Geomagnetic reversal rates following Palaeozoic superchrons have a fast restart mechanism

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Long intervals of single geomagnetic polarity (superchrons) reflect geodynamo processes, driven by core–mantle boundary interactions; however, it is not clear what initiates the start and end of superchrons, other than superchrons probably reflect lower heat flow across the core–mantle boundary compared with adjacent intervals. Here geomagnetic polarity timescales, with confidence intervals, are constructed before and following the reverse polarity Kiaman (Carboniferous–Permian) and Moyero (Ordovician) superchrons, providing a window into the geodynamo processes. Similar to the Cretaceous, asymmetry in reversal rates is seen in the Palaeozoic superchrons, but the higher reversal rates imply higher heatflow thresholds for entering the superchron state. Similar to the Cretaceous superchron, unusually long-duration chrons characterize the ~10 Myr interval adjacent to the superchrons, indicating a transitional reversing state to the superchrons. This may relate to a weak pattern in the clustering of chron durations superimposed on the dominant random arrangement of chron durations.
non-reversing (superchrons) and reversing geodynamo states punctuate the Phanerozoic and Proterozoic, with reverse polarity superchrons during the early-mid-Ordovician (Moyero superchron) and during the late Carboniferous–mid Permian (Kiaman superchron) and normal polarity superchrons during the late Cretaceous (CNPS) and Proterozoic. The superchron transitions are thought to reflect a threshold between reversing and non-reversing states, with reducing heat flow (at superchrons’ start) or increasing heat flow (at superchrons’ end), across the core–mantle boundary (CMB). The heat flow changes likely respond mainly to mantle convention, since convective turnover time in the core is much shorter than that in the mantle. However, geodynamo models driven by geologically constrained mantle convention models highlight the different model heat flow constraints required to simulate the CNPS compared with the Kiaman and Moyero superchrons. Plate tectonic-driven models of the mantle suggest an increased heat flux during the CNPS, not reduced heat flux, and require that the lower mantle may be de-coupled or insulated from the main mantle flow. This conundrum has resulted in the ‘superplume’ hypothesis, in which the thermal piles in the lower mantle are inferred to expand upwards and later shrink on an ~200 Ma repeat time controlling the superchron cycles. More rapidly reducing CMB heat flow is inferred during the collapse phase of the ‘superplume’. Support for this hypothesis is the asymmetry in the polarity-reversal rates on either side of the CNPS, with lower reversal reversal rates following the end of the CNPS. It is not clear whether there is support for this ‘superplume’ model in the Palaeozoic superchrons, since reversal rates are an open discussion because of a lack of robust data. In the Mesozoic geomagnetic polarity timescale (GPTS), recovery from the CNPS starts with two long-duration chron (C33r and C33n) of some 4–5.5 Myr duration. This is followed by either stationary reversal behaviour, or a slow increase in reversal rates, over some 50 Myr before reversal rates stabilize to what they have been over the last 30 Myr (refs 7, 8). Slowly increasing reversal rates are also inferred following the Kiaman superchron, during the mid and late Permian into the Triassic. Crucially, the characteristics of the palaeomagnetic properties (rates of reversals, secular variation and field intensity) around the superchron start and end transitions hold clues to interpreting the geomagnetic response. Initiation of all superchrons is normally inferred to be pre-empted by a more rapid drop in reversal rates, evidenced by longer chron immediately pre-superchron, as exemplified during the early Cretaceous.

This work examines the 0–15 Myrs before and following the end of the Palaeozoic superchrons by providing a quantitatively derived polarity timescale as a proxy for the core–mantle behaviour driving the geodynamo. Comparisons of the reversal structure adjacent to the Phanerozoic superchrons provide clues to differences in the ending and restart conditions of the reversing geodynamo.

Results

Construction of geomagnetic polarity timescales. A robust construction of the polarity-reversal structure following the Kiaman and Moyero superchrons has not been made. Proposed end-Kiaman GPTS has both long chron of 2–3.5 Myr duration or numerous briefer chron. The data sets immediately before the Kiaman and Moyero superchrons (mid-Carboniferous and late-Cambrian–early Ordovician respectively) are less extensive. In the 2012 GPTS timescale for the Permian, Carboniferous and Ordovician, while the faunal ranges are scaled and converted into a composite position-scale using CONOP, the polarity chron are qualitatively attached to the biozonal scale, at the last stage, after the biozones are scaled to Myrs (Fig. 1). These procedures lose the information in the relative duration of chron embedded in the section data, but introduce additional guesswork, since the chron boundary positions in the scaled biozones need to be estimated in ways that are not defined. The later step also has considerable uncertainty, since relationships between chron boundaries and relevant biozonal boundaries are often not well defined (see uncertainty in biozone placement in Figs 2–4). The 2012 mid-late Permian GPTS is also based on only limited primary data sets from the Nammal Gorge, Wulong and Linshi sections. The mid-late Ordovician GPTS in the 2012 timescale is based on the small composite figure in ref. 14, with primarily the old British Ordovician regional stages for reference. The Carboniferous and late-Cambrian–early Ordovician GPTS in the 2012 timescale have similar limitations. Here a new more inclusive numerical approach is used by attaching the radiometric dates directly to a composite geomagnetic polarity scale, which uses multiple section data. Critically, this removes the transfer through a biozone scale to get at the estimated radiometric ages for the chron (Fig. 1). The GPTS produced uses a numerical optimization procedure, which generates a statistical composite of the geomagnetic polarity, initially in a scale of composite section height. The numerical method finds a solution that minimizes the misfits between the final composite and the section data, subject to chosen transformations that modify the height scales of the sections. From the final model composite, several statistical measures are determined. First, $D_s$ that assesses the proportion of relative

