requires an unlikely low viscosity to trigger bulk turbulence together with an unlikely high obliquity (figure 5a and Supplementary Figure 5). Even under such favourable circumstances, the predicted flow velocity is hardly sufficient for magnetic induction to overcome ohmic dissipation (figure 5a). In addition, the magnetic field obtained in precession-driven dynamos does not match the properties of the modern geomagnetic field (figure 3d and table in box 2).

Unlike precession, tides can generate vigorous bulk flows in the early Earth. However, the flows and magnetic fields driven by tides are poorly known. Numerical simulations of tide-driven dynamos are in their infancy in deformed spheres, with a single proof-of-concept kinematic dynamo¹⁰⁶. A dynamo driven by tides in the past would require a substantial fraction of the power currently attributed to the oceans¹⁵⁰ to be dissipated in the core.

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722 **Author contributions**

723 All authors contributed equally to the thinking and writing of the article. ML contributed particularly to the convection energy
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731 **Competing interests**

732 The authors declare no competing interests.

733 **Code Availability Statement**

734 The authors declare the code for Figures 4 and 5 is available from [https://doi.org/10.6084/m9.figshare.
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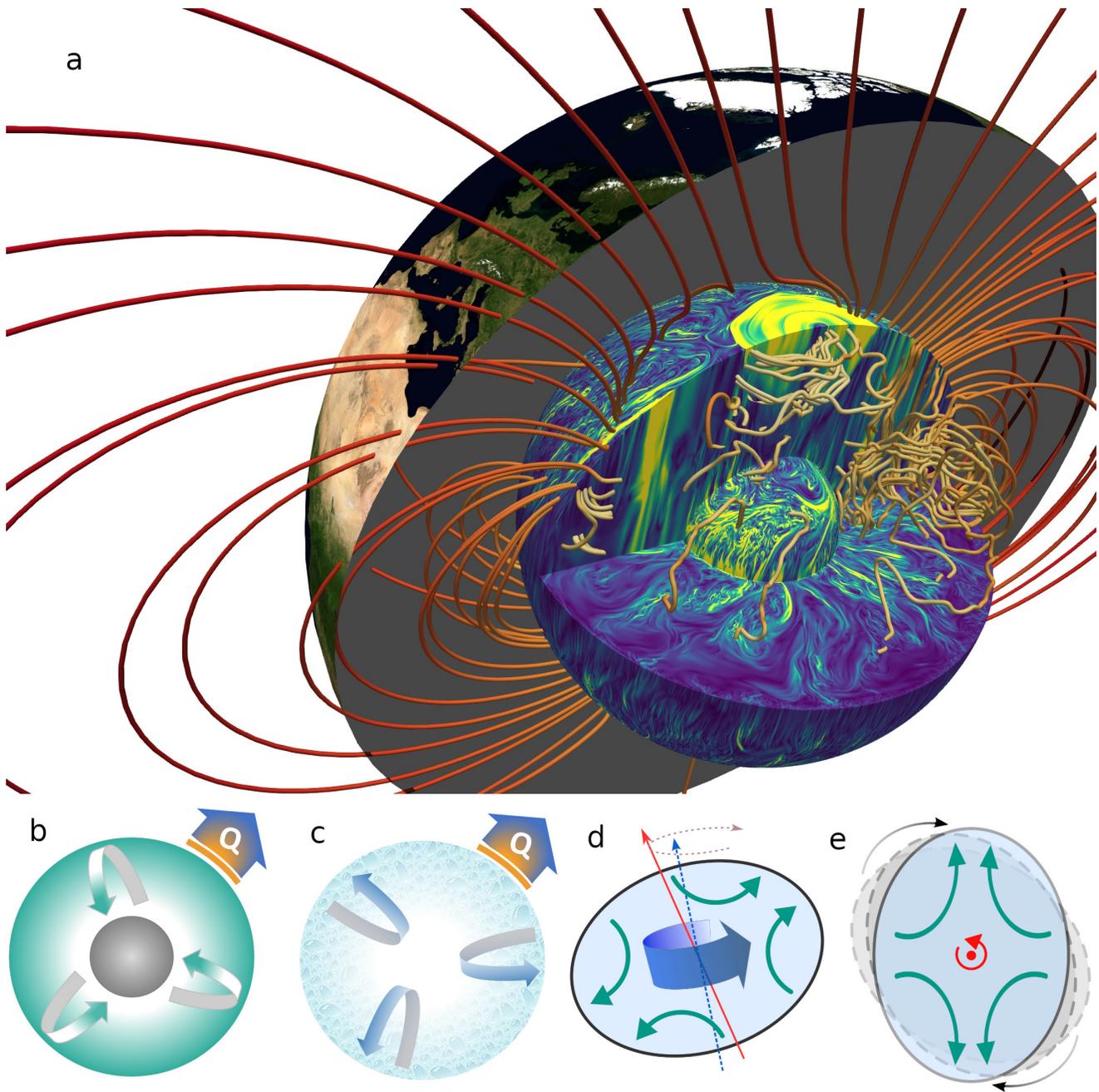


