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Evolving core conditions ca. 2 billion years ago detected by paleosecular variation

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ABSTRACT

Paleomagnetic data provide one of the few probes available to interrogate early evolution of the core. Here we apply this probe by examining the latitudinal dependence of paleosecular variation (PSV) data derived from high-quality paleomagnetic data collected from Proterozoic and Neoarchean rocks. These data define a Neoarchean geomagnetic field that was more dipolar than that during Proterozoic times, indicating a change in core conditions. The signals observed may reflect a change in forcing of the dynamo and an early onset of inner core growth. We propose a model that links evolution of the core, mantle and crust in three principal phases: (i) Before approximately 3.5 Ga, an entirely liquid core may not have hosted a geodynamo. If heat transport was sufficient across the core–mantle boundary, however, a geodynamo could have been generated. If so, sources in the shallow outer core could have been more important for generating the dynamo relative to deeper convection, resulting in a field that was less dipolar than that generated in later times. (ii) Cooling of the lower mantle between ca. 2 and 3.5 billion years ago was promoted by deep subduction and possibly coincided with inner core growth. The geodynamo during this episode was deeply-seated producing a highly dipolar surface magnetic field. (iii) After ca. 2 billion years ago, continued subduction led to large-scale core–mantle boundary compositional and heat flux heterogeneity. With these changes, shallow core contributions to the geomagnetic field grew in importance, resulting in a less dipolar field.

1. Introduction

There is currently much debate over the nature of Earth's early core. For example, estimates for the onset of solid inner core nucleation range from times younger than 1 Ga (Aubert et al., 2009) to 3.5 Ga (Gubbins et al., 2004). Paleointensity data indicate the presence of a geodynamo in Mesoarchean and Paleoarchean times (3.2–3.45 Ga) (Tarduno et al., 2007, 2010). But some models suggest that strong fields can be generated by an early dynamo without inner core growth (Sakuraba and Kono, 1999). We can gain insight into early core conditions by examining the morphology of the ancient geomagnetic field defined by paleomagnetic data. Specifically, we can track the importance of non-dipole fields in the past using the angular dispersion (S) of virtual geomagnetic poles (VGPs) derived from paleomagnetic data:

$$S = \sqrt{\frac{1}{N-1}\sum_{i=1}^{N} \Delta_i^2} \tag{1}$$

where N is the number of VGPs and Δ_i is the angle between the ith VGP and the mean paleomagnetic pole. McFadden et al. (1991) modeled S as independent dipole (S_D, antisymmetric) and quadrupole (S_Q, symmetric) families, with the latter dominating at the equator:

$$S(\lambda) = \sqrt{S_D^2 + S_Q^2} = \sqrt{(a\lambda)^2 + b^2} \tag{2}$$

where λ is paleolatitude, and a and b are constants. Complete independence of the two families is unlikely, but this interpretation (Model G) remains a useful framework to gauge past paleosecular variation (PSV).

Although lava flow sequences have yielded high resolution PSV values for the last 5 million years (e.g., Johnson et al., 2008), data on billion-year time scales are more difficult to obtain. A few extant lava flow sequences are available, but these must be supplemented with data from dike swarms. Any given regional data set may fortuitously overestimate or underestimate PSV. But if data sets are available spanning many latitudinal belts from multiple ancient cratons, a synoptic view of PSV can be derived. Smirnov and Tarduno (2004) found that such data suggest that the field at the time of the Proterozoic/Archean boundary (~2.5 Ga) was more dipolar than the field of the last 5 million years. This result was

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confirmed by an analysis of the same time window by Biggin et al. (2008a). Here, we expand our initial analysis, to assess any PSV pattern that might reflect changes in the Precambrian core conditions.

2. Application and results

We have identified two Precambrian time windows where global igneous units allow a new assessment (Table 1; Fig. 1A). We used the Global Paleomagnetic Database GPMDB4-6 (Pisarevsky, 2005) (www.tsrc.uwa.edu.au/data_bases), supplemented with recent results for our new data set. We exclude data from sedimentary, metamorphic, plutonic and silicic extrusive rocks. In particular, the silicic lavas often do not form easily distinguishable lava flows and may be deposited on slopes, which makes it difficult to assess the number of independent cooling units and structural corrections. Therefore, our analysis was confined to mafic and intermediate extrusive rocks and shallow mafic intrusions that can record distinct field directions. We further apply the following criteria: (1) Directions must be from ⩾10 sites each comprising ⩾3 samples. (2) Data must be from modern demagnetization and processing techniques (e.g., principal component analysis). (3) A primary origin of the magnetization must be convincingly demonstrated. (4) Data must be consistent with a thermoremanent magnetization, without evidence of chemical remanence. (5) Magnetization age must be reliably constrained. No site selection criteria based on the precision parameter (k) or maximum 95% confidence area (α_{95}) were applied. However, for most (575 of 585) of the accepted sites, α_{95} did not exceed 20°.

