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Alec R. Brenner
Harvard University

Roger R. Fu
Harvard University

David A.D. Evans
Yale University

Aleksey V. Smirnov
Michigan Technological University, asmirnov@mtu.edu

Raisa Trubko
Harvard University

See next page for additional authors

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Authors
Alec R. Brenner, Roger R. Fu, David A.D. Evans, Aleksey V. Smirnov, Raisa Trubko, and Ian R. Rose

GEOLOGY

Paleomagnetic evidence for modern-like plate motion velocities at 3.2 Ga

Alec R. Brenner[1]*, Roger R. Fu[1], David A.D. Evans[2], Aleksey V. Smirnov[3], Raisa Trubko[1,4], Ian R. Rose[2,5]

The mode and rates of tectonic processes and lithospheric growth during the Archean [4.0 to 2.5 billion years (Ga) ago] are subjects of considerable debate. Paleomagnetism may contribute to the discussion by quantifying past plate velocities. We report a paleomagnetic pole for the ~3180 million year (Ma) old Honeyeater Basalt of the East Pilbara Craton, Western Australia, supported by a positive fold test and micromagnetic imaging. Comparison of the 44°±15° Honeyeater Basalt paleolatitude with previously reported paleolatitudes requires that the average latitudinal drift rate of the East Pilbara was ≥2.5 cm/year during the ~170 Ma preceding 3180 Ma ago, a velocity comparable with those of modern plates. This result is the earliest unambiguous evidence yet uncovered for long-range lithospheric motion. Assuming this motion is due primarily to plate motion instead of true polar wander, the result is consistent with uniformitarian or episodic tectonic processes in place by 3.2 Ga ago.

INTRODUCTION

Plate tectonics has been the dominant surface geodynamical regime throughout Earth's recent geological history. One defining feature of modern plate tectonics is the differential horizontal motion of rigid lithospheric plates. The physiography and composition of Earth's modern crust bear evidence for plate tectonic or "mobile-lid" processes including subduction, collisional orogeny, rifting, and ocean spreading. The case for the Archean Earth [4.0 to 2.5 billion years (Ga) ago] is not so clear. The surviving Archean crust consists of ~35 cratons (*1*), most with a characteristic architecture of rounded granitoid intrusive domes rimmed by steeply dipping greenstone keels (*2*). The composition of extant Archean crust is substantially more mafic than modern oceanic crust, with a high fraction of ultramafic rocks such as komatiites (*2*–*4*). These structural and compositional differences have led to a number of proposals that the Archean crust was constructed by exotic processes, including plume tectonics (*5*, *6*), sagduction/drip tectonics (*6*, *7*), and a vertically overturning lithosphere (*8*, *9*). Since some of these processes are difficult to reconcile with plate mobility, alternative geodynamical regimes have been proposed for the Archean Earth, including stagnant-lid and sluggish-lid modes (*10*, *11*) in which the lithosphere was rendered immobile, or at least slowed, due to decoupling from the asthenosphere under elevated geothermal gradients (*12*). Other studies argue for a uniformitarian model of the Archean Earth, in which some variant of modern plate tectonics was in operation at least locally throughout Earth history (*13*). Complete understanding of the Archean lithosphere, hydrosphere, atmosphere, and biosphere are predicated upon distinguishing between these proposed Archean geodynamic modes. Insights into these components of the early Earth are foundational to the inner workings of terrestrial planets generally and what surface conditions and environments hosted the development of the first life.

[1]Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA, USA. [2]Department of Geology and Geophysics, Yale University, New Haven, CT, USA. [3]Department of Geological and Mining Engineering and Sciences, Michigan Technological University, Houghton, MI, USA. [4]Department of Physics, Harvard University, Cambridge, MA, USA. [5]Department of Earth and Planetary Sciences, University of California Berkeley, Berkeley, CA, USA.
*Corresponding author. Email: alecbrenner@g.harvard.edu

Arguments for alternative geodynamical regimes in the Archean are often based on inferences of a regime transition toward modern-style plate tectonics. Existing estimates for the age of such a transition range from the Neoproterozoic to the Hadean (see the Supplementary Materials) and invoke a range of observations including global and local geochemical records (*14*–*16*), field relations of possible syn-tectonic rocks (*17*–*19*), and paleomagnetic pole comparisons (*20*–*23*).

A key discriminant between stagnant- and mobile-lid regimes is the rate of horizontal motion of plates over Earth's surface. Absolute plate velocities have typically been ~2 to 10 cm/year (extremes from 0 to 25 cm/year) over the last 400 million years (Ma) (*24*), while hypothesized velocities for stagnant- and sluggish-lid models are typically less than 2 cm/year (*25*). Paleomagnetic methods may constrain the velocity of crustal blocks in deep geological time by measuring their apparent polar wander histories. However, robust paleomagnetic evidence for latitudinal motion has been lacking thus far for times before 2.8 Ga (*20*–*23*). Here, we produce a new paleomagnetic pole from ~3180 Ma volcanics in the East Pilbara Craton of Western Australia and use this result to assess the presence of plate tectonic–like processes on Earth before that time.

Geologic setting of the Honeyeater Basalt

Greenstone belts in the East Pilbara Craton of Western Australia and the Kaapvaal Craton of South Africa are two of the only well-exposed regions on Earth that preserve ≥3.0 Ga rocks metamorphosed to greenschist facies conditions or lower (250° to 500°C, 1 to 8 kbar), making them potentially suitable for paleomagnetic analysis. Archean paleopoles previously reported from the East Pilbara fall into Neoarchean (<2.8 Ga) and Paleoarchean (3.35 to 3.5 Ga) groups (table S3) and are discussed in detail in the Supplementary Materials ("Mesoarchean Paleogeography"). Paleomagnetic study of a rock unit with an intermediate age between existing Paleoarchean and Neoarchean poles can potentially refine the history of East Pilbara motion before 2.8 Ga. One such candidate is the ~3220- to 3170-Ma Soanesville Group (*26*), a predominantly mafic volcano-sedimentary succession that outcrops mainly along the western margin of the East Pilbara. The ~3220- to 3180-Ma lower Soanesville Group is a siliciclastic succession of up to 3500 m of sandstones, turbidites, conglomerates, and minor cherts and banded iron formations (*18*). The overlying Honeyeater Basalt (HEB; 3192 to 3176 Ma), which we sampled in this work, contains up to 1050 m of pillowed and massive flows of tholeiitic and komatiitic metabasalt (*18*). The lower Soanesville Group succession is cut by the mafic-ultramafic dikes and sills of the Dalton Suite, which is contemporaneous and probably comagmatic with the HEB (*18*). The 3192 to 3176 Ma age range of the HEB is derived from U-Pb Sensitive High-Resolution Ion Microprobe (SHRIMP) dating of magmatic and detrital zircons and baddeleyites from the Dalton Suite in five greenstone belts (*18*).