Figure 1. The Earth's dynamo requires turbulent motion of liquid iron in Earth's core. a) The three-layer structure of Earth's interior. The rocky mantle overlays a metallic core. The core, which is essentially made of iron, is divided into a liquid outer core and a solid inner core. The flow of liquid metal sustains dynamo action. The magnetic field lines (red to yellow lines) and the velocity field strength (blue to yellow) are taken from a direct numerical simulation of the dynamo driven by turbulent convection¹¹.

b-e) Candidate mechanisms to drive the geodynamo in the outer core. b) Convection driven by core cooling and inner-core growth. The heat Q_{cmb} leaving the core leads to the solidification of the inner-core, releasing light elements at the base of the outer core. At the outer edge of the core, cooling releases a cool, denser fluid that sinks into the deeper core. c) Convection driven by the exsolution of light oxides such as MgO or SiO₂. The iron-rich liquid released at the top of the core sinks into the core. d) Precession makes the fluid rotate along an axis (blue) that is different from the mantle rotation axis (red); a secondary circulation (teal arrows) is induced by the non-spherical shape. e) Tides induce a deformation that rotates around the liquid core in about 1 day, inducing a recirculation.

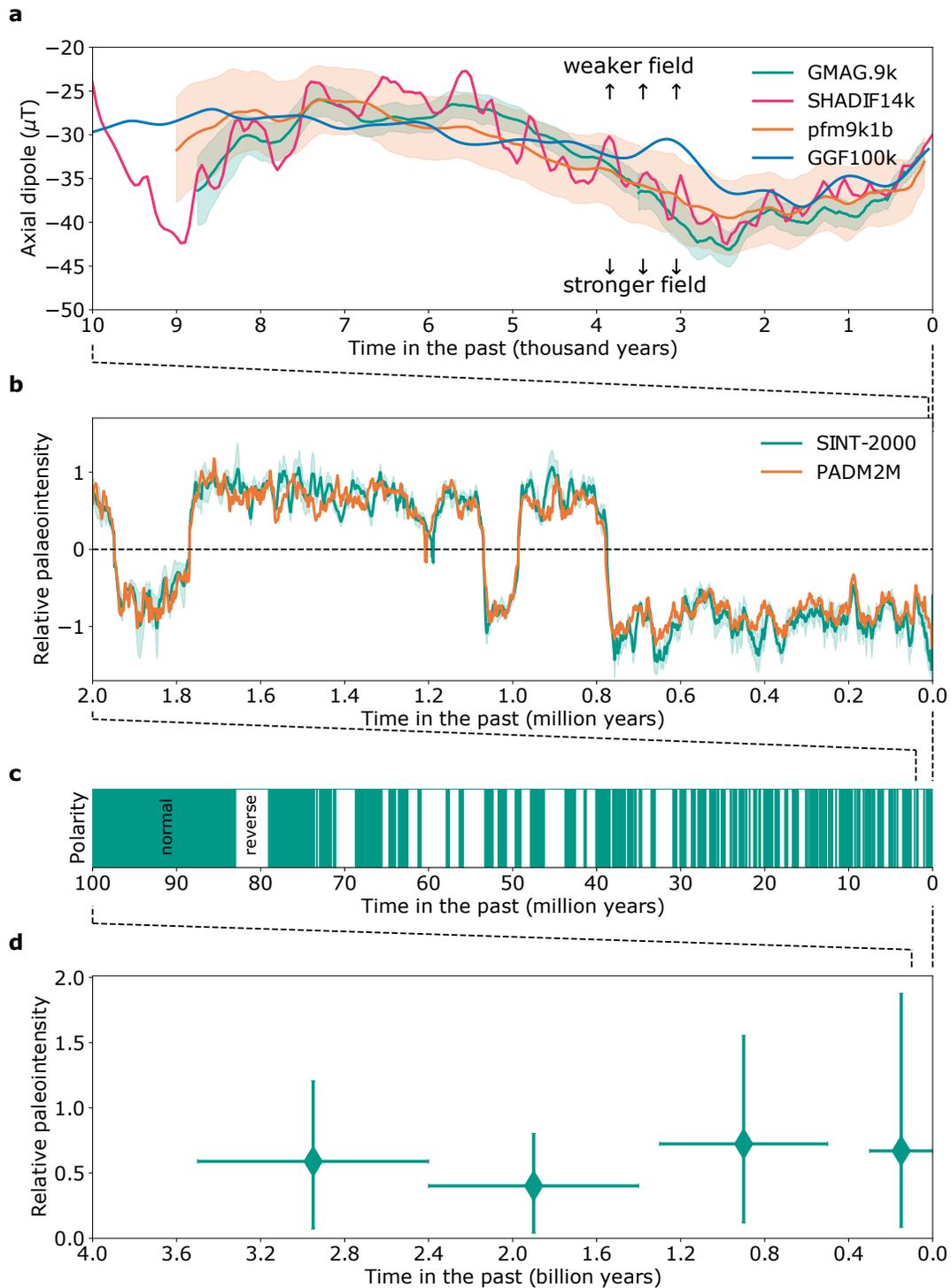


Figure 2. Earth's dynamo operates on a broad range of time scales. a) Evolution of the geomagnetic axial dipole (the g_1^0 Gauss coefficient) according to four reconstructions: SHADIF14k¹⁸⁰, GGF100k¹⁸¹, GMAG.9k¹⁸², and pfm9k1b¹⁸³, with the standard deviation uncertainty for the latter two. The axial dipole coefficient is given in μT . b) Fluctuation of the signed relative paleointensity of the magnetic field according to the SINT-2000 model¹⁸⁴, with its uncertainty, and the PADM2M model¹⁸⁵. To normalise, the average virtual axial dipole moment $m_{\text{ref}} = 7.46 \times 10^{22} \text{ A m}^2$ since the last reversal is used. c) Geomagnetic polarity¹⁸⁶, showing the variability of reversal frequency. Intervals of normal (present-day) polarity are in teal. d) Virtual dipole moment from the PINT database⁶¹, normalised by $m_{\text{ref}} = 7.46 \times 10^{22} \text{ A m}^2$. The coordinates of a symbol are the median value of the age interval and the median value of the dipole moment of the dataset selected for the time interval of interest. The horizontal bar spans the corresponding time interval, and the vertical bar covers the values found in the dataset. Historical geomagnetism and paleomagnetism provide key constraints on the operation of the geodynamo.