![Figure 1](https://example.com/image1.png)

**Figure 1** | Types of procedures used in polarity timescale constructions. (a) Procedures in geomagnetic polarity timescale construction for the Permian, Carboniferous and Ordovician as used in the 2012 timescale versus (b) the optimization method used here. The main additional procedures used in the 2012 polarity timescale in (a), but not used in the optimization method here, are the transfer through the CONOP-scaled biozones, the need to estimate polarity boundary position in the scaled biozones and the absence of chron uncertainty assessment.
misfits of the overall model; second, \( \sigma_T \) (in Myr), the s.d. of the transformed chron positions in the sections (about the position of the chron in the composite); third, \( D_T \) that measures the relative misfit of the magnetozone data in each section with respect to the final composite. These allow comparative assessment of the data consistency and the chosen transformations. Importantly, the method also produces confidence intervals on the chron ages. This new quantitative approach allows a more comprehensive assessment of the GPTS interval, since, most importantly, it directly uses the proxy for duration embedded in magnetozone-relative thicknesses in numerous sections.

In the new Palaeozoic polarity composites, the magnetochrons have been labelled in groups, corresponding to polarity dominance with a series pre-fix (for example, MI1 to MI7 for the Mississippian in the Carboniferous, Supplementary Fig. 1; GU1 to GU3n, LP1 to LP3, for the Guadalupian (Middle) and Lopingian (Upper Permian; Figs 2 and 3). The Permian Russian regional labelling scheme is also shown in Fig. 2. The late Cambrian–Ordovician magnetochrons are labelled mC (informal mid Cambrian), FU (Furongian), and LO, MO and UO (for Lower, Middle and Upper Ordovician, respectively, Fig. 4 and Supplementary Fig. 2). Data used in this compilation are all based on modern palaeomagnetic cleaning techniques, with good sampling density coverage, and data used are displayed with respect to section height, giving magnetostratigraphic data-quality values typically in excess of 7. The compositing procedure used here needs the largest possible subset of the stratigraphically most detailed and most reliable polarity data, with high sampling density, reasonable biostratigraphy and the largest stratigraphic coverage. Filtering the data set to that with the highest quality criteria would result in a lower resolution and less detailed polarity timescale. Some data interpretations have been slightly modified from original publications to maintain a consistency in polarity boundary interpretation between data sets.

**Russian Permian data.** In Russian sections, the end of the Kiaman superchron is in the upper Urzhumian\(^1\). Multiple sections, borehole cores and studies through the Kazanian and Lower Urzhumian have failed to detect any normal polarity intervals below the Russian magnetozone NRP; therefore, the end of the Kiaman Superchron is clearly expressed in these extensive data. The NRP Russian polarity interval shows two major reverse polarity intervals, the upper one of which is subdivided by a normal polarity submagnetozone. The structure of the lower reverse magnetozone in the NRP polarity interval is less clear and appears to be best represented by the Cherumuska section (type section of the Urzhumian), which is particularly thick. The uppermost normal polarity parts of magnetozone \( R_3P \) (that is, \( n_1R_3P \) and \( n_2R_3P \)) are missing from some sections, but...
are clearly present in the Oparino core and Boyevaya Gora section (and other sections not illustrated in Fig. 2).

Russian workers split the normal magnetozone sometimes seen straddling the Vyatkin–Vokhmian boundary into \(n_RnS\) and \(n_Sn\). The Vyatkin part is below a major increase in susceptibility and remanence intensity (because of a regional magnetite abundance increase) and a Vokhmian part above. The latest Permian magnetozones, LPn2 to LP3, are variably removed by erosion at the base of the Vokhmian, suggesting that magnetozone \(n_RnS\) is the equivalent of LPn2m (Figs 2 and 3).

The middle and upper Permian magnetostratigraphies from marine sections and Chinese sections display more detail in magnetozones over the Lopingian part of the timescale (Fig. 3) than the Russian non-marine sections, whereas the Russian platform sections show more detail in the Guadalupian (GU1 to GU3n intervals; Fig. 2). This is probably because of the absence of the mid and upper Changhsingian in many Russian sections.

Permian marine sections and Chinese sections. The middle and upper Permian magnetostratigraphy from marine sections and Chinese sections display the greatest detail and resolution in the Wuchiapingian and Changhsingian (that is, Abadeh, Linshui and Wulong sections, whereas the Capitanian and Wordian magnetostratigraphy is less well defined (Fig. 3). The end of the Kiaman Superchron is shown in the West Texas/New Mexico data, the Taiyuan section (Fig. 3) and in data from Japan. The magnetozone GU1n is the ‘Word-N’ magnetozone of ref. 26, which in the sections in the American south west appears to place it in the early Wordian.

