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Palaeosecular variation, field reversals and the stability of the geodynamo in the Precambrian

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SUMMARY

Palaeosecular variation (PSV), as estimated from the scatter of remanent magnetization directions or poles, can be used to shed light to processes in the geodynamo, and potentially, to model the growth of the solid inner core. To understand the temporal aspects of the geomagnetic field behavior in the far past, we have calculated the scatter of palaeomagnetic poles (S) from a set of 55 high-quality observations derived from a new Precambrian paleomagnetic data compilation. Our analysis supports former Phanerozoic and Precambrian analyses of PSV, which favour a lower reversal rate, a higher stability of the geodynamo and a substantially smaller size of the inner core prior to 1.5 billion years ago.

Key words: Magnetic mineralogy and petrology; Palaeomagnetic secular variation; Reversals: process, timescale, magnetostratigraphy.

1 INTRODUCTION

Palaeosecular variation (PSV) of the geomagnetic field accounts for changes observed in its intensity and the direction during stable polarity epochs, on timescales ranging from decades to millions of years. For explaining recently observed changes in the field, several global models of the field variation, such as CALS3K (Korte *et al.* 2009), CALS7K (Korte & Constable 2005) and GUFM1 (Jackson *et al.* 2000), have been developed. Each of these models can be used to reconstruct the geomagnetic field over terms of spherical harmonic coefficients at a given time in their validity period. In much longer timescales, particularly in the Precambrian, the geometry of the field cannot be described, though parametric PSV models (e.g. McFadden *et al.* 1988; Camps & Prévot 1996; Johnson *et al.* 2008; Smirnov *et al.* 2011) have been constructed for analyses of the broad-scale evolution of the field.

One of the various ways of applying PSV data is studying the stability of the geodynamo and its possible implications on the reversal rate of the field throughout the geological time. Among the several measures of PSV, the most amenable ones for this purpose are the dispersion of palaeomagnetic directions (s) (Irving & Ward 1964) and that of virtual geomagnetic poles (S) (Cox 1970). Plotting the S parameter with respect to palaeolatitude has proven that the field may possess time-varying non-dipolar features, particularly components of the symmetric family such as an axial quadrupolar component. Nonetheless, most of PSV studies (e.g. Tauxe & Kent 2004; Johnson *et al.* 2008; Harrison 2009) deal with the geomagnetic field of the last 5 Ma and it is unclear whether its features, such as the estimated ratio of dipolar and non-dipolar terms, prevail in longer timescales. An extension of this approach to the Precambrian data is possible, yet one must acknowledge the poor resolution of observations and the caveat that spatiotemporally limited data sets may give highly biased estimates of PSV.

One of the most interesting hypotheses derived from Precambrian PSV data casts doubt on the existence of the solid inner core of the Earth before 3.5 Ga, but supports a dipole-dominated field thereafter and the turning of the field to a more non-dipolar structure towards the Neoproterozoic (Smirnov *et al.* 2011). Based on their bistable dynamo model, Schmitt *et al.* (2001), demonstrated that the high latitudinal variation of VGP dispersion (S parameter), associated with low values near the equator, is typical of periods with low or moderate reversal frequency of the field as suggested by McFadden *et al.* (1991). This hypothesis gains support from the reversal chronology of last 165 Ma, where the Cretaceous Normal Superchron (118 to 83 Ma) is characterized by a dipole-dominated field (Aubert *et al.* 2010). According to Biggin *et al.* (2008a), the same phenomenon is visible even in the geomagnetic field of the late Archaean and early Proterozoic and may be associated with the smaller, or even nonexistent inner core (Aubert *et al.* 2009), consistent what Smirnov *et al.* (2011) and Biggin *et al.* (2008a) claim for their PSV data.