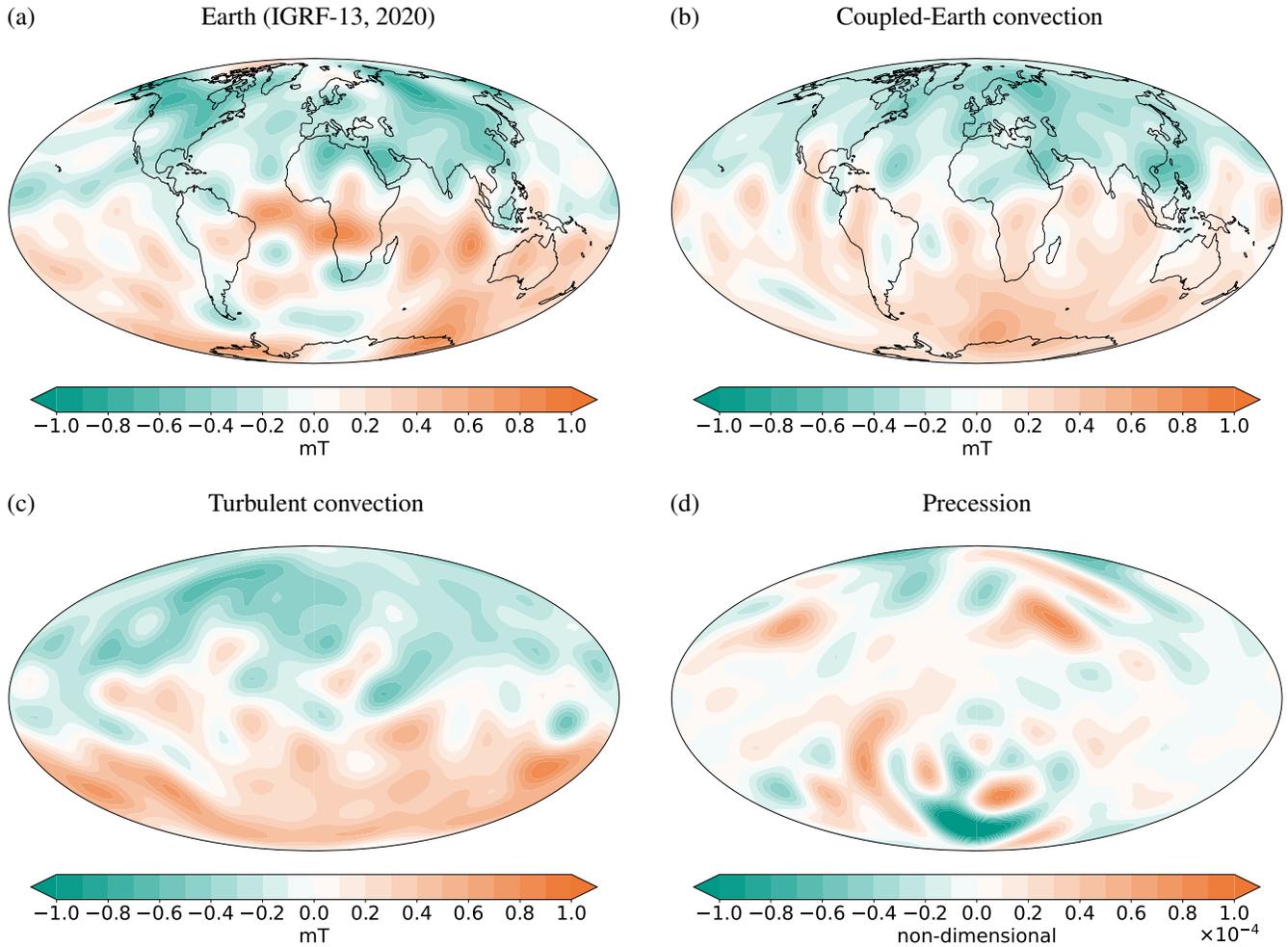


Figure 3. The morphology of the Earth’s magnetic field is best reproduced by convection-driven dynamos. Snapshots of the radial component of the magnetic field at the core surface (Mollweide projection). Only the largest scales (spherical harmonic degree $\ell < 13$) that are resolved for the Earth’s core are shown. a) Earth’s magnetic field in 2020, according to the International Geomagnetic Reference Field model⁴⁶. b) Coupled Earth direct dynamo model^{12,187} ($Ek = 3 \times 10^{-5}$, $Rm = 930$, $Pm = 2.5$, see Box 2 for definitions), rescaled using dynamo scaling laws¹⁸⁷. c) Direct numerical simulation¹¹, ($Ek = 10^{-7}$, $Rm = 514$, $Pm = 0.1$), rescaled so that the average axial dipole moment matches m_{ref} ¹⁸⁴ (see Figure 2). d) Low-viscosity turbulent precession dynamo¹⁰⁵, aligned on the fluid rotation axis ($Ek = 10^{-5}$, $Rm \simeq 1900$, $Pm = 0.3$). Not all candidate driving mechanisms yield a magnetic field that complies with the observed field.