S values were corrected for within-site dispersion following Doell (1970). In three studies, published information is insufficient to correct S, but the large number of samples and site-level statistics lead us to believe that any additional uncertainty is less than a few degrees. S confidence intervals (1σ) were calculated using a $N-1$ jackknife method (Efron, 1982).

As opposed to the small, select Matachewan dike data set used by Smirnov and Tarduno (2004), we use a new compilation (Evans and Halls, 2010); directions from the western subprovince of the Superior craton were rotated using an Euler pole at 51°N, 85°W and rotation angle of 14°CW. We exclude results from the Derdepoort basalts (Wingate, 1998) used by Biggin et al. (2008a) because

Table 1
Summary of paleomagnetic studies used for estimating the paleosecular variation.

Unit	Age (Ma)	B	Plat	$S \pm dS$	C/UC	S_{exp}	$S_{obs}-S_{exp}$	Sign
Bangemall Basin Sills[a]	1070	11	27.8	13.9 ± 2.6	C	11.6	2.3	+
Lake Shore traps[b]	1087	30	14.6	13.6 ± 0.8	C	10.6	3.0	+
Portage Lake Volcanics[c]	1095	28	16.9	14.5 ± 1.2	UC	10.7	3.8	+
North Shore Traps[d]	1098	34	27.3	11.4 ± 2.1	C	11.5	−0.1	−
Mamainse Point Upper N[e]	∼1100	21	22.4	4.8 ± 3.0	C	11.1	−6.3	−
Umkondo dolerites[f]	1110	15	6.8	14.2 ± 3.1	C	10.3	3.9	+
Cleaver dikes[g]	1740	17	39.1	14.4 ± 4.0	C	12.8	1.6	+
Taihang dikes[h]	1769	19	2.6	8.5 ± 2.6	C	10.2	−1.7	−
Flaherty volcanics[i]	1870	11	26.2	11.6 ± 1.7	UC	11.4	0.2	+
Mashonaland dolerites[j]	1880	16	28.8	14.8 ± 4.0	C	11.7	3.1	+
Fort Frances dikes[k,l]	2067–2077	12	35.1	11.2 ± 3.3	C	12.3	−1.1	−
Marathon dikes (R)[k,l]	2101–2106	13	37.0	14.0 ± 2.5	C	12.6	1.4	+
Marathon dikes (N)[k,l]	2121–2126	16	39.9	16.0 ± 2.6	C	12.9	3.1	+
Biscotasing dikes[m]	2169	12	38.4	12.6 ± 2.3	C	12.7	−0.1	−
Ongeluk lavas[n]	2200	32	14.1	7.1 ± 7.5	C	10.5	−3.4	−
Dharwar dikes[o]	2367	25	69.1	15.5 ± 1.1	C	17.1	−1.6	−
Matachewan dikes (N)[p]	2473–2446	28	14.8	6.4 ± 2.9	C	10.6	−4.2	−
Matachewan dikes (R)[p]	2473–2446	101	7.7	8.8 ± 5.1	C	10.3	−1.5	−
Karelia dikes[q]	2440	11	30.0	10.7 ± 1.3	UC	11.8	−1.1	−
Allanridge lavas[r,s]	2664–2709	17	43.2	12.1 ± 5.7	C	13.3	−1.2	−
Upper Fortescue lavas[t]	2715	16	35.0	12.9 ± 3.2	C	12.3	0.6	+
Fortescue Lower lavas[t]	2746	75	49.5	12.8 ± 2.0	C	14.2	−1.4	−
Fortescue Package 0[t]	>2772	24	58.3	16.1 ± 2.8	C	15.4	0.7	+