The correlation of PSV results with the field intensity has revealed an inverse relation between the reversal rate and the virtual dipole moment (VDM) in the last 320 Ma (Tarduno & Smirnov 2004). Due to the small number and high uncertainty of VDM measurements from oldest rocks, it remains questionable whether this analysis can be effectively extended to the Precambrian, and currently there is not enough evidence to reject the null hypothesis that the field intensity has been subject to a statistically significant change in the longest geological timescales (Dunlop & Yu 2004; Donadini 2007). The IAGA Global Paleointensity Database (Tauxe & Yamazaki 2007), now available at the website of the University of Liverpool (http://earth.liv.ac.uk/pint/index.htm) has faced some improvements, and a subsequent palaeointensity study has been performed on on Neoarchaean Australian rocks (Biggin *et al.* 2009), but there is still no escape from the fact that the current palaeointensity record of the Precambrian seriously lacks temporal coverage. The most notable gap is visible between 2.25 and 2.45 Ga, which led Ziegler & Stegman (2013) to suggest that this time interval may represent a transition between mantle and core dynamos, which left the Earth temporarily without a functioning geodynamo. However, PALEOMAGIA, our global online palaeomagnetic database (Veikkolainen *et al.* 2014a) comprises no less than 121 directional data from 2.25 to 2.45 Ga, for example the high-quality data sets from Dharwar dykes of India (Belica *et al.* 2014) and Widgiemooltha dykes of Australia (Smirnov *et al.* 2013). As the Dharwar data set has a well-defined U-Pb age and dual polarity paleomagnetic record, it necessitates at least one field reversal during that time. It is also evident that the 2.4 Ga Australian Erayinia dykes (Pisarevsky *et al.* 2014), which slightly postdate the Widgiemooltha swarm, have provided a field direction almost antiparallel to that of Widgiemooltha dykes. In the absence of the geodynamo this kind of reversal behaviour would be unlikely.

Since the reliability of Precambrian directional data generally overshadows that of intensity data, and dual-polarity palaeomagnetic data is prevalent throughout the Precambrian (Veikkolainen *et al.* 2014b), we have focused our analysis on the relation between PSV and the stability of the field via reversal rate (Coe & Glatzmaier 2006; Biggin *et al.* 2008a).

2 OBTAINING A MEASURE OF PSV

In paleomagnetism, Fisherian directions or VGPs are commonly assumed (Merrill *et al.* 1998; Deenen *et al.* 2011). The equation for the within-site scatter of directions (s) is simple and does not require any underlying assumption of the dipolar character of the field:

$$s = 81/\sqrt{k}, \tag{1}$$

The Fisherian precision parameter for the directional data, denoted by k, is used here. However, in terms of statistical validity, this approach is mathematically unjustified due to the latitude-dependent elongation of directional data which effectively disallows the assumption of Fisherian directions (Tauxe & Kent 2004; Deenen *et al.* 2011). Recent studies of PSV, especially those dealing with the lava flow data of last 5 Ma (McElhinny & McFadden 1997; Johnson *et al.* 2008; Kent *et al.* 2010; Cromwell *et al.* 2013a) systematically employ the scatter of VGPs, paying little or no attention to the scatter of directions, although in our study the scatter of directions is briefly discussed, too.

In a pure geocentric axial dipole field, the between-site scatter of VGPs (S_B) can be calculated as:

$$S_B = \left[\frac{1}{N-1}\sum_{i=1}^{N}\left(\Delta_I^2 - \frac{s_{Wi}}{n_i}\right)\right]^{1/2} \quad (i = 1 \ldots N), \tag{2}$$

Here N denotes the count of individual VGPs, Δ_i means the angular distance between the ith VGP and the mean of VGPs, and S_{Wi}/n_i is the within-site correction term (Cox 1969).