Previous studies have hypothesized a rift origin for the Soanesville Group (*18*, *27*, *28*), which may be linked to the initiation of a primitive Wilson cycle (*18*). Its fining-upward siliciclastic succession and subsequent basalts are consistent with a modern rift setting. In addition, the lower siliciclastic portion is syndepositionally deformed by extension and basement-involved normal faulting. If this interpretation is true, the Soanesville Group would represent structural evidence of Mesoarchean horizontal tectonics in the Pilbara Craton.

We sampled from two of the most extensive exposures of the HEB located in the Soanesville Syncline (SVS) and the East Strelley greenstone belt (ESGB; Fig. 1) (*18*). The SVS is a northeast-plunging open syncline that outcrops over an approximately100 km^2 region in the southern Soanesville greenstone belt (*29*). In the ESGB, a baddeleyite-bearing Dalton Suite sill gave a U-Pb SHRIMP age of 3182 ± 2 Ma, in agreement with the 3192 to 3176 Ma age range of the HEB based on U-Pb geochronology throughout the East Pilbara (*18*). Correlation of the HEB across the ESGB and SVS is well established thanks to its unique stratigraphic surroundings. The HEB just overlies a sequence (bottom to top) of komatiites, felsic volcanics, siliciclastics, and banded iron formations (the Kunagunarrina, Kangaroo Caves, Corboy, and Paddy Market Formations, respectively), all intruded by Dalton Suite sills and dikes (*18*). Diagnostically, this sequence repeats itself in both the ESGB and the SVS (*18*). The Corboy Formation is particularly distinctive, as it represents the first thick siliciclastic package in the East Pilbara (*27*). The age of folding in both the SVS and ESGB is ~2930 Ma based on structural comparison with similar folds of this age elsewhere in the Pilbara (*27*). This age corresponds to the widespread emplacement of granitoids of the Sisters Supersuite (c. 2955 to 2920 Ma) across the western Pilbara Craton, including those west of the SVS in the Yule Dome (*27*). Sinistral faulting in the region has been dated to 2936 to 2919 Ma and is synkinematic with granitic intrusions and folding (*29*, *30*).

In both the SVS and ESGB, the HEB is locally weakly metamorphosed to prehnite-pumpellyite facies (*31*), possibly representing the oldest unit in the East Pilbara, or on Earth, not metamorphosed to greenschist facies or higher. In addition, the available range of bedding attitudes across ~2930-Ma structures in both the SVS and ESGB enables a paleomagnetic fold test for constraining the age of magnetization in the HEB.

RESULTS

We extracted and performed thermal demagnetization on 235 oriented cores of the HEB from 22 sites in the SVS and 13 sites in the ESGB (Fig. 1). These samples represent five localities (SVA to SVE; SV indicates Soanesville Syncline) with distinct local bedding attitude from both limbs of the SVS, and two distinct localities (ESA and ESB; ES indicates East Strelley) in the ESGB. Together, these samples potentially permit two separate fold tests in the SVS and ESGB.

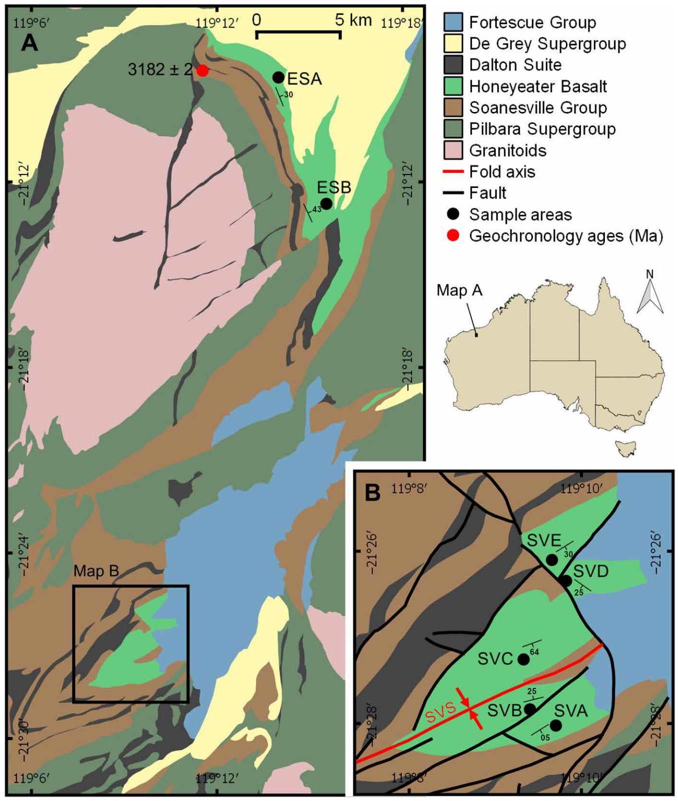

Fig. 1. Map of sampling localities. The HEB (light green) is especially well preserved in the Soanesville Syncline [SVS; see inset Map (**B**)] and East Strelley greenstone belt [ESGB; top of Map (**A**)], along with the underlying Soanesville Group sediments (brown) and Dalton Suite mafic intrusives (dark gray). The age of the HEB (3192 to 3176 Ma) in this area is derived from the age of the coeval Dalton Suite intrusives [e.g., one sample dated to 3182 ± 2 Ma in the ESGB (*18*)].

We identified three overprinting components of magnetization common to most samples. One component ("AF," alternating field) is unblocked by liquid nitrogen treatment and AF demagnetization up to 9 to 10 mT in 225 samples (Fig. 2). This component shows random directions in both in situ and stratigraphic coordinates with no observed relationship to the directions recovered during thermal demagnetization. Upon thermal demagnetization, a low-temperature component (L) found in 198 samples unblocks between ~75° and 300°C (Figs. 2 and 3A and fig. S2). Its direction in in situ coordinates (D, I = 357.4°, −44.2°; α_{95} = 7.0°) is consistent with the present-day geomagnetic field direction (I = −38.2° assuming a current Geocentric Axial Dipolar (GAD) geometry). We therefore conclude that the L component of magnetization is a recent viscous remanent magnetization. A medium-temperature component (M) appears in 181 samples, unblocks within ~200° to 500°C, and consistently displays a shallow northwest-upward mean direction in in situ coordinates (D, I = 310.3°, −19.0°; α_{95} = 5.7°) (Figs. 2 and 3A and fig. S2). In many samples from the SVS, a substantial fraction of this component unblocks between 315° and 355°C, suggesting the presence of pyrrhotite. We interpret the M component as a partial thermoremanent magnetization (pTRM) overprint acquired during low-grade regional metamorphism, possibly during the 2215- to 2145-Ma Ophthalmian Orogeny (*32*) or the 1830- to 1780-Ma Capricorn Orogeny, associated with tectonic accretion along the southern Pilbara margin or subsequent regional tectonic reactivation (*33*). This overprint direction is