B: number of units; Plat: paleolatitude; S, dS: angular dispersion of VGPs and confidence interval; C/UC: data corrected (uncorrected) for within-site dispersion; $S_{obs}-S_{exp}$: difference between observed S and that predicted (S_{exp}) from Model G fit to all data (a = 0.20 ± 0.04, b = 10.17 ± 0.90); Sign: result of Sign Test (see text). Paleomagnetic and age data sources: Global Paleomagnetic Database (GPD) (Pisarevsky, 2005) data identifier is listed.
See references cited for more recent works:
[a] Wingate et al. (2002) (GPD 3455).
[b] Diehl and Haig (1994) (GPD 2776).
[c] Hnat et al. (2006).
[d] Tauxe and Kodama (2009).
[e] Swanson-Hysell et al. (2009).
[f] Gose et al. (2006).
[g] Irving et al. (2004) (GPD 3609).
[h] Halls et al. (2000) (GPD 3394).
[i] Schmidt (1980) (GPD 1862).
[j] Bates and Jones (1996) (GPD 3088).
[k] Buchan et al. (1996) (GPD 3061).
[l] Halls et al. (2008).
[m] Halls and Davis (2004) (GPD 3644).
[n] Evans et al. (1997) (GPD 3175).
[o] Halls et al. (2007).
[p] Evans and Halls (2010).
[q] Mertanen et al. (1999) (GPD 3296).
[r] de Kock et al. (2009).
[s] Strik et al. (2007).
[t] Biggin et al. (2008a).

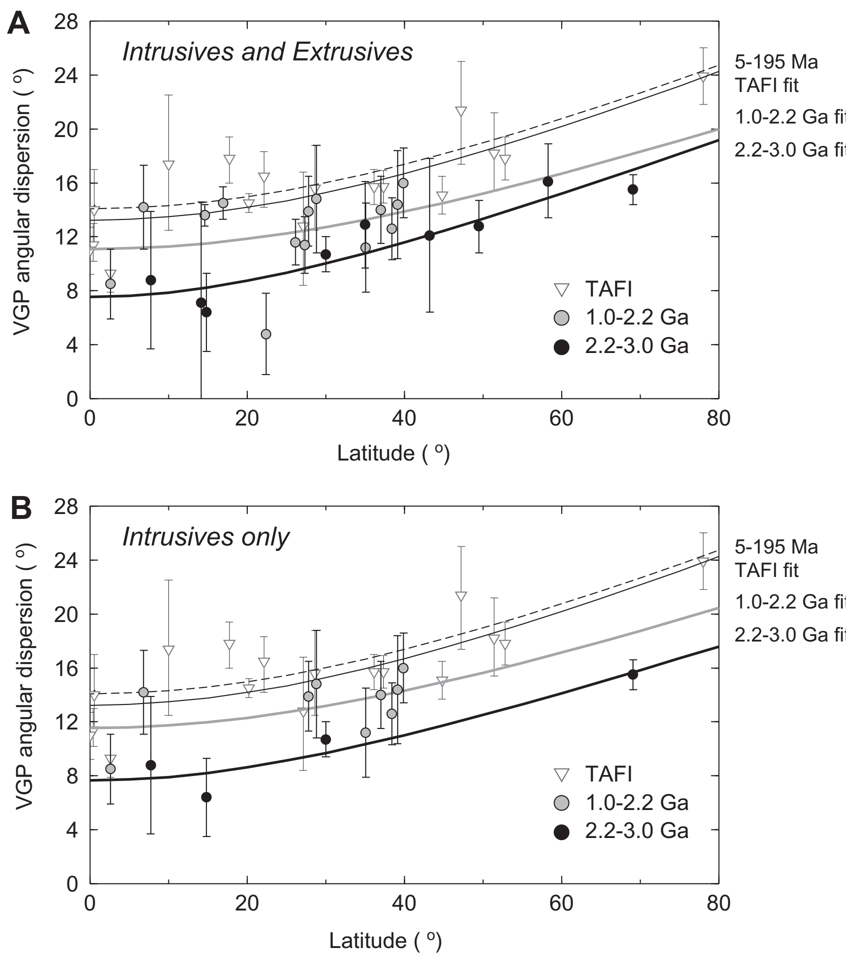

Fig. 1. (A) Latitudinal dependence of angular dispersion S of virtual geomagnetic poles for the Precambrian (solid symbols) intrusive and extrusive units and extrusives of the last 5 million years (open inverted triangles). See Table 1 for the Precambrian data sources. Gray and black symbols: younger and older than 2.2 Ga, respectively. Solid black, gray, and thick black lines: Model G fits for the 0–5 Ma (Time-Averaged Field Initiative, TAFI; Johnson et al., 2008; Lawrence et al., 2009; Kent et al., 2010; Opdyke et al., 2010), 1.0–2.2 Ga and 2.2–3.0 Ga data, respectively. The dashed line shows the Model G fit for the 5–195 Ma data (McFadden et al., 1991; Tarduno et al., 2002; note individual data points are not shown here) (see text). (B) Latitudinal dependence of S only for the Precambrian (solid symbols) intrusive units (see text).

the large apparent S value (24.3°) probably reflects uncertainties in bedding corrections for the lavas which are found in faulted basins (Wingate, pers. comm., 2009). This tectonic uncertainty can masquerade as PSV. We exclude magnetizations carried by hematite as these may record chemical remanences acquired after cooling.