The total scatter S_T consists of between-site (S_B) and within-site (S_W) contributions and for within-site data, Fisherian distribution is commonly employed (Biggin *et al.* 2008a). Typically, large values of within-site scatter may result, for example from problems in the orientation of samples or contamination from secondary magnetizations. Although information on the internal structure of the geomagnetic field should ideally be wholly described by S_B, in practice the dispersion caused by the natural variation of the field and that resulting from experimental errors may be difficult to distinguish from one another. For example, in their study of Quaternary lava flows, Böhnel & Schnepp (1999) obtained 5.4° for the total S_W value, about 60 per cent (3.3°) of which was attributed to natural variation. They also suggested that S_w values of fresh lava flows follow log-normal distribution rather than a Fisherian one, though for different rock types this may not necessarily hold true. It is also true that even azimuthally symmetric sets are not necessarily Fisherian (Camps & Prévot 1996), but a proper investigation of this possibility in the Precambrian deserves a separate study to be performed.

A widely used way to describe the scatter of VGPs is to apply a parametric model, such as the Model G (McElhinny & McFadden 1997), which is based on the distinction between antisymmetric (e.g. axial dipole g_1^0 and octupole g_3^0) and symmetric (e.g. axial quadrupole g_2^0) Gauss coefficients (Roberts & Stix 1972). Following this approach, several fits to observational data have been published, for instance those by Smirnov *et al.* (2011), also based on Model G applied to more recent compilations. The increase of S_B from the equator towards high latitudes during this time is primarily caused by the fact that the axial dipole, the coefficient most strongly governing the evolution of the long-term field, belongs to the antisymmetric family. On the other hand, symmetric terms should produce scatter of poles independent of latitude (Merrill *et al.* 1998). Since Model G is parametric and does not derive from any simulations of the geodynamo, it does not provide a direct way to discriminate among individual terms within symmetric and antisymmetric families. Neither does it take the possible longitudinal dependence of PSV (Constable & Johnson 1999) into account, albeit the quantity and quality of data do not allow any tests of that hypothesis except for the last few million years.

Typical timescales of non-dipolar geomagnetic fields have been thought to vary in the range of hundreds to perhaps thousands of years (Hulot & Gallet 1996). In terms of PSV, recent evidence points to statistically distinct PSV patterns in the Brunhes and Matuyama epochs. Features associated with longer periods, such as 100 Ma (Tarduno *et al.* 2002; Johnson *et al.* 2008) may be a result of various factors, such as the changes in the non-dipole field, oscillations of the moment of the central dipole, dipole wobble or a combination of them (Brock 1971). In the studies of Precambrian, however, the hypothesis of the time-varying non-dipole field is convenient to implement as its influence can be directly attributed to changes of the relative variation of symmetric and antisymmetric terms of the field. The current technique of calculating PSV, as discussed by Merrill *et al.* (1998), applies the scatter of the poles (S) which is shown to be dependent of the absolute value of paleolatitude ($|\lambda|$), following the theoretical Model G (McElhinny & McFadden 1997). Both the lava flow data of the last 5 Ma (Quidelleur *et al.* 1994; Harrison 2009; Opdyke *et al.* 2010), and observations of the last 195 Ma show the same phenomenon (Tarduno *et al.* 2002) with small values of S near equator and larger values at high latitudes. However, Johnson *et al.* (2008) and Biggin *et al.* (2008b) noticed that the observed latitudinal dependence of S may be partly an artefact caused by the poor quality of observations and the limitations of the commonly applied way in correcting for within-site scatter.