Magenetostratigraphic data from Permian limestones in Japan show GU3n in the mid Capitaniain overlain by reverse polarity in the early Wuchiapingian, where their section-8 has an additional normal polarity magnetozone near the start of the Wuchiapingian, which is not clearly shown in other data. Their magnetostratigraphy through the Wordian contains both reverse and normal polarity intervals, with the earliest normal polarity magnetozone in the Neoschwagerina craticulifera fusulinid zone (their section-2) of the early Wordian, which may be the base of GU1n or GU1r.1n. The GU3n chron is the ‘Capitan-N’ magnetozone of ref. 26. The fragmentary (but well-dated) nature of the sections in Japan does not allow them to be used in the composite GPTS construction.

In spite of these studies on the Changhsingian at Meishan (the Changhsingian GSSP), the agreement on the magnetic polarity is poor. The only consistency between these studies suggests a lower normal polarity interval (the LP2n.1n?) and
mixed polarity in the younger part of the Changhsingian (LP2n.3n–LP3r polarity interval).

The age of the base of the Linshui section (Member 5 of the Lungtan Fm) is based on regional correlation of brachiopod assemblages, suggesting a late Wuchiapingian age, supported by the presence of the conodont Clarkina klanghanensis from the basal beds of the Lungtan Fm, 300 m below the measured magnetostratigraphy (personal communication, Shu-zhong Shen, 2010).

The Ebian county magnetostratigraphy through the Emeishan Basalts along with the Guadalupian conodonts from previous attempts is considerably from previous attempts by not using the Meishan section for scaling, and the smaller number of brief normal polarity intervals in the lower Wuchiapingian (Fig. 3). The most recent inclusive attempt is somewhat similar in the Wuchiapingian–Changhsingian, but differs greatly in relative thickness of magnetozones, especially for the Capitanian–Wordian. The most complete and well-documented transition into chron LT1n (which includes the Changhsingian–Induan boundary) is at the Shangsi section, where the chron can be tied to a succession of conodont zones.

**Carboniferous geomagnetic polarity data sources.** The most detailed magnetic polarity data are available for the Mississippian (mid-Carboniferous) from North America. Rather, more fragmentary magnetic polarity data occur in the early
Pennsylvanian, which however define the base of the Kiaman superchron (Supplementary Fig. 1) in the late Bashkirian (upper part of European Yeadonian regional stage\(^1\)). The base Kiaman is approximately at \(\text{ca. 318.8 Myr, derived from an array of radiometric dates through the Bashkirian and Moscovian from Russian sections, which can be related to the European stages using biostratigraphy\(^1\).} A gap of some \(\text{ca. 6 Myrs occurs between the late Mississippian and early Pennsylvanian magnetic polarity data sets; therefore, the magnetic polarity data are incomplete before the start of the Kiaman superchron}\(^1\).}

**Late Cambrian–Ordovician geomagnetic polarity data sources.** Palaeomagnetic polarity data for the Ordovician define the base of the Moyero superchron in the Tremadocian\(^1\). Magnetic polarity data from the underlying late Cambrian are known from

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**Figure 5** | Optimized reversal stratigraphy for the Permian and Middle-Upper Ordovician. (a,c) The base of the transformed corresponding magnetozone from each section is shown on the y axis, along with the median chron level (\(P_i\)) on the x axis and the resulting reversal stratigraphy in the top panel. (b,d) \(\sigma_T\)—the s.d. of the data for chron \(T_i\) values displayed on the y axis. The s.d. of \(T_i\) is scaled to Myr using the chron durations derived from Figs 6b and 7b (see Supplementary Tables 5 and 7), which is why the scatter in \(T_i\) values on the y axis (in a,c) does not correspond visually to the magnitude of \(\sigma_T\) in Myr in (b,d). Carboniferous (Mississippian) and late Cambrian-Lower Ordovician optimization data shown in Supplementary Fig. 4.
Siberia, Australia, and China (Supplementary Fig. 2). The key Siberian data set (from the Kulyumbe section) can be related to the well-dated (using conodonts) Australian data set by a set of carbon isotope-negative and -positive excursions in the late Cambrian. Attempts at extending the GPTS further into the mid-Cambrian using the Kulyumbe section data are hindered by an interval of low-quality data in the Orakta Fm.

The upper boundary of the Moyero superchron is in the early part of the Middle Ordovician. No detailed magnetic polarity data are available for the mid Katian or younger in the late Ordovician. The mid and late Ordovician magnetic polarities from Siberia (Fig. 4) are tied to the Siberian regional stages (Supplementary Table 4). Other single-sample data from two sections (Fig. 4).

In Siberia, early Baksanian strata contain the graptolite Oepikograptus bekkeri, which has been used as an analogy for the cosmopolitan early Sandbian, Nemagraptus gracilis Zone. (ref. 45) correlates the base of the Chertovskian to the base of the Sandbian, using co-occurring N. gracilis, and the regional equivalence of shelly faunas. This is similar to the proposed magnetostratigraphic correlation to the Gullăgen section (Fig. 4). The upper boundary of the Moyero superchron is in the early part of the Middle Ordovician. No detailed magnetic polarity data are available for the mid Katian or younger in the late Ordovician. The mid and late Ordovician magnetic polarities from Siberia (Fig. 4) are tied to the Siberian regional stages (Supplementary Table 4). Other single-sample data from two sections (Fig. 4).