In all selected studies, the maximum deviation of VGPs from the mean paleopole did not exceed 35°. Therefore the application of a constant cutoff angle of 45° commonly used to exclude transitional VGPs (e.g., Johnson et al., 2008) did not modify any of the datasets. Because of the relatively small scatter of VGPs in our Precambrian datasets (Supplementary Table 1), we feel that the application of a variable cutoff (Vandamme, 1994) may remove some scatter related to the normal secular variation rather than to the transitional field. Therefore, we chose not to apply the variable cutoff in this study. However, when used, the variable cutoff affected only three datasets (Supplementary Table 1) and did not change the overall conclusions of this study (Supplementary Fig. 1).

We split the Precambrian dataset into two age groups at ~2.2 Ga (specifically, 1.0–2.2 Ga and 2.2–3.0 Ga) and fit the data in each group using Model G (Eq. (2)). The fitting was done using the Levenberg–Marquardt least-square iterative algorithm (e.g., Björck, 1996). We find Model G parameters $a = 0.21 \pm 0.09$ (1σ), $b = 11.10 \pm 1.46$ and $a = 0.22 \pm 0.02$, $b = 7.56 \pm 0.84$ for the <2.2 Ga and >2.2 Ga groups, respectively. While the statistically indistinguishable values of the parameter a indicate similar shapes of the fitting curves, their equatorial intercepts (defined by the parameter b) are different at the 95% confidence level.

Any individual Proterozoic or Neoarchean S value may overestimate or underestimate PSV because of under-sampling, and/or there may be trends on ten to 100 million year timescales related to the core–mantle boundary processes (McFadden et al., 1991; Tarduno et al., 2002); this may account for variability such as apparently low S observed from the ~1.8 Ga Taihang and ~1.1 Ga Mamainse Point lavas (Fig. 1A). We interpret here only the longer-term signal. To test for differences between the pre- and post-2.2 Ga data, we use a non-parametric Sign Test. The combined data were fit with Model G to produce an expected S curve (Table 1). When the difference between the observed and fit data is negative (positive), a minus (plus) is assigned. This comparison versus the Model G fit to all data suggests that the pre- and post-2.2 Ga data are different at the 78% confidence level.

Directions from extrusives are usually obtained from spatially limited stratigraphic sections and these may be particularly prone to undersampling of the field due to rapid lava emplacement. In contrast, studies of dikes often represent greater spatial sampling and are less likely to sample extremely short magmatic pulses. To further test our conclusions, we fit Model G to the intrusive data sets only (Fig. 1B), yielding $a = 0.21 \pm 0.07$, $b = 11.56 \pm 1.43$ and $a = 0.20 \pm 0.03$, $b = 7.66 \pm 1.13$ for the <2.2 Ga and >2.2 Ga groups, respectively. These values are indistinguishable from those obtained from fitting the total data set, supporting the difference in PSV.

3. Discussion and conclusion

According to the Model G (Eq. (2)), the equatorial intercept (the parameter b) of a PSV curve reflects the contribution only from the quadrupole (symmetric) family. Consequently, the lower values of b indicate a stronger contribution from the dipole (antisymmetric) family that includes the axial dipole and octupole. As noted by the authors of Model G, and emphasized by others (e.g., Hulot and Gallet, 1996), the model by itself does not discern the relative strength of the dipole and higher order components such as the octupole. However, analyses of the time-averaged 0–5 Ma field (e.g., Johnson et al., 2008), Cretaceous field (Tarduno et al., 2002) and Proterozoic field (Evans, 2006) have failed to detect significant octupole components, and very large contributions relative to those of the dipole would be needed to influence our interpretations. Therefore we feel that the interpretation of lower b values as reflecting higher contribution from the axial dipole is justified, although we note that *sensu stricto* some higher octupole contribution cannot be excluded from the data analyses we present alone.