In principle, PSV can be obtained from all rock types present in the novel online database of Precambrian palaeomagnetic data (Veikkolainen *et al.* 2014a), that is igneous, sedimentary and metamorphic rocks. Successive lava flows are the best source of data, since the typical timescale of their eruption covers the average secular variation timespan, roughly 10^3–10^5 yr (Tanaka *et al.* 1995; Laj *et al.* 1999; Smirnov *et al.* 2011). This must be short enough to allow the approximation that no apparent or true polar wander has taken place at the same time (Brock 1971), but long enough to avoid the undersampling of the field. Using sedimentary successions should be avoided, since the acquisition of their remanence may be delayed due to the complex lock-in processes, and the question whether the individual sample mean truly represents a spot reading of the geomagnetic field may be difficult to answer (Vigliotti 2006). In addition, the definition of an individual sampling site, which is crucial in calculating within-site dispersion, is indeterminate in many cases with just one sedimentary section sampled. Therefore, unaltered igneous rocks give most reliable estimates of PSV, and out of them, dykes, for example Biscotasing dyke swarm in Canada (Buchan *et al.* 1993; Halls & Davis 2004) and Dharwar dyke swarm in India (e.g. Halls *et al.* 2007; Piispa *et al.* 2011; Belica *et al.* 2014), and lavas, for example Portage Lake lava flow in Canada (Hnat *et al.* 2006), are the most useful in studies of the Precambrian PSV. Although proven to provide useful information to gauge PSV, they occur sporadically as pulses (Condie 1997), which are contrasted by periods with little or no igneous activity, leading to an uneven temporal resolution of the PSV record.

One of the problems in using Precambrian rocks as a source to obtain PSV is the fact that often the sampling sites, whether in dykes, lavas or sedimentary layers, record either apparent polar wander or transitional directions of polarity reversals, coupled with the ongoing PSV. For example, the lower reversed part (R2) of the Keweenawan (*ca.* 1.1 Ga) Mamainse Point lava section records rapid apparent polar wander with the resulting directions being systematically elongated along the apparent polar wander path (APWP) due to the rapid plate motion (Swanson-Hysell *et al.* 2009). This is unlike the situation in a typical secular variation case where the directions are either more spherically scattered around the overall mean (Pesonen 1979) or perpendicular against the APWP swathe (e.g. Donadini *et al.* 2009), thus lacking any elongation along the APWP. An attempt to calculate the S value in cases with an elongated pole pattern will lead to an erroneously high estimate as demonstrated, for example by lower reversed lavas of the Mamainse Point formation. Most Keweenawan studies, however, appear to have been conducted on rock units that span a short enough duration of time relative to APWP to be relatively unbiased by such motion and therefore the general pattern of S_B in Keweenawan (Halls & Pesonen 1982) is close to the model prediction for 1.0–2.2 Ga (Smirnov *et al.* 2011), with an overall increase of S towards higher latitudes. No elongation of poles along the APWP is visible, for example in the VGP pattern of the stratigraphically thick Portage Lake Volcanics, aged 1092–1098 Ma and with normal polarity only (Hnat *et al.* 2006; Donadini 2007). The Umkondo-age dolerites (Gose *et al.* 2006), which are coeval with Lake Superior results, have a relatively shallow value of S_B ($13.1^\circ \pm 0.5^\circ$), as predicted for nearly equatorial results. Similarly, reversed-polarity Central Arizona diabases (Elston 1989; Harlan 1993; Donadini *et al.* 2009), which occupied nearly polar latitudes at 1.1 Ga, yield larger S_B ($28.4^\circ \pm 2.1^\circ$), as expected.

3 MODELLING AND RESULTS

In our PSV analysis, particular attention was paid to the scatter of VGPs and the comparison of our novel data set with the Model G of PSV and its Precambrian counterparts (McFadden *et al.* 1991; Smirnov *et al.* 2011):

$$S^2 = (a\lambda)^2 + b^2. \tag{3}$$

In eq. (3), the dipolar (antisymmetric) or odd family is denoted by a, and the quadrupolar or even family by b. The original a and b parameters in the Model G had been calculated on the grounds of 3719 lava flow observations from the period of 0–5 Ma, leading to $a = 0.26$ and $b = 11.9$, whereas the more recent fit by Smirnov *et al.* (2011) had $a = 0.25$ and $b = 13.2$, corresponding to a slightly larger quadrupolar contribution than previously concluded. These values, however, tell little about the strengths of individual spherical harmonic terms, but rather, describe the overall temporal variation of all even and odd terms combined. Knowledge of the underlying physical processes is dubious, yet it is unlikely that these two types of families are truly independent of one another. It must be pointed out here that the value of S for any standing spherical harmonic term alone is zero by definition, and even the influence of the temporally averaged field on the VGP scatter curves may not be significant (Hulot & Gallet 1996). However, the relative strengths of b and a may give implications on the stability of the geodynamo, with larger b/a ratios pointing to a more unstable field with a higher proportion of excursional directions and reversals (Coe & Glatzmaier 2006). Therefore PSV data can be fruitfully applied to studies of the long-term stability of the time-averaged geomagnetic field.