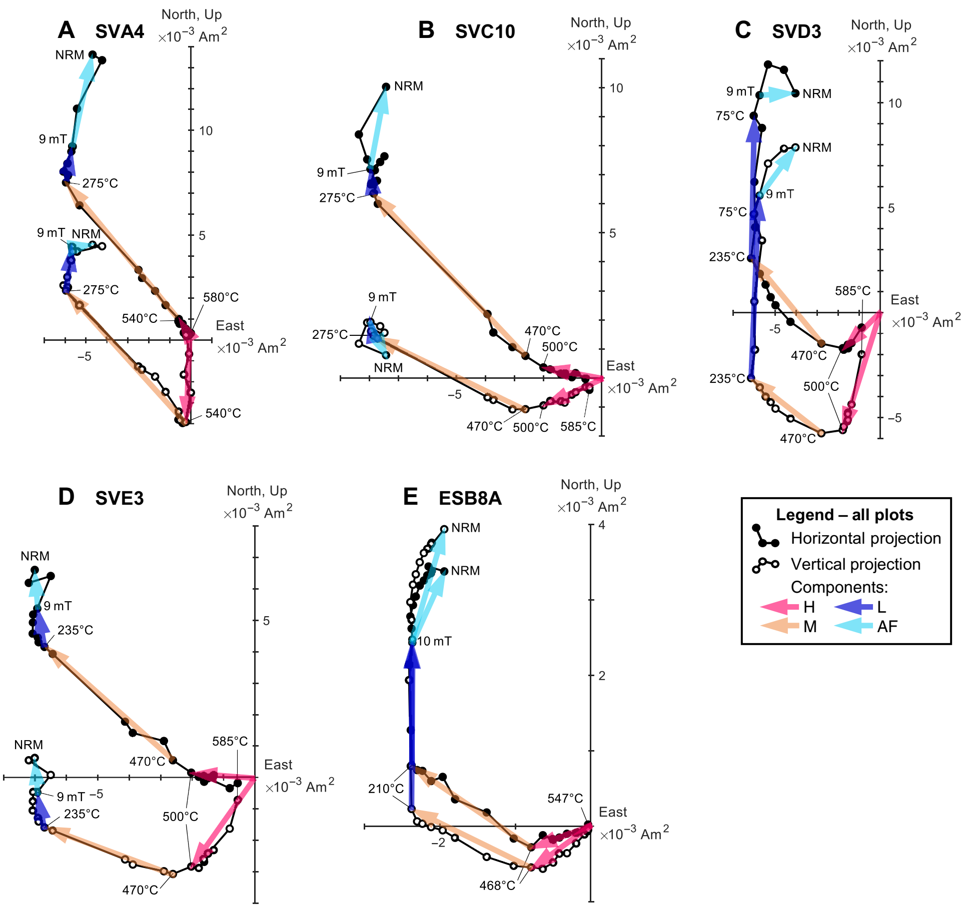

Fig. 2. Orthogonal projection diagrams of representative samples. All are in in situ coordinates and show four components; AF, light blue; L, dark blue; M, orange; and H, magenta (see the main text for discussion). (**A**) Sample SVA4. (**B**) SVC10. (**C**) SVD3. (**D**) SVE3. (**E**) ESB8A.

similar to the reversed counterparts of overprint directions from the Hamersley Basin and Tom Price iron formations ($D, I \sim 305°, -10°$) (*34*). Our interpretation of this magnetization as a metamorphic pTRM is also consistent with its typical maximum unblocking temperature range of ~400° to 500°C. The corresponding metamorphic temperature range in single-domain magnetite-dominated samples assuming time scales of 10 to 100 Ma for orogenesis is ~250° to 400°C (*35*), overlapping the conditions of prehnite-pumpellyite facies metamorphism observed in the HEB within our study area (*31*) (fig. S3).

The highest-temperature component of magnetization ("H") unblocks between ~400° and 590°C in 128 samples (Fig. 2). This maximum unblocking temperature and its origin-trending direction suggest that the H component is a magnetite-hosted TRM or thermochemical remanent magnetization (TCRM). In localities SVB and ESA, no sites showed coherence in H component directions and were thus excluded from further analysis. Localities SVA, SVC, SVD, SVE, and ESB contained between two and three sites each with coherent within-site directions ($n \geq 2$, $k \geq 15$; table S1). Site mean directions in these localities display reasonable scatter around the locality mean, with k values between 10 and 66, while locality means are mutually scattered in in situ coordinates (Fig. 3A).

After correction for folding (see Materials and Methods), localities SVA, SVC, SVD, and SVE H component site and locality means cluster about a southwest and down direction ($D, I = 203.7°, 59.0°$; $\alpha_{95} = 11.3°$; $N = 12$ sites; Fig. 3B and table S1). A three-limb fold test (*36*) using directions from SVC, SVD, and SVE using the nominal mapped SVS geometry (trend/plunge = 069°/41°) (*29*) passes with minimum scatter at 100 ± 30% (2σ) unfolding (Fig. 3C). Inclusion of locality SVA shifts the point of minimum scatter to ~79 ± 22% unfolding (Fig. 3D). In general, such an intermediate degree of unfolding at peak concentration may be due to subtle amounts of component mixing, diachronous magnetizations, syn-folding magnetization acquisition, local structural variability within the fold, or unrecognized vertical axis rotation (*36*). A final possibility is that the high-temperature magnetization in SVA cooling units was acquired over a short time interval insufficient for averaging paleosecular variation.