When compared with data for the last 5 million years (Johnson et al., 2008; Lawrence et al., 2009; Kent et al., 2010; Opdyke et al., 2010; $a = 0.25 \pm 0.03$, $b = 13.24 \pm 0.81$) we find that data from both time windows suggest a more dipolar field; the further observation that data from the pre-2.2 Ga window suggest a more dipolar field than that in the post-2.2 Ga interval implies that the process causing the change of PSV was operating by at least ~2.2 billion years ago. The trend is even more expressed when the comparison is made versus the data for 5–195 Ma ($a = 0.25 \pm 0.04$, $b = 14.10 \pm 1.24$; McFadden et al., 1991; Tarduno et al., 2002) (Fig. 1). We note that Biggin et al. (2008b) claimed that differences in their analysis of Mid-Cretaceous and Jurassic PSV differed from those of McFadden et al. (1991) because the latter authors used a constant value to correct site-level data, which can impart a bias that is especially apparent for low latitudes. Although it is as yet unclear whether this explanation is correct (e.g., it may more simply relate to the use of some more extensive data sets), some bias does exist with the use of a constant value. Because this affects low latitude preferentially, lowering the S value, if present this bias would lead us to conclude that the 5–195 Ma average has a lower-than-actual b value. In this sense, our handling of the data is conservative because it would lead us to believe there was less of a difference between the 5–195 Ma data and the Proterozoic/Neoarchean data sets than what actually existed.

There are several processes that could be recorded by the PSV signal. Forcing of the dynamo could have changed in the absence of inner core growth (e.g., Olson and Christensen, 2006; Hori et al., 2010). However, these models predict lower CMB heat flow during superchrons, something that appears to be inconsistent with geological observations during the best known superchron, the Cretaceous Normal Polarity Superchron. Specifically, this interval is marked by extraordinarily high mantle plume activity, during which the giant oceanic plateaus such as Ontong Java formed (Larson, 1991; Tarduno et al., 1991). Some studies suggest that classical values of core heat flux (e.g., Sleep, 1990) based on hotspot topography should be raised by as much as a factor of three (e.g., Bunge, 2005). These considerations lead to the prediction of high values of CMB heat flow during the Cretaceous Normal Polarity Superchron. Moreover, relying on a change in forcing to explain the PSV signal requires a change in core–mantle boundary conditions. While this is straightforward to change in a numerical model, it carries with it several important implications for the ancient Earth. We outline some of the important issues below.

First, a potential problem with relying on conditions at the core–mantle boundary alone to account for the signal we have observed relates to the overall dipolarity of the signal. Notwithstanding models which do predict dipolar fields (e.g., Olson and Christensen, 2006), a range of other experimental results and numerical simulations suggest that a thin shell dynamo (possible when core–mantle boundary heat flux alone drives the system) may produce highly nondipolar fields (e.g., Stanley and Bloxham, 2004, 2006). While the deeper core would probably also convect in the case of Earth without an inner core and driven by CMB heat transport alone, the relative importance of the shallow convection is greater, leading to a more nondipolar field. While we have detected a trend toward more nondipolar fields from Neoarchean to Proterozoic times, it should be emphasized that overall the field throughout the entire interval was dipolar. Second, the most typical way one evokes a change in CMB boundary conditions in the Mesozoic to Recent Earth is to call upon deep subduction changing D″ heterogeneity. For the Neoarchean to Proterozoic interval under consideration, some feel plate tectonics (e.g., Stern, 2005) and therefore deep subduction as a causal agent in creating CMB change was not operating. While we favor an early onset of subduction because there is evidence that it acted at least locally in Archean times (e.g., Mints et al., 2010), we nevertheless recognize that it is not a trivial matter to dramatically change CMB conditions in a way that is compatible with the current generation of numerical models (e.g., Hori et al., 2010) for the earlier times represented in our data set.

In contrast to arbitrarily changing forcing, a simpler way to ensure large scale flows consistent with a highly dipolar field is to have a source of compositionally-driven buoyancy at depth, namely the onset of inner core growth. We emphasize that the principal effect of inner core growth we call upon here is buoyancy, not geometry (Coe and Glatzmaier, 2006).