Secular variation is generally determined for a stable-polarity field, or sometimes as temporally averaged using data before and after a field reversal or excursions, and VGPs of transitional data are not considered. They show high values of scatter, which typically do not occur during stable-polarity periods. Several values for the cutoff angle for excluding the low-latitude VGPs have been applied, ranging from 35° (Quidelleur *et al.* 1994) to 45° (McElhinny & McFadden 1997). Both variable (Vandamme 1994) and constant (Johnson *et al.* 2008) cutoff criteria have been previously used, but we applied neither, since the VGPs allowed by the authors for the calculation of the palaeomagnetic poles were in practically all cases clustered within 45° from the mean pole position. Whenever applied, cut-off invalidates the assumption of Fisherian-distributed VGPs and may even remove scatter actually caused by PSV, thus causing the calculated values of scatter to be systematically smaller (Lawrence *et al.* 2006; Cromwell *et al.* 2013b). We acknowledge the fact that in sedimentary sequences, such as the 1.88 Ga Stark Formation of the Slave craton, Canada (Bingham & Evans 1976) and the late Mesoproterozoic Siberian strata (Gallet *et al.* 2000), transitional poles may be frequent, but in our igneous-only compilation this problem was virtually nonexistent.

To estimate whether it is reasonable to use a combination of N and R data from the same rock unit in studying PSV, one may need to consult the reversals test (McFadden & McElhinny 1990), since the improper use of statistically different N and R subsets together may lead to an anomalously high values of scatter. Along with this criterion, our analysis follows the convention that a combined entry for dual-polarity data has been used only in cases where N and R records are roughly of the same age and pass the reversals test, for example for the data of Mashonaland sills (Bates & Jones 1996), which were also included in the data set of Smirnov *et al.* (2011) as a combined entry. Conversely, precise U-Pb datings of Matachewan and Marathon dykes (e.g. Buchan *et al.* 1996; Halls *et al.* 2008; Evans & Halls 2010) have revealed demonstrably different ages for N and R polarities. These are thus treated independently in our analysis, with the block rotation corrected if necessary. For the 2.17 Ga Biscotasing dyke data, no rotation parameters were used, but separate S_B values were calculated for western (Halls & Davis 2004) and eastern (Buchan *et al.* 1993) blocks.

Our study applies the selection of data following the MV (Modified Van der Voo) ≥3 quality criterion (Van der Voo 1990; Veikkolainen *et al.* 2014c) on igneous rocks only, preferring mafic and intermediate extrusive rocks, flat-lying intrusive rocks and nearly vertical dykes with well-defined isotopic age information and evidence of a primary magnetization, preferably carried by magnetite instead of hematite. For instance, in the North Shore traps (Tauxe & Kodama 2009) both magnetite and hematitic remanence data were available, but only magnetite records were used for the analysis as they are less likely to have been affected by low-temperature alteration. Data with good structural coherence, and hence no need for tilt correction of directional data, are essential, since incorrectly applied tilt corrections can cause erroneous VGP scatter estimates, yet in the data of Portage Lake volcanics (Hnat *et al.* 2006) and Purcell lava (Elston *et al.* 2002), there is no escape from the fact that the effect of post-emplacement tilting needs to be removed before considering the directional data robust. No data from sediments, metamorphic rocks and slowly cooled intrusive rocks, for example plutons, were included in the compilation. In general, it was required that the original research article provided adequate sample-mean or specimen-mean statistics (α_{95} or k) for the proper determination of the within-site scatter of poles (S_{w}), although in a few cases, including the Portage Lake volcanics (Hnat *et al.* 2006), these were obtained via personal communication. Further requirements included the presence of at least nine sampling sites and 30 samples after the removal of sites without adequate statistical information. Typically the sites rejected from the analysis included not more than one or two samples, or were already excluded by the authors as having anomalously large error parameters. Moreover, sites with statistics determined via demagnetization circles only were not considered as they do not allow the calculation of within-site scatter in a straightforward way.