The steep H direction of SVA is similar to that of the ~2750-Ma Mount Jope Volcanics (MJV) (*37*). This possibility is unlikely, however, since the MJV pole was reported only from lavas in the Hamersley ranges far to the south, and the SVS-ESGB region is much closer to other Fortescue Group lavas (e.g., P0) with substantially different directions (*37*). The low metamorphic grade of the HEB and its ferromagnetic mineralogy (see below) also provide evidence against complete ~2930-Ma, syn-folding remagnetization. Further, although SVA is situated on a small fault-bounded block that may have rotated relative to the north limb of the syncline, the steep inclination of the SVA locality H directions implies that vertical axis rotations cannot substantially improve the concentration of unfolded directions. On the other hand, the high among-site precision parameter of SVA

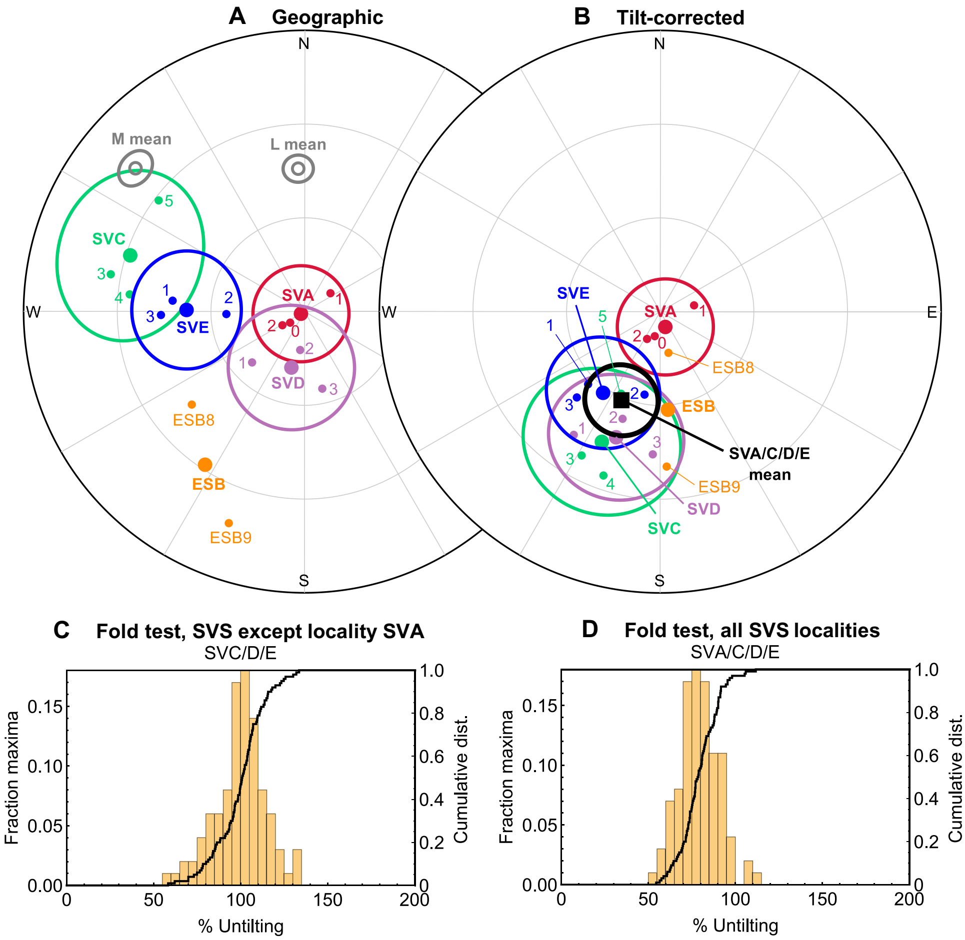

Fig. 3. Fold test on the H component. Equal-angle stereonet projections show measured directions in (**A**) in situ and (**B**) 100%-fold–corrected coordinates. Mean H component directions from each site (small points) are color-coded by locality means (large points with error circles; SVA, red; SVC, light green; SVD, light purple; SVE, blue; ESB, orange). The large error circle is omitted from the ESB mean direction for clarity. L and M component in situ means are gray. The tilt-corrected mean of all sites in SVA, SVC, SVD, and SVE is shown in black. Tilt correction in (B) includes correction for the plunge of the SVS fold axis. (**C**) Histogram (orange bars) and cumulative distribution (bold curve) of maximum post-unfolding fit in a fold test on the SVC, SVD, and SVE directions after Tauxe and Watson (*36*), using the 069°/41° mapped axis of the SVS. The fold test passes with best-fit unfolding 100 ± 15%. (**D**) A fold test on the SVA, SVC, SVD, and SVE directions with best fit at 79 ± 11% unfolding.

(k = 65.4), which is the highest of our sampled localities, supports the hypothesis that the site directions are not adequately sampling paleosecular variation. Such a scenario would apply, for instance, to a set of flows erupted in rapid succession atop one another, such as on the south flank of Kilauea, Hawai'i today, such that paleosecular variation is minimal over the time intervals between sites. We therefore posit that the SVA directions are biased by insufficient sampling of paleosecular variation, although an age of emplacement near 3180 Ma but distinct from SVC, SVD, and SVE by at least several million years cannot be ruled out. The latter scenario would imply substantial latitudinal velocity by ~3.18 Ga. In either case, positive fold test in three SVS localities constrains the age of the H component to at least 2930 Ma (*27*).

Area ESB's distinct direction may be attributable to its location within the ESGB block, which may have undergone unknown structural rotation with respect to the SVS (see the "Tests for Archean plate motion" section). While the placement of ESB supports its exclusion from our analysis, the similarity of the ESB H component direction (D, I = 175.5°, 58.5°; α_{95} = 89.1°; N = 2 sites; Fig. 3B and table S1) to those in the SVS hints at consistency of this component across multiple fold structures. Further sampling of HEB lavas with a range of bedding attitudes in the ESGB would be required to produce a fold test that could include locality ESB. We limit further discussion to our preferred H component direction and pole defined by localities SVA, SVC, SVD, and SVE.

Petrological and rock magnetic properties of the HEB may provide further insight into the timing of remanence acquisition. The low metamorphic grade of the HEB relative to the 590°C maximum unblocking temperature of the H component rules out a thermal overprint origin after 3182 Ma. To constrain the origin of H magnetization, we mapped the spatial distribution of magnetic sources hosting the natural remanent magnetization (NRM) using a Quantum Diamond Microscope (QDM; see Materials and Methods) (*38*). For this analysis, we chose SVD14, a sample with a very strong H component relative to total NRM moment (J/J_0 = 61%). The strongest sources of NRM were visible under an optical microscope as dark, opaque, sub-equant grains <5 μm in size, which is consistent with rock magnetic experiments on two samples from site ESB8 identifying pseudo-single-domain (PSD) magnetite as the major remanence-hosting phase (fig. S5). We find that these H magnetization-carrying grains are localized along the rims of several-millimeter laths of skeletal augite surrounding chlorite cores, most likely originally zoned clinopyroxenes where the cores were serpentinized to chlorite on the seafloor (figs. S3 and S4). The textural association of magnetic sources with the rims of these relict zoned pyroxene grains, in conjunction with rock magnetic data, suggests that they consist of magnetite precipitated during serpentinization, during which the reaction front would have been localized to the contact between the altering phases and the surrounding high-temperature fluids (*39*). The H component therefore likely represents a shortly post-emplacement seafloor TCRM, implying that the H magnetization was acquired near the time of eruption.