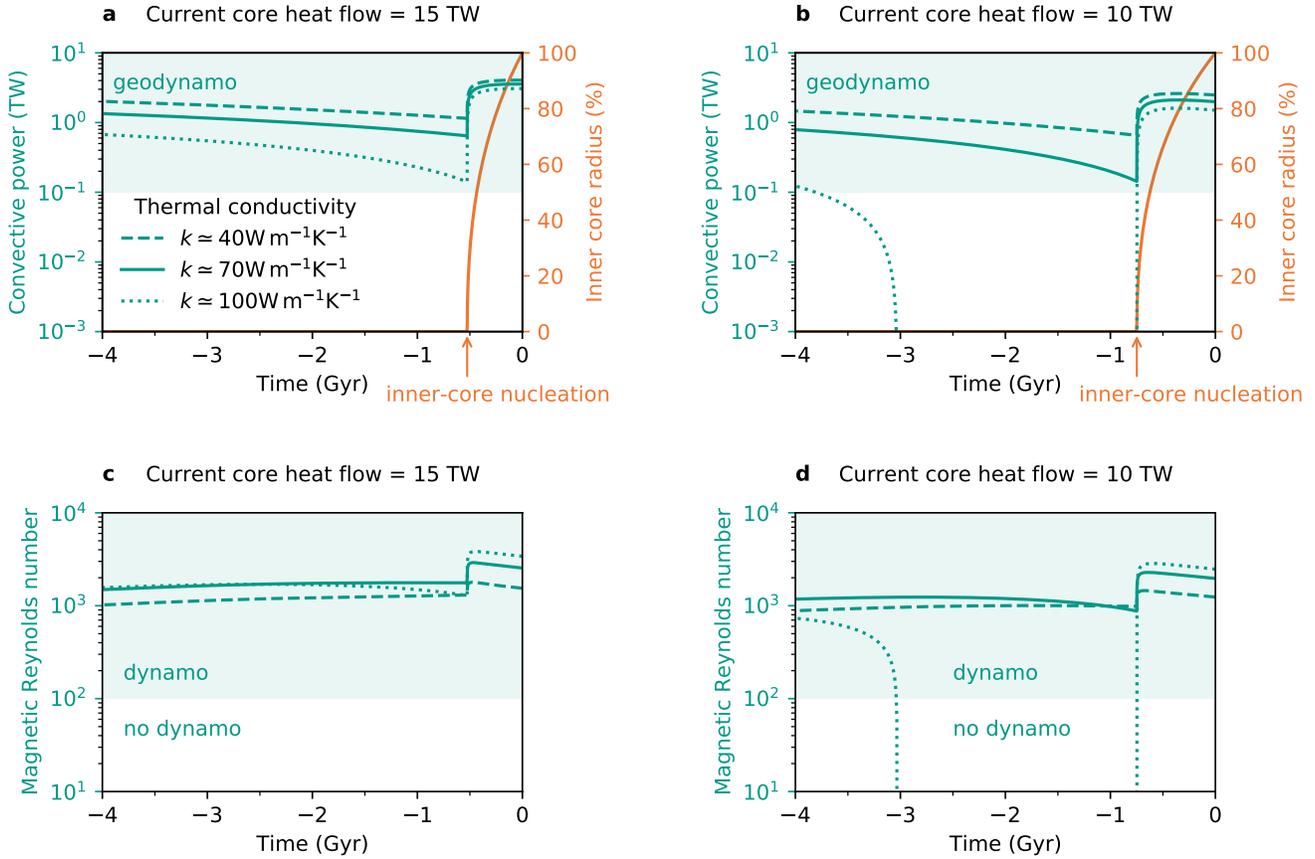


Figure 4. Convection can power the geodynamo in the distant past when thermal conductivity is lower than 100