Our final set of 55 Precambrian entries (Table 1) possessed an average quality rating MV = 4.9 as opposed to the average rating (MV = 2.9) in the PALEOMAGIA database as a whole. For comparison, Smirnov *et al.* (2011) previously used 23 points of data in their analysis, with an average rating MV = 5.0. More than half of the data used in our compilation ($N = 31$) were derived from the present-day North America, including Greenland. Isotopic ages were available for most rocks observed, with the majority (69 per cent) of results having uranium-lead age as the primary method of dating. The quantity of sampling sites used varied highly: from the reversed-polarity Matachewan dykes (Evans & Halls 2010) as much as 123 sites were included in the analysis, as opposed to our minimum requirement (nine sites), which was barely met, for example in the case of Malley dykes (Buchan *et al.* 2012).

Repeated studies on the same rock unit should ideally produce nearly identical values of S, and applying the $n-1$ jackknife method (Efron 1982) to calculate error parameters, one may study whether the data sets are statistically similar or distinct. For example, studies on the well-defined 2.37 Ga magnetization of Dharwar dykes (Halls *et al.* 2007; Piispa *et al.* 2011) can be used to test this presumption. In the former study, seven normal- and one reversed-polarity site give $S_{\mathrm{B}} = 18.5^{\circ} \pm 0.3^{\circ}$, and in the latter one, the corresponding value is $S_{\mathrm{B}} = 22.1^{\circ} \pm 2.9^{\circ}$ as derived from five sites, since one of the originally included sites was later proven to be of different age than others (Belica *et al.* (2014). While the data set used by Piispa *et al.* (2011) is too small for a robust estimate of PSV, the S_{B} estimate of Halls *et al.* (2007) evidently falls into the error limits of the new corresponding value derived from the large and well-constrained compilation of Belica *et al.* (2014), with $S_{\mathrm{B}} = 18.4^{\circ} \pm 0.3^{\circ}$. In all these cases, error parameters correspond to two standard deviations (2σ) from the mean, and this is the standard procedure in our paper.

Several previous studies deal with the relation between the geologic age and the behaviour of PSV (Halls & Pesonen 1982; Biggin *et al.* 2008a; Smirnov *et al.* 2011). In our study, the data were divided into two temporal intervals, with the first one ranging from 500 to 1500 Ma ($N = 28$) and the second one from 1500 to 2900 Ma ($N = 27$). Although the relation of inclination I and palaeolatitude λ in the GAD field ($\tan I = 2 \tan \lambda$), predicts a small proportion of observations at high palaeolatitudes, our compilation includes not more than 4 records (7.3 per cent of all data) with $|\lambda|$ greater than 60° which is fewer than the expected percentage (13.4 per cent). Smirnov *et al.* (2011), on the other hand, drew conclusions from a data set with just one entry (Dharwar dykes) at high palaeolatitudes ($|\lambda| > 60^{\circ}$) and even this entry (Halls *et al.* 2007) is replaced in our study by a newer and more reliable one (Belica *et al.* 2014). Another high-latitude palaeomagnetic record, namely that of the 1192 ± 10 Ma Harohalli alkaline dykes (Dawson & Hargraves (1994), Radhakrishna & Mathew (1996), Pradhan *et al.* 2008), with $|\lambda| = 77.4^{\circ}$ and $S_{\mathrm{B}} = 20.4^{\circ} \pm 5.5^{\circ}$ has fairly large error limits, yet not greater than a number of entries included by Smirnov *et al.* (2011) in their analysis have. Unfortunately, we could not include the steep directions derived from 930 Ma dolerites of southern Sweden (e.g. Bylund 1992; Pisarevsky & Bylund 1998) for our analysis since those most likely represent uplift-cooling magnetizations despite revealing field reversals also within dykes.