The base of sub-magnetochron LO1r.1n is used as the end of the Moyero superchron, one of two brief submagnetozones in the Siberian Volginian and lower Kirenskian stages in the lower Rozhzhova, Polovinka and Moyero sections. Normal magnetochron MO1n.1n (Fig. 4), seen within the Kirenskian–Kudrinoan regional stages, is the equivalent magnetozone to that seen in the lowest part of the Holen Limestone in Sweden, which is dated to the Lenodus variabilis conodont zone (Fig. 4). This is correlated with the Da1–Da2 (Darriwillian) zonal boundary (Fig. 4). The MO1r interval is reverse polarity-dominated, with only the Gullăgen section spanning this entire interval. The chron MO2n is best represented in the Siberian Kudrino section, but inferred to be present in the upper Rozhzhova and Gullăgen sections. The high $\sigma_T$ values (Fig. 5d) indicate that there is much variability in magnetozone-relative thicknesses in the MO1r.1n to UO1n.1n interval. The Darriwillian–Sandbian boundary is well defined in the Mójca and Gullăgen sections within chron MO2r.2r (Fig. 4).

It has been suggested that the upper boundary of the Moyero superchron is at the base of MO1n; however, there is evidence for at least two brief normal chronos below this that are validated by single-sample data from two sections. Other single-sample intervals are not validated; therefore, the assumption used here is that the end of the Moyero Superchron is the base of LO1r.1n.

Radiometric age constraints. The Permian GPTS uses 20 U-Pb dates that can be directly related to the magnetostratigraphy (Fig. 6b), either within the sections or in sections that can be reliably tied to the sections with the magnetostratigraphy (Supplementary Table 2). Radiometric dates best constrain the Permian GPTS in the Changhsingian, and are sparse during the Wuchiapingian, Capitanian and Wordian (Fig. 6b). The composite Shangsi section data have a large number of associated U-Pb dates, making this section important for date-constraining the Lopingian magnetostratigraphy. The U-Pb radiometric ages from the Meishan section, using the EARTHTIME tracer calibration, are offset (ranging from $-0.087$ to $-0.157\%$) from the pre-EARTHTIME ages by some $-0.126\%$ on average (for beds 22, 25 and 28). This may bias the age calibration of the GPTS in the late Changhsingian; however, such offsets may not be systematic, compared with older pre-EARTHTIME calibrations, so the most recently published U-Pb dates are used (Supplementary Table 2).

The Permian age model indicates the good definition of the chron ages between 251 and 255 Myrs (Fig. 6b and...
However, the small number of radiometric dates available for the 257 to 266 Myrs interval produce larger 95% confidence intervals for chron ages older than 259 Myrs, which are some two to three times larger than those in the Changhsingian (Supplementary Table 5). The single Wordian radiometric date (Supplementary Table 2) has the biggest impact on the age model in the Guadalupian (Fig. 6b).

The Carboniferous age model has five radiometric dates (Supplementary Table 3), which constrain a linear relationship between the optimized scale and age (Fig. 6a). A sixth radiometric date (Supplementary Table 2) has the biggest impact on the age model in the Guadalupian (Fig. 6b).

Figure 8 | Smoothed reversal rates and 95% confidence intervals. Reversal rates determined using local regression techniques (a) Before the superchrons (0 Myr = superchron start) and (b) after the superchron (0 Myr = superchron end). Cretaceous data from ref. 63. Typical mid Cambrian reversal rates and Triassic rates shown as grey bars in a,b. Chron ordinal-age relationships, from which the reversal rates are derived, are shown in Supplementary Fig. 7. The three time intervals coloured accordingly. Confidence intervals use the point-wise method based on the ordinal-age gradient determination.

Figure 9 | Chron durations and chron patterns. Chron durations (a,b) and Shermans statistic (c,d), using a six-chron duration sliding window, showing classification of chron intervals into periodic (approximately equal duration), random and clustered (two groups of duration). The x axes are scaled in Myr from the older (a,c) and younger (b,d) boundaries of the three superchrons. Chron ages, durations and confidence intervals shown in Supplementary Tables 5–7.

Supplementary Fig. 3). However, the small number of radiometric dates available for the 257 to 266 Myrs interval produce larger 95% confidence intervals for chron ages older than ~259 Myrs, which are some two to three times larger than those in the Changhsingian (Supplementary Table 5). The single Wordian radiometric date (Supplementary Table 2) has the biggest impact on the age model in the Guadalupian (Fig. 6b).