$\text{W m}^{-1} \text{K}^{-1}$. a) Convective power (teal) and inner-core radius (orange, % of its present-day value) generated by convection as a function of time. The shaded teal band delineates the range of admissible convective power for the geodynamo. A reduced model¹²¹ was computed using three different thermal conductivities $k = 40$ (dashed), 70 (plain) and $100 \text{ W m}^{-1} \text{K}^{-1}$ (dotted) corresponding to present-day conducted heat flows²⁷ $Q_{\text{is}} = 6 \text{ TW}$, 10 TW and 15 TW, respectively. The assumed present-day core heat flow is $Q_{\text{cmb}} = 15 \text{ TW}$. b) Same as a) with $Q_{\text{cmb}} = 10 \text{ TW}$. c) Magnetic Reynolds number Rm (see Box 1 and Box 2) generated by convection as a function of time for the same models as in a) ($Q_{\text{cmb}} = 15 \text{ TW}$). Dynamo action is expected only for $Rm \geq 100$ (shaded teal band). d) Same as c) with $Q_{\text{cmb}} = 10 \text{ TW}$. The core heat flow Q_{cmb} is arbitrarily assumed to increase at 2.9 TW Gyr^{-1} , which is close to the lowest estimates from models that couple the thermal evolution of the core and mantle¹²⁷. To compute the past evolution of the conducted heat flow $Q_{\text{is}}(t)$ equations (1-2) and the evolution of core temperature from the energy budget of the core^{120, 121} are used. $Rm = U D / \eta$ is deduced from dynamo scaling law^{12, 135}, where $U \simeq 2 \Omega D p^{4/9}$ is the convective velocity, $p = \Phi_{\text{conv}} / (\rho \Omega^3 D^2 \mathcal{V})$ the dimensionless convective power⁸⁴, Ω the time-varying Earth's rotation rate¹⁵⁸, \mathcal{V} the outer core volume and D the evolving thickness of the outer core. The electric conductivity is assumed to depend on the thermal conductivity through the Wiedemann–Franz law. Convection provides enough power and kinetic energy to power the geodynamo over geological times.