We measured magnetic hysteresis loops, backfield remagnetization of saturation remanence, and temperature dependence of low-field magnetic susceptibility (see Materials and Methods) on two samples with well-defined H components from site ESB8. The magnetic hysteresis behavior in both samples is dominated by paramagnetism (likely from clays). After paramagnetic slope correction, the measured magnetic hysteresis loops have a regular shape and indicate the presence of a low-to-intermediate coercivity magnetic phase (H_c = 6 to 11 mT; fig. S5, A and B). Magnetic hysteresis data did not reveal the presence of magnetically hard phases such as hematite. The M_{rs}/M_s (saturation remanence to saturation magnetization) and H_{cr}/H_c (coercivity of remanence to coercive force) ratios (ESB8B: $M_{rs}/M_s = 0.076$ and $H_{cr}/H_c = 3.72$; ESB8G: $M_{rs}/M_s = 1.42$ and $H_{cr}/H_c = 3.00$) suggest PSD magnetite (*40*, *41*) as a remanence carrier in both samples (fig. S5C).

The low-field susceptibility [$k(T)$] curves measured from ESB8B and ESB8G are irreversible, suggesting heating-induced alteration (fig. S5, D and E). For both samples, the preheating low-temperature $k(T)$ run shows a monotonic decrease of k with increasing temperature, whereas the high-temperature heating legs indicate the presence of a magnetic mineral phase with a broad (~400° to 600°C) range of Curie temperatures (T_c). We interpret this phase as cation-deficient magnetite. The strongly irreversible behavior observed above 600°C may indicate an additional metastable iron-rich phase with T_c > 700°C formed by thermal alteration of paramagnetic minerals (e.g., clays) (*42*). This phase subsequently converts to nearly stoichiometric magnetite as indicated by the Hopkinson peak at ~580° to 600°C seen in the cooling $k(T)$ legs for both samples and by the well-expressed Verwey transition (*43*) in the post-heating low-temperature $k(T)$ curves. The irreversible signal above 600°C may also be partially due to paramagnetic minerals directly converting to a nearly magnetite phase. The broader humps on the cooling legs of high-temperature $k(T)$ runs likely represent primary titanomagnetite (with the Curie temperatures shifted to lower values, potentially, due to stress relaxation during thermal treatment). In summary, these rock magnetic measurements are fully consistent with the magnetic mineralogy identified from our micromagnetic imaging experiments.

On the basis of the fold test result, the uniquely low metamorphic temperature of the HEB, the petrographic context of the remanence-carrying grains, and the high unblocking temperature and origin-trending nature of H magnetization, we conclude that the H component is a primary TRM or TCRM dating to emplacement of the HEB at 3182^{+10}_{-6} Ma ago (*18*). The paleopole for this direction (λ, ϕ = 60.6°S, 83.9°E; α_{95} = 15.3°), which we call the HEBh pole for "HEB, high temperature," corresponds to a paleolatitude of λ_p = 43.7° ± 15.3° (Figs. 4and 5A and table S2). The HEBh pole reduces the size of the largest unsampled gap in the available (i.e., after 3.5 Ga) paleomagnetic record (the ~580 Ma separating poles EBm and P0) by 170 Ma.

Accepting a primary origin for the H magnetization, the angular dispersion of its virtual geomagnetic poles (VGPs) may be used to characterize the variability of the geodynamo during the Mesoarchean. We compute an angular standard deviation of S_B = 19.5° ± 7.9° (N = 11 VGPs). This is greater than the ~12° dispersion expected at ~45° paleolatitude based on a fit to 2.2- to 3.0-Ga poles (*44*) and somewhat greater than values of S_B observed over the last 5 Ma (*45*). VGPs of the older, lower-latitude Duffer Formation (DFM) pole (*46*) show lower scatter with S_B = 9.7° ± 5.5° (N = 11 VGPs), potentially suggesting a more stable pre–3.0-Ga geodynamo. An important caveat of these calculations is that both have small sample sizes, which may bias S_B to lower values (*47*). In addition, the probable early serpentinization origin of the H magnetization complicates the interpretation of its VGP dispersion, as magnetite ingrowth may have occurred over periods long enough to partially average paleosecular variation. Chemical remanence directions are also commonly more scattered than thermal remanence directions [e.g., in serpentinized ultramafic intrusions (*48*)]. However, many sites display coherent directions distinct from those of adjacent sites, demonstrating that some directional changes due to paleosecular variation were indeed recorded and supporting our interpretation that the observed inter-site scatter reflects geodynamo stability.

DISCUSSION

Tests for Archean plate motion

Together with paleolatitudes from other poles previously reported from the East Pilbara, the new HEBh pole places new lower bounds on the horizontal drift rate of the East Pilbara between ~3.35 and 2.77 Ga relative to Earth's rotation axis, assuming a geocentric axial dipole field geometry at ~3.2 Ga. The youngest pre-HEBh pole from the East Pilbara is that of the 3350- to 3335-Ma Euro Basalt (EBm), which provides an estimated paleolatitude of 8.09° ± 5.3° (*49*). This paleolatitude is distinguishable from that of the HEB at 2σ and requires that the average latitudinal velocity of the Pilbara exceeded 0.23° ± 0.10°/Ma or ≥2.50 ± 1.15 cm/year between 3.35 and 3.18 Ga (Fig. 5). The oldest post-HEBh pole is that of Fortescue Group Package 0 (P0), with an age most likely within 10 Ma of its minimum age of 2772 ± 2 Ma (*37*, *50*). Its paleolatitude of 57.3° ± 8.0° implies an average latitudinal velocity of 0.03° ± 0.04°/Ma for the Pilbara between 3.18 and 2.77 Ga or ≥0.37 ± 0.47 cm/year (Fig. 5). These values are lower bounds on the rate of lithospheric motion for several reasons. The longitude of the Pilbara during the relevant time interval is unknown, implying that all above drift rates are lower bounds. In addition, the unknown polarity of the geomagnetic field precludes these rates from accounting for the sign of the paleolatitude.