Being aware of the persistent uncertainty of our data at high palaeolatitudes, we fitted the model parameters a and b first for the entire Precambrian (Fig. 1), and later for the Mesoarchaean-Mesoproterozoic and Meso-Neoproterozoic subsets separately (Figs 2 and 3). The fitting was done using a Python script applying nonlinear iterative least squares method based on the Levenberg–Marquardt algorithm (Pujol 2007). Although we were not able to get all within-site statistical data used by Smirnov *et al.* (2011) for comparison, a large proportion of their entries seem to have a very high degree of uncertainty which leads us to suspect their way of calculating error parameters. For example, their result for the moderate-latitude ($|\lambda| = 43.2^{\circ}$) 2.66–2.71 Ga Allanridge lavas (Strik *et al.* 2007; De Kock *et al.* 2009) yielded $S_{\mathrm{B}} = 12.1^{\circ} \pm 5.7^{\circ}$ and that for the low-latitude ($|\lambda| = 7.7^{\circ}$) 2.46 Ga Matachewan *R* dykes yielded $S_{\mathrm{B}} = 8.8^{\circ} \pm 5.1^{\circ}$, even though their estimates are based on one standard deviation from the mean unlike our estimates which apply two standard deviations. Our Matachewan *R* data yielded $S_{\mathrm{B}} = 9.8^{\circ} \pm 0.1^{\circ}$, a result based on 123 sites and well within the large error limits provided by Smirnov *et al.* (2011). The regionally filtered Matachewan data sets used by Biggin *et al.* (2008a) have a few degrees of statistical uncertainty, most probably because none of them has more than 14 sampling sites. This also demonstrates the fact that for well-behaving populations, the statistical error inherent in the jackknife method decreases with the increasing number of entries, as expected.

Using the non-parametric sign test, we estimated whether the Model G curve fitted for the TAFI data of last 5 Ma (Smirnov *et al.* 2011) serves as a reasonable proxy for our Precambrian records. In our pan-Precambrian data set ($N = 55$), as much as 49 entries fell below the model curve, corresponding to a statistical probability (two-tailed p value) of less than 0.0001. However, in the 0.5–1.5 Ga subset, 5 out of 28 entries show higher S_{B} values than predicted from the TAFI model fit, corresponding to two-tailed p value of 0.0009. In 1.5–2.9 Ga data, all entries except that of the Satakunta Subjotnian dykes (Salminen *et al.* 2014) show lower values than expected from the model at similar palaeolatitudes. Since the age of magnetization of Satakunta dykes is 1565 ± 35 Ma, it is the youngest entry in the older subset. The functionality of Model G seems to decrease with geologic time, which is in line with the conclusion of Smirnov *et al.* (2011) based on their comparison of Precambrian data with TAFI records. In the Precambrian, large differences between the

Table 1. Summary of Precambrian PSV data used for this study. Ages are in Ma. Here B/N means how many sites and samples were used and S(B) stands for the within-site scatter (°) along with error parameters corresponding to two standard deviations from the mean. Mean paleolatitudes (°), MV quality grades (Van der Voo 1990; Veikkolainen *et al.* 2014c) and references to the original studies are also given. For a more detailed description of the data, the reader is referred to Supporting Information Appendices A and B.