The Carboniferous age model has five radiometric dates (Supplementary Table 3), which constrain a linear relationship between the optimized scale and age (Fig. 6a). A sixth radiometric date (Supplementary Table 2) has the biggest impact on the age model in the Guadalupian (Fig. 6b).
Figure 10 | Nomenclature used in the chron optimization method. Data from two hypothetical sections, with magnetozone boundary heights \( H_{j,i} \) with respect to the zero level of each section. Shown in red is the individual section \( E_{j,i} \), values \( E_{j} \) is the squared offsets, equation 1) used to measure the residuals from the optimized composite level of the fifth chron boundary \( P_{S} \). The medians of the magnetozone boundary levels (horizontal lines) give \( P_{S} \) for each section. Shown in blue are the positions of magnetozone boundaries that scale the two sets of \( H_{j,i} \) data to a nominal 0–1 scale. The transformation shift offsets for each section \( \beta_{j,i} \) serve to best align all the equivalent magnetozone boundary levels between sections. The actual heights in the two sections are also subject to the rate transformation \( f(\cdot) \) subject to height transformation \( T_{ij} = (H_{j,i}) + \beta_{j} \), producing transformed relative heights of \( T_{ij} \), which are used to determine the chron positions \( P_{S} \), in the optimization procedure. The example illustrated shows only linear rate transformations.

date from the mid Arnsburgian (Supplementary Table 3, date B9), if projected into this relationship (Fig. 6a), suggests that the magnetic polarity data ends in the earliest part of the Arnsburgian (late Serpukovian\(^1\)). Carboniferous optimization and confidence interval data are shown in Supplementary Figs 4a and 5.

In the mid–late Ordovician data, the only directly section-related radiometric date is from the Kinnekulle-K bentonite \( ^{40}\text{Ar}/^{39}\text{Ar} \) date of 458.0 ± 2.7 Myr, recalculated\(^1\)) at the top of the Dalby Limestone (Fig. 4). However, this part of the section has no magnetostratigraphy; therefore, its position with respect to the magnetostratigraphy cannot easily be inferred, hence is not used in the final data set to scale the polarity (Supplementary Table 4). Instead, CONOP-based age estimates of biostratigraphic boundaries (Table 20.1 in ref. 16) are used. The only radiometric date used is a \( ^{206}\text{Pb}/^{238}\text{U} \) date (456.9 ± 2.1 Myr) from strata in the upper part of the Amorphognathus tvaerenisi zone in Sweden, related to the magnetostratigraphy by the conodont zonations (but it has a large 2\( \sigma \) confidence interval of 2.1 Myrs; Fig. 7b). The Gullhögen section data produce very different scale to age relationships in the younger and older parts of the optimized composite (Fig. 7b), since it has thinner normal magnetozones over the MO1n and MO1r intervals, and is the only section that constrains the relative thickness of MO1r.1r. The age model uncertainties for the mid–late Ordovician GPTS show larger 95% confidence intervals on chron ages compared with those for the Carboniferous and Permian due to the paucity of age control (Supplementary Tables 5–7).

The late Cambrian and earliest Ordovician have a paucity of radiometric dates\(^5\), and only two can be directly related to the available magnetostratigraphy (Fig. 7a and Supplementary Table 4). These come with much larger analytical uncertainties than stratigraphic uncertainty. The proposed linear relationship between optimized scale and age is compatible with the base of the Guzhangan Stage at ca. 494.4 ± 3.5 Myr (ref. 52; Fig. 7a), an age relationship suggested by the carbon isotope data from the Kulyumbe section in Siberian (Supplementary Fig. 2).

Discussion

Initial post-Kiaman and post-Moyero polarity-reversal rates are \( \sim 2.5–4\) r/Myr—rates that are larger than those in the first 30 Myr following the CNPS (Fig. 8b). This shows that, unlike the CNPS, the reversal rates following the Palaeozoic superchrons have an initial fast restart. However, some 4–10 Myr following the Palaeozoic superchrons, reversal rates did decline and were comparable to the lower rates following the CNPS (Fig. 8b). Before all superchrons, reversal rates generally decline from higher values (some 10–15 Myr) before the superchron start, although the Carboniferous data are incomplete (Fig. 8a), and there is some cyclicity in the reversal rates. In the interval 7–15 Myr prior to the initiation of the three superchrons, reversal rates seem to get progressively larger back in time. The same may be the case for the 2 Myr prior to the superchron (Fig. 8a). If the heatflux change across the CMB provides the threshold for entering or leaving the non-reversing state (as suggested by geodynamo models\(^3\)), then the starting and ending of the Palaeozoic superchrons imply either a more rapid change in the heatflux or progressively higher heat flow threshold for older superchrons when entering or leaving the superchron state. This suggests that for geodynamo models to effectively model superchrons through the Phanerozoic, they need to include elements of slow progressive thermal evolution (perhaps via inner core growth or differing thermochemical piles in the lower mantle), as well as the interaction with mantle dynamics.

Some hypotheses of superchron start and end utilize support from the asymmetry in reversal rates either side of the CNPS, since this matches the modelled slow growth (superchron end, lower reversal rates) and faster collapse (superchron start, faster reversal rates) of superplumes\(^2\). Rate asymmetry can be considered over different time windows. For a ca. 2 Myr time window prior and following the CNPS and the Moyero superchrons, there is evidence for higher reversal rates at the superchron start (Fig. 8). The same can be inferred for a ca. 15 Myr time window either side of the CNPS and Moyero superchrons. Confirmation of this asymmetry for the Kiaman superchron requires more magnetic polarity data for the 8 Myr before the superchron start (Fig. 8a). However, these data do tentatively suggest that the superplume model may be a viable hypothesis for generating the low-frequency alternation between non-reversing and reversing states in the Palaeozoic.