In addition to the latitudinal translations quantified by inclination changes, declination changes between the P0, HEBh, and EBm poles may imply substantial vertical axis rotations. Approximately 124° of clockwise (CW) rotation occurred in the ~170 Ma between the EBm and HEBh poles (0.73°CW/Ma) and ~101° of counterclockwise (CCW) rotation in the ~410 Ma between the HEBh and P0 poles (0.25°CCW/Ma). These rotations may originate from plate-tectonic motion of the East Pilbara Craton, true polar wander (TPW), or post-emplacement local block rotations, perhaps during ~2950- to 2930-Ma deformation in the Lalla Rookh-Western Shaw Structural Corridor (LWSC) that sinistrally sheared the majority of the SVS and

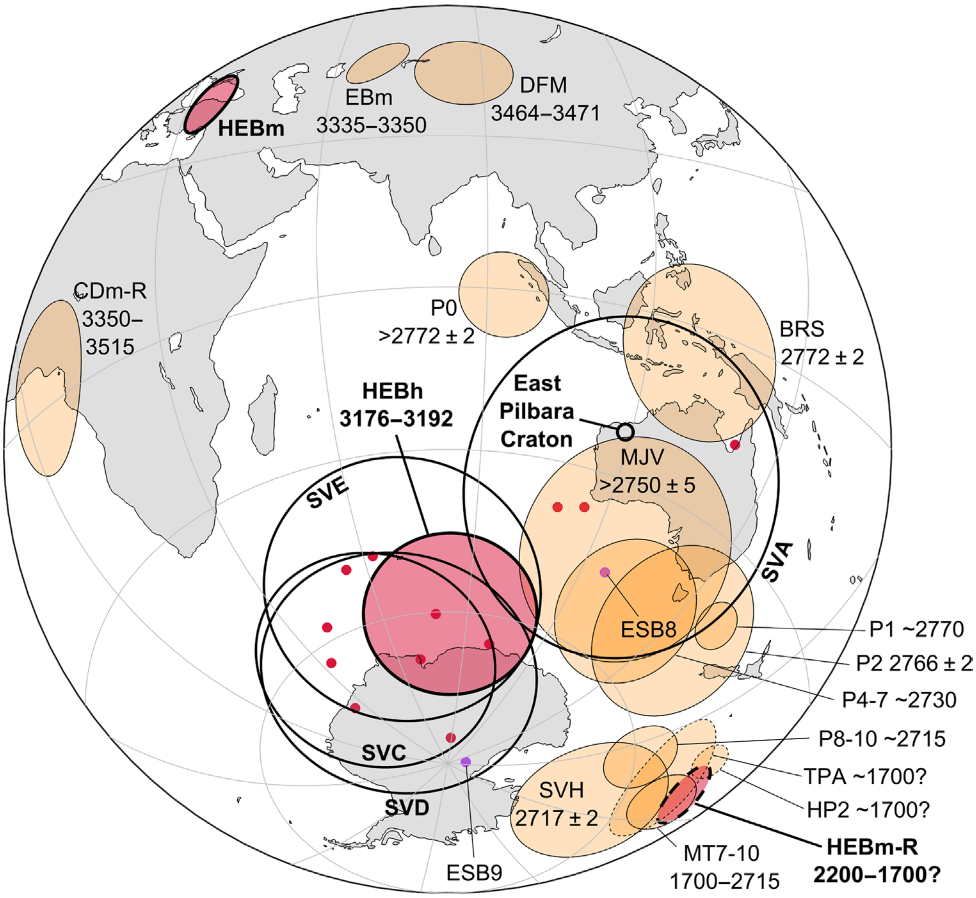

Fig. 4. Paleopoles from the Pilbara Craton. Stratigraphic coordinates apply except for HEBm and HEBm-R overprints in in situ coordinates. Numbers are ages in million years. Poles on the hemisphere opposite the viewer have dashed outlines. Poles reported in this study are red-filled circles and ellipses, locality-mean poles are black hollow circles, and site VGPs are red points (ESB sites in purple; ESB locality-mean pole not shown for clarity). Poles from other studies in the Pilbara are orange.

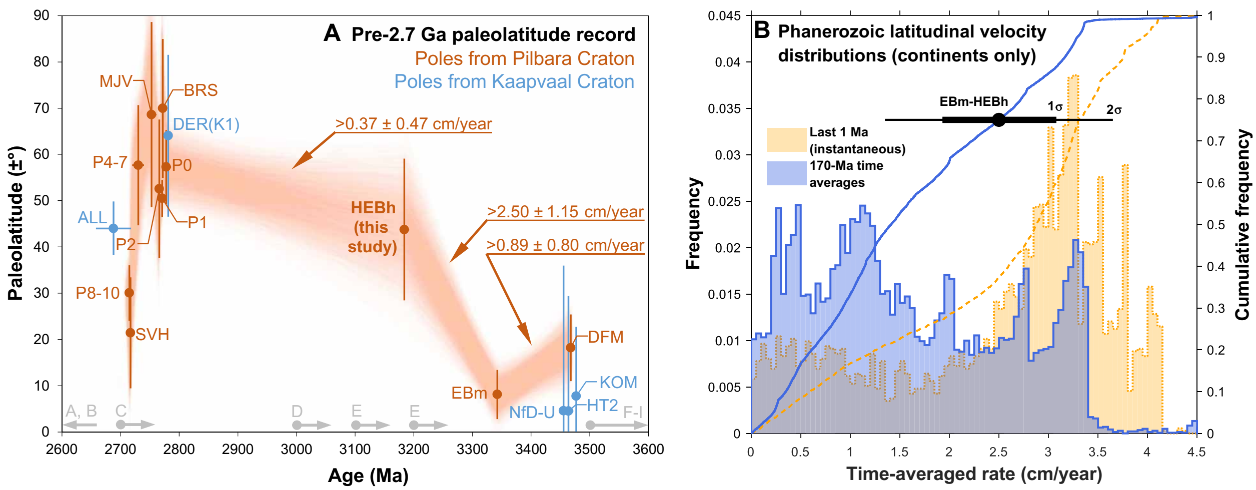

Fig. 5. Tests for Archean plate motion. (**A**) Meso-Paleoarchean paleolatitude records from the Pilbara Craton (Western Australia, orange) and Kaapvaal Craton (South Africa, blue), including this study (HEBh). Paleolatitude error bars are 2σ; age error bars are 2σ or the full range allowable given available constraints. The orange shaded region is a sample of the range of minimum-velocity paths allowed by existing poles. Gray arrows at the bottom are constraints on the onset date of mobile-lid plate tectonics from a sample of previous studies, with letters referencing studies listed in the Supplementary Materials. (**B**) Distributions of time-averaged latitudinal velocities on the continents (*24*) recorded at $N = 3736$ randomly selected points from the present to 1 Ma (yellow with dashed lines) and over 170 Ma windows throughout the Phanerozoic (blue with solid lines) as both histograms and cumulative distribution frequencies (line overlays). The 170 Ma distribution tends toward slower rates because it often samples reversals in plate velocity. For comparison, the minimum motion rate implied by the EBm and HEBh poles is shown by the black point and error bars. This rate, which is averaged over a 170 Ma interval, is comparable with geologically modern rates.