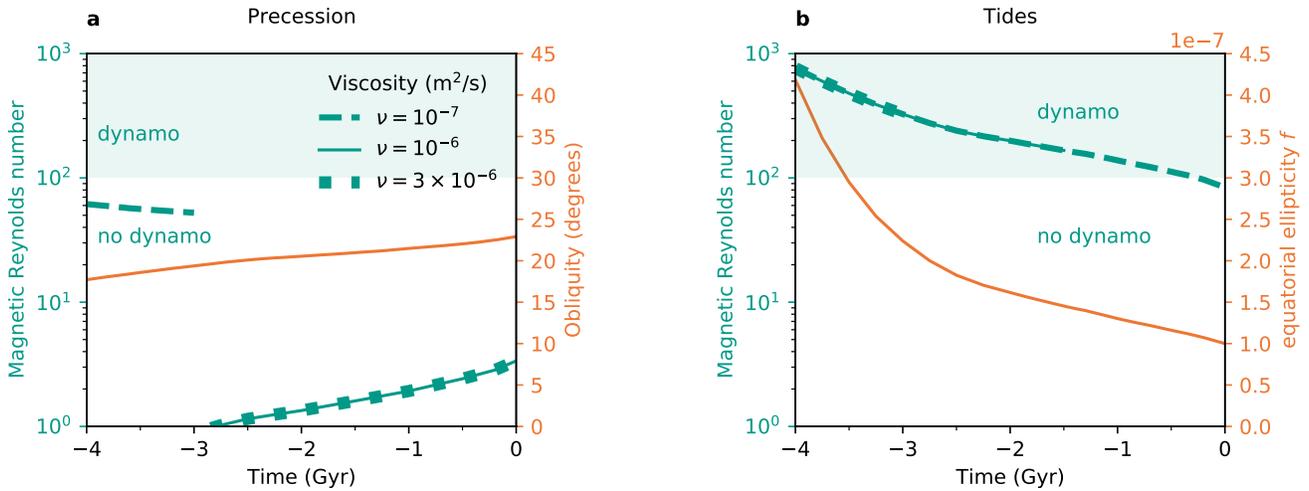


Figure 5. Precession cannot drive a dynamo but tides generated strong flows that could drive a dynamo in the early Earth. a) Magnetic Reynolds number Rm (teal lines, see Box 1) generated by precession as a function of time for three values¹⁵⁷ of the kinematic viscosity in the outer core. The Rm jump at -3 Ga for $\nu = 10^{-7}$ m²/s is due to the triggering of bulk instabilities (instability onset is reached at this point). Dynamo action is expected only for $Rm \geq 100$ (shaded teal band). On the right vertical axis, the precession angle (obliquity) used for the calculation is shown. b) Same as a) for flows produced by tides but showing the equatorial ellipticity f used for the calculation on the right vertical axis. The estimates of Rm in b) are all associated with bulk instabilities, and the curves start at their respective onset for each viscosity. Criterion (10), with $K = 2.62$, is used to predict whether bulk instabilities are triggered. $Rm = U r_c / \eta$, where r_c is the core radius, is estimated using the following assumptions for the velocity, U . For times when bulk instabilities are excited, $U = f u$ is assumed, where u is respectively the differential velocity between the outer core and the mantle for precession and the speed of the tidal bulge at the core-mantle boundary for tides, and f is respectively the polar or tidal ellipticity^{160, 161, 188}. For precession a), for times -3 Gyrs to 0 when bulk instabilities are not excited, the flow velocity $U = 0.15 \Omega r_c Ro^{3/2}$ is used for precession viscosity-driven instabilities¹⁰⁵, where Ro is the Rossby number $Ro = u / (\Omega r_c)$. Although precession could not have driven a dynamo in the Earth's history, tides could have driven a dynamo prior to 1.5 Gyrs ago.

744 **Box 1: Basics of dynamo action**

A dynamo converts mechanical energy into electromagnetic energy and produces a self-excited magnetic field^{8, 189, 190}. The induction equation governs the evolution of the magnetic field \mathbf{B} given the velocity field \mathbf{u} of an electrically conducting but neutral fluid:

$$\partial_t \mathbf{B} = \nabla \times (\mathbf{u} \times \mathbf{B}) + \frac{1}{\mu \sigma} \nabla^2 \mathbf{B}, \quad (11)$$

with μ the magnetic permeability and σ the electrical conductivity. The above equation shows that the rate of change of the magnetic field results from the induction term (first term on the right-hand side), which involves the fluid motion and is responsible for the production of magnetic field, and the magnetic diffusion (second term on the right-hand side) related to ohmic dissipation. When induction is much larger than magnetic diffusion, a runaway growth of the magnetic field is possible. Comparing the order of magnitude of these two terms yields the dimensionless magnetic Reynolds number

$$Rm = \mu \sigma LU = \frac{UL}{\eta}, \quad (12)$$

745 with L the system length scale, U the typical fluid velocity, and $\eta = 1/\mu\sigma$ the magnetic diffusivity. A necessary condition
 746 to self-generate a magnetic field is $Rm \gg 1$. With U the maximum fluid velocity, the search for an optimal dynamo suggests
 747 $Rm \gtrsim 44$ in the sphere¹⁹¹. Another condition is that the velocity field is sufficiently complex and its topology adequate. For
 748 example, a solid-body rotation or strictly two-dimensional flows are incapable of dynamo action¹⁹². In contrast, helical flows
 749 are prone to dynamo action^{9, 191}.

750 Such a dynamo process is different from the magnetic field produced by the motion of charged matter for which the field
 751 is simply proportional to the velocity and charge. In the case of a dynamo process, motions of a neutral fluid can lead to
 752 spontaneous generation of electric currents and magnetic field.

753 **Box 2: Hard-to-reach dynamos**

754 Multiple time scales control the Earth dynamo. The ordering of these time scales controls the regime in which the dynamo
 755 operates. When evaluated at the largest length scale, typical time scales range from a day for the rotational time to 10^5 years for
 756 magnetic diffusion.