We outline the following three-stage scenario of core evolution (Fig. 2) to explain the PSV signal. Prior to the oldest record of the geodynamo at 3.45 Ga (Tarduno et al., 2010), the core may have been entirely liquid. A geomagnetic field may not have been generated if a dense liquid layer existed at the base of the mantle, the relict of which is now found as layers of ultra-low velocity near the core–mantle boundary (Labrosse et al., 2007). A null or weak field at 3.8–3.9 Ga is suggested by a hypothesis seeking to explain lunar nitrogen values through transport from Earth's atmosphere by the solar wind (Ozima et al., 2005). After breakdown of the dense liquid layer, a geodynamo may have been present. But sources of field generation in the shallow outer core, related to convection associated with heat transport across the core–mantle boundary (Fig. 2A) could have produced a less dipolar field than that of latter times. Plate tectonics may have started very early on Earth, but cooling relevant for generation of the geodynamo requires cooling of the lower mantle. We envision this cooling accumulating with the penetration of slabs into the lower mantle, favoring super-adiabatic conditions and possibly inner core growth (Fig. 2B). The geodynamo at this time was deeply seated, related to compositional convection associated with inner core growth. The resulting field was highly dipolar, and is recorded by the oldest time window examined here (Fig. 2B). Subsequent subduction resulted in core–mantle boundary compositional and heat flux heterogeneity, resulting in sources for shallow field generation in addition to deeper sources near the inner core/outer core boundary. While the overall field was still dipolar, it was less so than prior to ca. 2 Ga (Fig. 2C). These field generation regions are similar to those of more recent times (Hoffman and Singer, 2008), with the exception that the inner core was smaller. We note that the decrease in the field dipolarity may also be promoted by increase in the CMB heat flow as suggested by some models (e.g., Hori et al., 2010), but we caution that these models still need to be