ESGB (*29*, *30*). The region in which the EBm pole was measured (*49*) would not have rotated substantially during this episode (*29*, *30*). Local effects related to the impingement of the nearby Shaw, Yule, and Strelley batholiths apparently both produced the SVS and rotated its central portion CW (*29*, *30*), thus potentially accounting for some of the observed EBm-HEBh rotation. This would also imply that the rotation measured between the HEBh and P0 poles is an underestimation, since CW structural rotation in the LWSC would reduce the observed CCW rotation. Regardless, we cannot determine whether either of the observed vertical axis rotations is due to purely local structural effects, rotation of the entire East Pilbara, or a combination of both processes.

The paleomagnetic record thus indicates that the East Pilbara experienced horizontal motion of ≥2.50 ± 1.15 cm/year in the ~170 Ma between 3350 and 3180 Ma and ≥0.37 ± 0.47 cm/year in the ~410 Ma between 3180 and 2772 Ma (Fig. 5A). The former rate is generally inconsistent with hypothesized "hard" stagnant-lid surface velocities and observed rates of Phanerozoic net lithosphere rotation (typically <2 cm/year) (*25*, *51*). It falls between the 48th and 99th percentiles of the Phanerozoic distribution (since 410 Ma) of latitudinal continental plate velocities (*24*) time-averaged over 170 Ma windows, as estimated from random point sampling of continental lithosphere (Fig. 5B; see Materials and Methods). We chose these windows to cover the last full Wilson cycle, sampling from the full range of tectonic settings encountered on the geodynamically "modern" Earth (figs. S6 and S7). The paleolatitudinal velocity of the Pilbara between 3.35 and 3.18 Ga is therefore comparable to those observed in modern plate motion, conforming to either a uniformitarian or episodic model of Archean tectonics (Fig. 5B). An episodic regime would combine a stagnant-lid state with occasional rapid mobile-lid plate motion (*52*) and is possible for both the pre– and post–3.18-Ga periods. In this case, our minimum rates would represent a mix of substantial tectonic drift velocities (>2.50 cm/year) and episodes of slower motion. Hence, we cannot rigorously rule out episodic transitions between a stagnant- or sluggish-lid regime and intervals of modern-style tectonics before 2.8 Ga without further paleomagnetic studies sampling the motion of multiple Archean cratons over identical, closely spaced time intervals.

The relative contributions of TPW (*53*, *54*) and differential plate motion to these rates are unknown, representing another opportunity for further work. Resolving the TPW and differential components of motion during 3.35 to 3.18 Ga would require ~3.35-Ga and 3.18-Ga paleopoles from another craton. Distinguishing between these contributions to net motion is an important test of Archean tectonic style, since motion of a stagnant lid would manifest as a net rotation of the lithosphere (i.e., indistinguishable from TPW in the paleomagnetic record), while mobile-lid tectonic processes would produce substantial differential motion. Measurement of substantial TPW during the Archean would be a noteworthy result, suggesting a less stable moment tensor, and perhaps more vigorous convection or a hemispherically asymmetric lithosphere.

CONCLUSIONS

While plate tectonics have characterized Earth's geodynamics in recent geologic time, it is unknown whether long-range horizontal motion of lithospheric plates occurred before ~2.7 Ga. Resolving this uncertainty would fundamentally contribute to understanding the formation settings of Earth's earliest crust and nascent biosphere and the evolution of geodynamics in terrestrial planets in general. We use paleomagnetic methods to isolate a high-temperature magnetization in the 3180-Ma HEB of the East Pilbara Craton that unblocks beyond the peak metamorphic temperature of the basalt (~250°C). The restored paleomagnetic directions robustly pass a fold test within the ~2930-Ma SVS. These observations, combined with magnetic microscopy and rock magnetic results, suggest a primary origin for the magnetization. This result establishes a paleolatitude of 43.7° ± 15.3° for the East Pilbara at ~3180 Ma and substantially reduces the duration of the longest unsampled gap in the available paleomagnetic record. With this new record, we show that paleolatitudes of the East Pilbara require an average latitudinal drift rate of at least 0.23°/Ma or 2.50 cm/year between 3350 and 3180 Ma. Since this time-averaged rate of horizontal motion is typical of modern plates, the paleomagnetic record is suggestive of either uniformitarian or episodic operation of plate tectonics in the East Pilbara before ~3.2 Ga. We cannot yet distinguish between differential plate motion, TPW, and net rotation explanations for the observed latitudinal motion, nor can we rule out modern-style or episodic plate tectonic motion before 3180 Ma. Constraining the relative contributions of each of these components represents a promising direction for future paleomagnetic studies in the East Pilbara and other cratons.

MATERIALS AND METHODS

We extracted 235 oriented cores of the HEB from 22 sites in the SVS (Fig. 1B) and 13 sites in the ESGB (Fig. 1A). These samples represent five localities (SVA to SVE) with distinct local bedding attitude from both limbs of the SVS and two distinct localities (ESA and ESB) in the ESGB. We drilled and field-oriented five to nine 2.5-cm core samples per site using magnetic and solar compasses while avoiding topographically prominent outcrops susceptible to lightning strikes where possible. Within each locality, we used a single paleohorizontal (bedding) attitude across all sites, thus attempting to average local variability of bedding attitudes we measured on rubbly flowtops and potentially undulatory chert interbeds.

We measured the magnetic moment of core samples using 2G Enterprises Model 755 Superconducting Rock Magnetometers at the Massachusetts Institute of Technology Paleomagnetism Laboratory and the Yale Paleomagnetic Facility. We subjected the SVS samples to AF demagnetization up to 9 mT in steps of 3 mT followed by thermal demagnetization up to 590°C in steps of 10° to 40°C. For ESGB samples, we first subjected them to immersion in liquid nitrogen to minimize the signal of multi-domain magnetite grains. We then applied AF demagnetization up to 10 mT in steps of 1 mT followed by thermal demagnetization up to 590°C in steps of 10° to 60°C. For some samples in the SVS, we averaged pairs of the highest-temperature demagnetization steps (>500°C) to suppress noise and to correct for a weak laboratory thermal remanence acquired in the oven. Because the cores were oriented such that they alternately pointed into and out of the oven during thermal demagnetization steps, this weak remanence, which was oriented approximately along the oven axis, was effectively eliminated by averaging pairs of thermal steps.