The CNPS, Kiaman and Moyero superchrons are 42.3, 51.7 and 19.0 Myr in duration, respectively. The Permian chron GU3n is a long chron (\( \sim 4.3 \) Myr duration) like the Cretaceous C33r (3.7 Myr) and C33n (5.6 Myr), making these three chron the longest known in the Phanerozoic (apart from the superchrons;
Before the superchrons, there is also a small cluster of longduration chrons (M1n, M3r, FU1r, all > 1.5 Myr in duration) within ca. 7 Myr of the superchron start (Fig. 9a). In the Cambrian data, this is shown by the low reversal rates in the 3–7 Myr interval before the start of the Moyoyer superchron (Fig. 8a). These pre- and post-superchron data together suggest that there is a ca. 10 Myr window either side of the superchron in which there is an enhanced likelihood of an unusually long chron occurring, that is, a transitional state either side of the superchron. A speculative suggestion is that the post-superchron transitional state may also be evidenced in ‘memory-related’ relics of the previous superchron.

Methods

Construction of an optimized GPTS. The optimized composite comprises the relative height of magnetozone boundary, $P_i$, in a succession of $N$ magnetozone boundaries (that is, magnetozone boundaries $i = 1$ to $N$, for example, base of GU1n to base LT1n for the Permian), from the $N$ sections of data (Fig. 10). The height of magnetozone bases is used from each original section, $\delta_i$ ($i = 1$ to $N$ sections), magnetozone boundary $i$ will have relative height in that section of $H_i$. A measure of the mismatch, $E_i$, (Fig. 10), between the position of the magnetozone boundary in the optimized composite (that is, $P_i$) and the equivalent boundary in each section is $(P_i - H_i)$. Summing across all sections containing this magnetozone ($N_i$) gives the total mismatch of:

$$E_i = \sum_{j=1}^{N_i} (P_i - H_j)^2$$  \hspace{1cm} (1)

A section-dependent rate transformation $f(H_j)$ and height shift ($\beta_j$) was applied to the magnetozone heights in each section, giving new transformed heights $T_{ij}$, where $T_{ij} = f(H_j) + \beta_j$ (Fig. 10). This transformation is in effect a sedimentation rate function, and $\beta_j$ is a shift in the height applied to all magnetozones in that section. In hand-drawn composite these two factors correspond to a linear stretch (or shrinkage) of the magnetozone scale of each section, and a relative height shift that visually best matches the magnetozone anchor points used in the correlation diagrams. Using these transformations, and summing across all sections containing this magnetozone, gives the residual in the chron mismatch, $E_i$ expressed as:

$$E_i = \sum_{j=1}^{N_i} (P_i - f(H_j) - \beta_j)^2$$ \hspace{1cm} (2)

In practice, the number of magnetozone boundaries contributing to $E_i$ for each chron will not be the same throughout the composite. To remove this bias, the value of $E_i$ was normalized by the number of magnetozone boundaries ($N_i$) that contribute to $E_i$, which will be $\leq N_i$ therefore:

$$E_i = E_i / N_i \hspace{1cm} (3)$$

If there are $N_i$ total magnetozone boundaries in the optimized composite, then any expression that represents the sum of the mismatch across all sections and magnetozone boundaries is:

$$E_{tot} = \sum_{i=1}^{N_i} \left( \sum_{j=1}^{N_i} (P_i - f(H_j) - \beta_j)^2 \right) / N_i \hspace{1cm} (4)$$

The value of $P_i$ is some average (a median was used here, since there are small numbers of equivalent magnetozone boundaries) of the relative heights of the magnetozones $i$ in the $N_i$ sections, subject to the unknown rate transformation and height shift operations (that is, unknowns are $f(H_j)$ and $\beta_j$). For a perfect set of rate transformation and height shift values $E_{tot} = 0$. The unknowns for this system can be solved by numerical optimization, which attempts to minimize the value of $E_{tot}$. This also determines the minimum $E_i$ values for each magnetozone. The values determined by the optimization are the parameters of the rate transformation $f(H_j)$ and the constant $\beta_j$ for each section (that is, minimum of two unknowns per section, see below).

A second proxy (in addition to $E_i$) for the chron mismatch is the s.d. of $T_{ij}$ for each individual magnetozone (that is, $\sigma_i$; Fig. 5). $\sigma_i$ was scaled to Myr (from the final age model) to give an uncertainty estimate of the position of the chron.