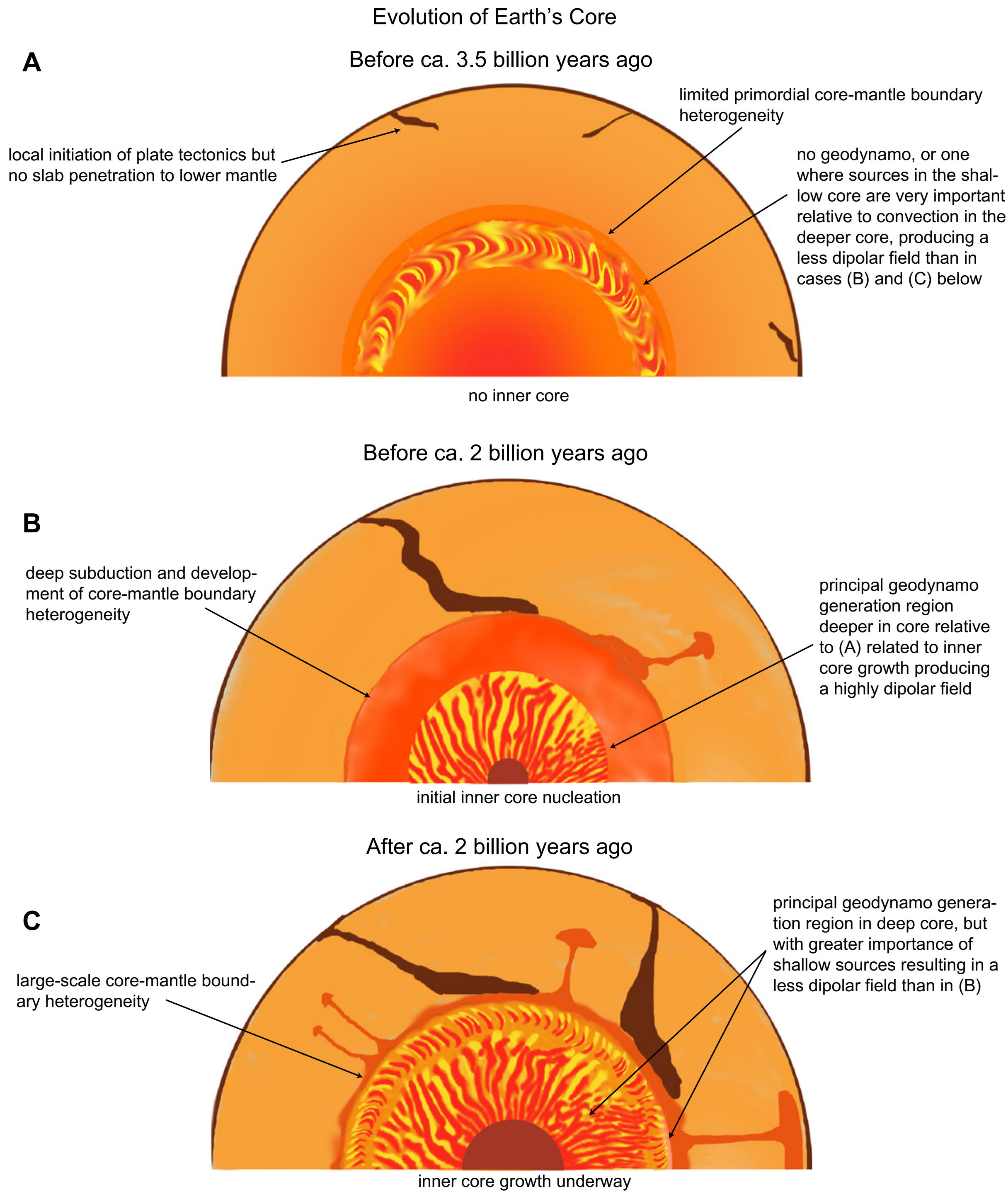

Fig. 2. One scenario for core evolution consistent with paleosecular variation and paleointensity data discussed here. Hypothetical equatorial Earth cross sections, highlighting regions of convective flow within the core most important for the dynamo. (A) Before approximately 3.5 Ga, an entirely liquid core may not have hosted a geodynamo (e.g., Labrosse et al., 2007; Ozima et al., 2005). However, given sufficient heat transport across the core–mantle boundary, a geodynamo could have been generated. If so, sources in the shallow outer core may have been more important for generating the dynamo relative to deeper convection, resulting in a field that was less dipolar than that generated in later times (B–C). (B) Onset of inner core nucleation sometime before approximately 2 Ga is driven by secular cooling of the lower mantle, possibly related to deep subduction. This results in a geodynamo that is more deeply seated in the core producing a highly dipolar field. (C) With the development of core–mantle boundary heterogeneity by continued deep subduction, shallow core contributions to the geomagnetic field grow in importance, resulting in a less dipolar field than in (B).

rigorously examined against the Mesozoic-Recent interval where geologic data may be used to infer changes in CMB conditions.

The relatively old inner core age implied by our PSV analysis favors radioactive heat sources in the core (Buffett, 2002). However, we note that the inner core nucleation age we call upon is older than that envisioned in many models (e.g., Aubert et al., 2009, 2010). Resolution of this important question should come as numerical models improve and are able to accommodate values representing the real Earth, and PSV data sets become larger. In particular, the possibility that a change in core cooling explicitly related to deep subduction as envisioned here led to inner core formation should be considered in future modeling. Finally, we note that the paleomagnetic data (Table 1) were generally collected for paleolatitude (tectonic) studies. The change we have identified is testable through renewed paleomagnetic studies of igneous units, with an eye toward dense sampling needed to reduce uncertainties in PSV analyses. Additional efforts should also be made to obtain robust PSV estimates for the time periods for which such estimates are currently rare or absent (for example, for the early/mid-Mesoproterozoic).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.pepi.2011.05.003.

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Supplementary Table 1. Summary of the data used for estimating the paleosecular variation after applying the variable cutoff filter (Vandamme, 1994). The three datasets affected by the procedure are highlighed by bold font.