We quantified the directions of magnetization components using principal components analysis (*55*) and excluded components with mean angle of deviation greater than 20° from further analysis. For all but seven samples in which the highest-temperature component was not origin trending (likely as a result of weak pTRMs acquired within the oven), we forced the highest-temperature component of magnetization to include the origin. Use of non–origin-forced fits to all samples did not significantly change our results. We also rejected data from 20 samples that displayed strong, single-component origin-trending magnetizations characteristic of lightning remagnetization. We then calculated site mean directions from sample components after removal of 30 samples with outlier directions. Outliers were samples with H component directions that differed from well-clustered H components from other samples within the same site by more than ~30°, often because H component fits in such samples were based on too few points to reliably average noise in the demagnetization path. We retained in our analysis sites with at least two passing core samples and within-site Fisher precision parameter $k \geq 15$.

To compute the correct paleomagnetic directions in bedding coordinates, we determined the trend and plunge of the SVS fold hinge to be 069°/41° based on mapped bedding orientations (*29*). We therefore rotated site mean directions by 41° around the horizontal axis with trend 159° and applied a second horizontal axis rotation based on field observations of lava flow bedding attitudes to restore local paleohorizontal. Samples from the ESGB required only a single horizontal axis rotation to restore their original orientation, as the plunge of the SVS does not apply to the rocks of the ESGB.

To identify the remanence-carrying magnetic grains, we observed oriented sections of several cores with the QDM (*56*) at the Harvard Paleomagnetics Laboratory. The QDM uses optically detected magnetic resonance spectroscopy of nitrogen-vacancy (NV) centers in a diamond to produce maps of the surface vector magnetization on the samples with several-millimeter field of view and ~2-μm resolution. From each sample, we prepared a 1- to 2-mm-thick, 25-mm-diameter disk-shaped section and polished its upper face using 1-μm Al_2O_3 grit, taking care to avoid overprinting the sample's NRM. Placing the polished face of our samples in contact with the sensing diamond, we measured the magnetic field intensity along the [111] direction of the diamond crystal lattice using projective magnetic microscopy (*56*). We isolated the magnetic field signal of the ferromagnetic grains by measuring the sample successively under two mutually antiparallel 0.9 mT bias fields and computed the out-of-plane (B_z) magnetic field using a spectral algorithm (*57*). Once we located magnetic sources in these maps, we used a Horiba Scientific XploRA Plus Raman microscope (532 nm excitation wavelength) to identify the mineral phases surrounding them, thus providing petrographic context for the source(s) of remanence.

We measured magnetic hysteresis parameters (coercivity, H_c; coercivity of remanence, H_{cr}; saturation remanence, M_{rs}; and saturation magnetization, M_s) at room temperature using a Model 2900 Princeton Measurement Corporation Alternating Gradient Field Magnetometer at Michigan Tech. We used a maximum applied field of 1 T for magnetic hysteresis loops and of 0.5 T for backfield demagnetization of M_{rs} measurements. We also measured temperature dependences of low-field magnetic susceptibility [$k(T)$] upon cycling the samples between room temperature and 700°C (in argon) using an AGICO MFK1-FA magnetic susceptibility meter equipped with a high-temperature furnace and cryostat at Michigan Tech. The $k(T)$ curves were also measured upon heating from ~−180°C to room temperature before and after the high-temperature thermomagnetic runs. The field applied in all $k(T)$ experiments was 250 μT.

To measure paleosecular variation captured by the HEBh pole, we calculated the angular standard deviation S_B of its VGPs using the within-site dispersion correction after Doell (*58*)

$$S_B = \sqrt{S_{tot}^2 - \frac{S_w^2}{\bar{n}}} = \sqrt{\frac{1}{N-1}\sum_{i=1}^{N} \Delta_i^2 - \frac{1}{N}\frac{\left(\sum_{i=1}^{N}\frac{81°}{\sqrt{K_{wi}}}\right)^2}{\sum_{i=1}^{N} n_i}} \tag{1}$$

where N is the number of sites each with n_i samples, S_{tot} is the total dispersion, S_w is the average within-site dispersion, $\bar{n}$ is the average number of samples per site, Δ_i is the angular distance between the ith VGP and the mean of all N VGPs, and K_{wi} is the precision parameter of sample VGPs within the ith site. We calculated K_{wi} from individual site direction precisions k_i and paleolatitudes λ_i after Cox (*59*)

$$K_{wi} = k_i\left[\frac{1}{8}(5 + 18\sin^2\lambda_i + 9\sin^4\lambda_i)\right]^{-1} \tag{2}$$

We used only VGPs from localities SVA, SVC, SVD, and SVE and adopted a common cutoff angle of $\Delta_i < 45°$ to exclude transitional directions (*44*, *60*). We then used jackknife ($N - 1$) resampled values S_{Bi} of the mean $\bar{S}_B$ and estimated its uncertainty σ_{SB} after Efron (*61*)

$$\sigma_{S_B} = \sqrt{\frac{N-1}{N}\sum_{i=1}^{N}(S_{Bi} - \bar{S}_B)^2} \tag{3}$$

We used Monte Carlo resampling ($N = 10^6$) of the ages and paleolatitudes of each pole pair to estimate the uncertainty range of latitudinal plate motion rates between them. For poles DFM, EBm, HEBh, and P0, we assumed normally distributed age uncertainties with mean ± 2σ values 3467.5 ± 3.5, 3342.5 ± 7.5, 3184 ± 8, and 2778 ± 8 Ma, respectively. We took each pole's paleolatitude uncertainty to be normally distributed about its mean with $2\sigma = \alpha_{95}$. We converted rates from degrees per million years to centimeters per year with the factor 11.1 (cm/year)/(°/Ma).

To evaluate whether the latitudinal motion rates we measure are consistent with a mobile-lid setting, we selected 10,000 randomly distributed latitude and longitude points and filtered these to those on the continents only ($N = 3736$) since the HEB erupted atop the western margin of East Pilbara continental crust. We displaced these points from their present to their reconstructed positions at 1 Ma (i.e., over the 0- to 1-Ma window) and bracketing four 170-Ma-long windows (0 to 170, 80 to 250, 160 to 330, and 240 to 410 Ma) throughout the last Wilson cycle (fig. S6). We reconstructed points using GPlates with the paleomagnetic reference frame and the 410-Ma to present plate motion model of Matthews *et al.* (*24*). We then calculated the latitudinal displacement of all points across each reconstructed window. The distribution of latitudinal changes over the last 1 Ma approximates the instantaneous latitudinal drift rate distribution. The distribution over the total of all four 170 Ma windows simulates the distribution of rates calculated from two samples from the same location but 170 Ma apart, as is the case for the EBm and HEBh poles.

SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/6/17/eaaz8670/DC1

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