Sedimentation rate transformations. Without independent evidence of sedimentation rates for sections, derived perhaps from sedimentological data, or other means of rate determination, a simple and widely used assumption is a constant sedimentation rate in each section (22). This assumption was used for most sections, since there is often insufficient information for most sections about the sedimentology or accompanying radiometric dates to properly evaluate any sedimentation rate changes. However, different rate functions were used on some sections, showing poorer model fits (Permian nonlinear ones shown in Supplementary Fig. 6). The additional functions used were either simulating increasing sedimentation rate upwards through the section (simulating transgressive sequences) or decreasing sedimentation rates (simulating regressive sequences). The transformation functions used were either constant, transgressive or regressive. For a constant rate, a transformation $T_{ij} = \gamma_i \times H_j + \beta_j$ (where $\gamma_i$ is in effect the sedimentation rate constant for each section, where $\gamma_i > 0$). For a transgressive rate, simulating smoothly increasing sedimentation rate, $T_{ij} = \beta_j + \gamma_i \times H_j$ (where $\gamma_i$ is the exponential distribution probability density function with $\lambda > 0$). For the regressive rate, simulating smoothly decreasing sedimentation rate, $T_{ij} = \beta_j + \gamma_i \times H_j$ (where $\gamma_i$ is the exponential distribution probability density function with $\lambda > 0$ and where $x_j > 0$).

Therefore, the unknowns were: which of the rate transformations to use and the values of $\gamma_i$, and $\beta_j$ (and $i$, if a non-constant rate function is used). Decisions on selecting rate functions depend on evaluating the magnetostratigraphic or geological data (helped by statistics outlined below). The $N$ sets of constants $\gamma_i, \beta_j$ (and potentially $\lambda$) are derived numerically by the optimization, giving up to a maximum of three unknown variables for each section.

Optimization of the rate constants. Minimization of $E_{tot}$ was performed using the Solver function in Microsoft Excel (Supplementary Fig. 8). However, to solve this two equality constraints are necessary to fix the vertical extent of the scale for $P_i$, an upper and lower limit (L, U). These limits were located at selected magnetozone boundaries in appropriate sections that show the upper and lower limits of the magnetic polarity pattern clearly (Supplementary Tables 8–10). For the Permian data these were the base of GU1n at Monastyrki and base of LT1n at Shangsi (Figs 2 and 3). For the late Ordovician these were the base of UO1n.2n and base of L01r.1n at the Rzhokova section (Fig. 4). For the Carboniferous and late Cambrian–Ordovician data sets, these are shown on Supplementary Figs 1 and 2. The L and U values in the appropriate spreadsheet cells are keyed into Solver as constraints on the optimization (Supplementary Notes).

The initial $H_i$ values were derived from Coreldraw correlation charts (Figs 2–4 and Supplementary Figs 1 and 2). The $H_i$ values were then re-scaled using $L$ and $U$ to produce new $H_i$ values ranging from $\sim 0$ to $\sim 1$ (see Supplementary Data). The optimization yields new transformed height values ($T_{ij}$) for each magnetozone boundary. In each section the median of the $T_i$ values (for each magnetozone boundary) was used as an estimate of $P_i$, the magnetozone position in the optimized composite, since there are relatively few data points per magnetozone (Supplementary Tables 8–11).

Goodness-of-fit statistics of the optimized composite GPTS. The residual between each magnetozone boundary in each section and the optimized composite is $T_{ij} - P_i$, and the average mismatch in each section per magnetozone is

$$\sum_{i=1}^{N_i} (T_{ij} - P_i)^2 / \left(1 / (N_i - 1) \right) \hspace{1cm} (5)$$

$N_i$ = number of magnetozone boundaries in section number $i$ ($N_i - 1$) in equation 5, since one more polarity boundary than magnetozones.

However, since the expectation is that the size of the mismatch is related to magnetozone duration (27), a better measure of relative (that is, between-section) mismatch is to normalize it by the average transformed chron height in that section. Therefore, a measure of the mean mismatch, $D_i$, between the optimized composite and the magnetozone boundary data in each section is:

$$D_i = \left( \sum_{i=1}^{N_i} (T_{ij} - P_i)^2 \right) / \left( N_i / (N_i - 1) \right) \hspace{1cm} (6)$$

where Ge is the geometric mean of the transformed chron height in the composite.

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over those chronos which occur in the section. The geometric mean, because chron durations (and so magnetozone-relative heights), vary enormously in relative scale. $D_i$ expresses the proportion of chron mismatch with respect to the mean magnetozone height in the section, and typically ranges from $-0.1$ to $0.3$ (Supplementary Tables 8–11), with larger values indicating worse matches. For these data sets $D_i$ greater than $0.25$ indicates a larger than usual misfit. In addition, across all sections

$$D_i = \frac{\sum N_i T_i - P_i}{N_i} = 0$$

$D_i$ provides a measure of the misfit of the entire set of section data. The data sets here have $D_i$ values of 0.07–0.28, indicating some 7–28% uncertainty in chron heights in the optimized composites.

For a perfect set of matching section data $T_i - P_i = 0$ for all chronos (that is, all chron boundaries in a study section would be perfectly aligned with the optimized composite). This difference is amenable to a paired t-test, which evaluates if the mean difference is statistically equivalent to zero, that is, $H_0$. Consequently, values of t-test probability, $TP < 0.05$, would suggest that the $T_i - P_i$ differences are biased away from zero, rather than being equally distributed about zero. The most likely reason for small $TP$ probabilities is that the rate transformation for that section is not well-matched to the optimized chron positions. In this situation, possible rate functions were investigated. Therefore, $D_i$ and $TP$ values allow identification of the most problematic section data—that with the largest $D_i$ and/or smallest $TP$ pointing to the most anomalous data set. Alternative rate transformations or correlation scenarios investigated were focussed on those section data highlighted by such values.