Unit	Age (Ma)	B′	Δ_{max}	A$_c$	Plat	$S \pm dS$	C/UC	S_{exp}	S_{obs}-S_{exp}	Sign
Bangemall Basin Sills[1]	1070	11	24.1	31.1	27.8	13.9 ± 2.6	C	11.3	2.6	+
Lake Shore traps[2]	1087	30	18.3	29.7	14.6	13.6 ± 0.8	C	10.1	3.5	+
Portage Lake Volcanics[3]	1095	28	26.2	31.1	16.9	14.5 ± 1.2	UC	10.3	4.2	+
North Shore Traps[4]	1098	34	23.6	26.1	27.3	11.4 ± 2.1	C	11.2	0.2	+
Mamainse Point Upper N[5]	~1100	21	10.1	14.8	22.4	4.8 ± 3.0	C	10.7	-5.9	–
Umkondo dolerites[6]	1110	15	28.0	31.3	6.8	14.2 ± 3.1	C	9.7	4.5	+
Cleaver dikes[7]	1740	17	30.7	33.2	39.1	14.4 ± 4.0	C	12.6	1.8	+
Taihang dikes[8]	**1769**	**18**	11.6	**16.2**	**2.3**	**5.9 ± 1.5**	**C**	**9.6**	**-3.7**	–
Flaherty volcanics[9]	1870	11	17.4	25.9	26.2	11.6 ± 1.7	UC	11.1	0.5	+
Mashonaland dolerites[10]	1880	**15**	21.2	**28.5**	**27.4**	**11.8 ± 5.1**	**C**	**11.2**	**0.6**	+
Fort Frances dikes[11,12]	2067-2077	12	26.0	26.0	35.1	11.2 ± 3.3	C	12.1	-0.9	–
Marathon dikes (R)[11,12]	2101-2106	13	21.2	31.1	37.0	14.0 ± 2.5	C	12.4	1.6	+
Marathon dikes (N)[11,12]	2121-2126	16	26.3	34.8	39.9	16.0 ± 2.6	C	12.8	3.2	+
Biscotasing dikes[13]	2169	12	20.7	28.2	38.4	12.6 ± 2.3	C	12.6	0	0
Ongeluk lavas[14]	2200	32	19.5	20.1	14.1	7.1 ± 7.5	C	10.1	-3.0	–
Dharwar dikes[15]	2367	25	24.7	32.8	69.1	15.5 ± 1.1	C	17.4	-1.9	–
Matachewan dikes (N)[16]	2473-2446	28	16.7	17.5	14.8	6.4 ± 2.9	C	10.1	-3.7	–
Matachewan dikes (R)[16]	2473-2446	**99**	18.3	**20.4**	**7.7**	**8.0 ± 5.5**	**C**	**9.8**	**-1.8**	–
Karelia dikes[17]	2440	11	16.6	24.2	30.0	10.7 ± 1.3	UC	11.5	-0.8	–
Allanridge lavas[18–19]	2664-2709	17	25.5	29.0	43.2	12.1 ± 5.7	C	13.2	-1.1	–
Upper Fortescue lavas[20]	2715	16	21.8	30.0	35.0	12.9 ± 3.2	C	12.1	0.8	+
Fortescue Lower lavas[20]	2746	75	28.7	30.7	49.5	12.8 ± 2.0	C	14.2	-1.4	–
Fortescue Package 0[20]	>2772	24	26.4	36.5	58.3	16.1 ± 2.8	C	15.6	0.5	+

B′: number of units not rejected by the variable cutoff; Δ_{max}: maximum angular distance of the non-excluded VGPs from the mean paleopole; A$_c$: the variable cutoff angle; Plat: paleolatitude; S, dS: angular dispersion of VGPs and confidence interval; C/UC: data corrected (uncorrected) for within-site dispersion; S_{obs}-S_{exp}: difference between observed S and that predicted (S_{exp}) from Model G fit to all data ($a = 0.21 \pm 0.03, b = 9.62 \pm 0.94$); Sign: result of Sign Test (see text). Paleomagnetic and age data sources: Global Paleomagnetic Database (GPD) (Pisarevsky, 2005) data identifier is listed; see references cited for more recent works: [1]Wingate et al., 2002 (GPD 3455), [2]Diehl and Haig, 1994 (GPD 2776); [3]Hnat et al. 2006; [4]Tauxe and Kodama, 2009; [5] Swanson-Hysell et al., 2009; [6]Gose et al. 2006; [7]Irving et al. 2004 (GPD 3609); [8]Halls et al. 2000 (GPD 3394); [9]Schmidt 1980 (GPD 1862); [10]Bates and Jones, 1996 (GPD 3088); [11]Buchan et al. 1996 (GPD 3061); [12]Halls et al. 2008; [13]Halls and Davis 2004 (GPD 3644); [14] Evans et al., 1997 (GPD 3175); [15]Halls et al. 2007; [16]Evans and Halls 2010; [17]Mertanen et al. 1999 (GPD 3296); [18] de Kock et al. 2009; [19] Strik et al. 2007; [20]Biggin et al. 2008.

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Supplementary Figure 1. (A) Latitudinal dependence of angular dispersion S of virtual geomagnetic poles for the Precambrian (solid symbols) intrusive and extrusive units and extrusives of the last five million years (open inverted triangles). The Precambrian datasets are filtered using the variable cutoff angle (Vandamme, 1994). See Supplementary Table 1 for the Precambrian data sources. Grey and black symbols: younger and older than 2.2 Ga, respectively. Solid black, grey, and thick black lines: Model G fits for the 0-5 Ma (Time-Averaged Field Initiative, TAFI; Johnson et al., 2008; Lawrence et al., 2009; Kent et al., 2010; Opdyke et al., 2010), 1.0-2.2 Ga and 2.2-3.0 Ga data, respectively. The dashed line shows the model G fit for the 5-195 Ma data (McFadden et al., 1991; Tarduno et al., 2002; note individual data points are not shown here) (see text). (B) Latitudinal dependence of S only for the Precambrian (solid symbols) intrusive units (see text).