757 The most relevant time scales are the rotation time $t_\Omega = 1$ day, the time $t_B = D\sqrt{\rho\mu}/B \sim 3$ yr of the hydromagnetic waves
 758 called Alfvén waves, the fluid overturn time $t_U \sim 300$ yr, the magnetic diffusion time $t_\eta = D^2/\eta \sim 10^5$ yrs and the viscous
 759 diffusion time $t_\nu = D^2/\nu \sim 10^{11}$ yr, where D is the outer core thickness, ρ the outer core density and ν the kinematic viscosity.
 760 Numerical simulations cannot cover such a range, but relevant regimes are reached when the ordering of these time-scales is
 761 preserved.

762 Dimensionless numbers evaluate the ratios between these times. These numbers are important to isolate the dominant
 763 physical processes. They are also necessary to properly compare numerical simulations, analogue experiments and Earth's
 764 dynamo (Supplementary Figure 1).

765 The magnetic Reynolds number introduced in Box 1 $Rm = t_\eta/t_U$ measures the importance of magnetic field production to
 766 magnetic diffusion.

767 The Ekman number $Ek = t_\Omega/t_\nu$ expresses the relative importance of viscous stress to the Coriolis force.

768 The Rossby number $Ro = t_\Omega/t_U$ quantifies the relative magnitude of inertial forces against the Coriolis force.

769 The magnetic Prandtl number $Pm = t_\eta/t_\nu$, is the ratio of kinematic viscosity ν to magnetic diffusivity η .

770 The Reynolds number $Re = t_\nu/t_U = Rm/Pm = Ro/Ek$ measures the degree of turbulence, that is the importance of inertial
 771 to viscous forces.

772 Estimates of these numbers for Earth's core are $Rm \sim 10^3$, $Ek \sim 10^{-15}$, $Ro \sim 10^{-6}$, $Pm \sim 10^{-6}$ and $Re \sim 10^9$. The values
 773 of Ro and Ek reflect the importance of planetary rotation on the dynamics. That of Pm is thought to be responsible for the scale
 774 separation between the large-scale magnetic field and the small-scale, turbulent velocity field.

775 The best way to assess whether a simulation is in the Earth's dynamo regime is to check that it does reproduce observed
 776 features of the Earth's magnetic field. The following table lists the current ability (marked with a ✓), or inability, (marked
 777 with a ✗), of convection-driven simulations, precession-driven simulations, and analogue experiments to account for selected
 778 properties of Earth's dynamo.

Feature	Convecting simulations ^a		Precessing simulations ^a		dynamo experiments (turbulent)
	laminar/turbulent	laminar/turbulent	laminar/turbulent	laminar/turbulent	
dipole-dominated field ^b	✓/✓	✓/✓	✓/✗	✓/✗	✓
surface morphology ^c	✓/✓	✓/✓	✗/✗	✗/✗	?
fast dynamics ^d	✗/✓	✗/✓	?	?	?
slow dynamics ^d	✓/✓	✓/✓	?	?	?
reversals ^e	✓/?	✓/?	✗/✗	✗/✗	✓
strong-field ^f	✓/✓	✓/✓	✓/✗	✓/✗	✗
scale separation ^g	✗/✓	✗/✓	✗/✗	✗/✗	✓

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^a The simulations are separated into two classes, laminar and turbulent, which correspond to a ratio of inertial to viscous forces smaller or larger than $\sim 10^3$, respectively. ^b Dipole-dominated refers to a dynamo field whose main component is dipolar. ^c Surface morphology refers to the large-scale morphology of the geomagnetic field at the core surface, which is well-known down to a scale of 1500 km. ^d Fast dynamics refers to the presence of waves operating on the magnetic time scale t_B , which is substantially lower than the advective timescale t_U characterizing the slow dynamics of core flow. ^e Reversals refers to the capacity for the field to reverse its polarity in an irregular fashion over geological time scales. ^f Strong-field refers to a dynamo operating with a magnetic energy larger by several orders of magnitude than the kinetic energy. ^g Scale separation implies that the flow energy spectrum peaks at a scale much smaller than the dominant magnetic scale. For further details the reader can consult previous studies^{15,105,193}.