For each chron boundary, $E_i$, provides a measure of the chron misfit; therefore, larger $E_i$ values flag-up problematic boundaries in the optimized GPTS with a poorer fit. A rather more geologically meaningful parameter than $E_i$ expressing chron misfit is $\sigma_i$ (Fig. 5), but it is only generated at the age model stage.

Hiatus and data limits in sections. Additional information is contained in the stratigraphic extent of the last polarity magnetozones at the top and bottom of the sections, since the rate functions applied to the polarity boundaries also apply to these parts of the GPTS (Fig. 2). The top of the Sukhoto section would be expected to be below the base of the overlying normal chron (that is, $N_{R3P}$, which is $LP_{2n.3n}$) in the optimized GPTS. Likewise, the base of the Pizhma section should be above the base of the GU3n chron in the optimized composite. These acted as ‘top and base constraints’ in the numerical optimization procedures for other data sets is outlined in the Supplementary Information.

Rationale for optimization procedures. In GPTS optimization, the rate transformations start off with linear rate functions, which progressively evolved in some sections to nonlinear rate models, guided by the above statistics. For the Permian data, rate functions were initially assessed using the data in Figs 2 and 3 independently, and then applied to the whole Permian data set (as model P-1 in Supplementary Table 8 and finally P-4 in Supplementary Fig. 6). This allowed more consistency in regional data sets than between regions. Evaluation of $E_i$ and $D_i$ identified potentially anomalous sections allowing testing of new rate functions, with the Permian optimization models evolving to the final P-4 model (Supplementary Table 8). The best and simplest optimized GPTS models (lowest degrees of freedom) for the Carboniferous and mid-Ordovician data sets are using linear rate functions. Nonlinear rate functions were tested for these but gave little or no improvement in $D_i$—an overall rationale that attempted to minimize the $E_i$ and $D_i$ values while using the minimum degrees of freedom in the optimization (minimum in unknowns and acting constraints). The best late Cambrian–early Ordovician model uses only nonlinear rate functions. Guidance on using these procedures for other data sets is outlined in the Supplementary Information.

Radiometric dates, GPTS and reversal rates. The dates were attached to the optimized GPTS using the relative distance from magnetozone boundaries estimated from the individual sections (Supplementary Tables 2–4). The Permian GPTS is constructed from three piecemeal segments, two linear, in the older part, and a spline with generalized validation in the youngest part of the timescale (Fig. 6b). For the spline, the uncertainty (in Myr) on the U-Pb dates was weighted $(1/\sigma^2_i)$, with both the uncertainty ($\sigma_i$) in the U-Pb dates (including uncertainty in tracer calibration and $238U$H and the stratigraphic uncertainty ($\sigma_i$ in placing the U-Pb date on the magnetostratigraphy, an approach used in other timescales). $\pm \epsilon_i$, was estimated initially relative to a chron adjacent to the date position. A final $\pm \epsilon_i$ (in Myr) was determined by placing each U-Pb date on the optimized magnetostratigraphy scaled to Myr. For final spline segment fitting, $\epsilon_i$ was converted to $d_i$ by $\sigma_i = \sqrt{(d_i^2 + 12)}$ and the $\sigma_i$, used for weighting is given by $\sigma_i = \epsilon_i + a_i \epsilon_i$ (ref. 60). Linear age models used regression with errors in age ($\sigma_i$) and stratigraphic uncertainty ($\pm \epsilon_i$) in optimized level units (Fig. 6). The Carboniferous, mid-Ordovician and late Cambrian–early Ordovician GPTS used only linear age models between radiometric ages and the optimized scales. The uncertainty in the ages of magnetozone boundaries was determined by linear regression of each date against its estimated age, the procedure used in the geologic timescale giving a confidence interval for the ages ($CM_{3n}$), using the lower and upper 95% confidence bounds from the regression (Supplementary Tables 5–7). For the late Cambrian–Ordovician GPTS confidence intervals on chron ages could not be determined because of having only two age control points (Fig. 7a).

Polarity-reversal rates (Fig. 8) were determined by evaluating 1/gradient of the chron ordinal versus age relationship (Supplementary Fig. 7 and Supplementary Tables 5–7), using the method of local regression and likelihood as implemented in the LOCFIT routines in R. Local quadratic polynomials were used for smoothing, with the small window selection using generalized cross-validation, and CP statistics. In this, the weighting of the chron ages used were 1/$\sigma_i^2$ derived from the confidence interval $CM_{3n}$. The late Cambrian–early Ordovician GPTS reversal rates used 1/$\sigma_i^2$ for weighting since this age slice has no $CM_{3n}$ values. Cretaceous chron confidence interval estimates exist for the post-CNPS interval, but not for the pre-CNPS interval. Confidence intervals were estimated from the gradients and the reciprocal provided confidence intervals on the reversal rates.

Data availability. The author declares that the numerical data supporting the findings of this study are included in the Supplementary Information Data file, in Microsoft Excel format, along with literature sources of the primary data in Supplementary Table 1.

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