Rock unit	Age	B/N	S(B)	$\lvert\lambda\rvert$	MV	References
Fortesque Package 0	2822	24/127	16.1 ± 0.6	56.0	5	Biggin *et al.* (2008a)
Modipe gabbro – N+R	2784	11/56	15.4 ± 1.5	64.5	5	Denyszyn *et al.* (2013), Evans & McElhinny (1966)
Pilbara Packages 1–2, 6–7	2747	46/278	17.7 ± 0.6	51.0	5	Biggin *et al.* (2008a)
Nyanzian lavas – N+R	2680	11/68	6.2 ± 0.8	25.2	6	Meert *et al.* (1994)
Allanridge basalts	2675	20/165	14.4 ± 1.0	43.6	5	Strik *et al.* (2007), De Kock *et al.* (2009)
Matachewan dykes – R	2460	123/687	9.8 ± 0.1	8.8	4	Evans & Halls (2010), Bates & Halls (1990, 1991), Vandall & Symons (1990), Buchan *et al.* (1990,1996), Smirnov & Tarduno (2004), Halls *et al.* (2005), Halls & Shaw (1988), Irving & Naldrett (1977), Pesonen (1973), Aibangbee (1982)
Karelian dykes	2458	10/56	9.6 ± 1.0	27.0	4	Mertanen *et al.* (1999)
Matachewan dykes –N	2446	63/322	7.9 ± 0.2	13.9	4	Evans & Halls (2010), Bates & Halls (1990, 1991), Smirnov & Tarduno (2004), Halls & Shaw (1988), Irving & Naldrett (1977)
Widgiemooltha dyke suite	2415	19/137	11.2 ± 0.7	48.9	5	Smirnov *et al.* (2013), Evans (1968)
Dharwar dykes A – N+R	2367	57/404	18.4 ± 0.3	72.2	6	Belica *et al.* (2014), Halls *et al.* (2007), Piispa *et al.* (2011), Kumar and Bhalla (1983), Radhakrishna *et al.* (2013a,b), Dash *et al.* (2013), Bhalla *et al.* (1980), Radhakrishna & Joseph (1996), Venkatesh *et al.* (1987), Dawson & Hargraves (1994)
Malley dykes	2231	9/45	9.8 ± 1.8	34.3	5	Buchan *et al.* (2012)
Ongeluk lavas	2222	18/103	8.5 ± 1.1	12.3	6	Evans *et al.* (1997)
Senneterre dykes	2216	12/55	6.6 ± 1.2	25.5	6	Buchan *et al.* (1993)
Tulemalu dykes – R	2190	12/68	10.8 ± 1.4	23.0	5	Fahrig *et al.* (1984)
Biscotasing East dykes	2169	15/65	11.9 ± 1.1	42.7	5	Buchan *et al.* (1993)
Biscotasing West dykes	2169	9/72	9.5 ± 1.8	43.1	5	Halls & Davis (2004)
Marathon dykes – N	2124	18/133	16.2 ± 0.9	39.7	5	Halls *et al.* (2008), Buchan *et al.* (1996)
Marathon dykes – R	2106	13/70	13.8 ± 1.0	34.4	5	Halls *et al.* (2008), Buchan *et al.* (1996)
Lac de Gras dykes	2026	10/49	11.2 ± 2.1	35.8	5	Buchan *et al.* (2009)
Uauá dykes	1983	20/68	16.8 ± 1.3	56.2	3	D'Agrella-Filho & Pacca (1998)
Bundelkhand NW–SE dykes	1979	10/137	11.4 ± 1.1	6.0	5	Pradhan *et al.* (2012)
Mashonaland dolerites – N+R	1878	16/78	14.4 ± 1.8	28.8	6	Bates & Jones (1996)
Taihang NNW trending dyke swarm	1769	19/125	8.5 ± 1.2	2.6	5	Halls *et al.* (2000)
Cleaver dykes	1741	17/99	14.6 ± 1.0	39.1	5	Irving *et al.* (2004)
Melville Bugt diabase – N+R	1625	9/54	13.0 ± 2.0	16.9	5	Halls *et al.* (2011)
Åland dykes –R	1576	23/154	6.2 ± 0.8	10.2	5	Pesonen, L.J.
Satakunta Subjotnian N–S and NE–SW dykes	1565	18/112	14.6 ± 1.0	2.3	6	Salminen *et al.* (2014)
Purcell lava	1443	10/166	7.3 ± 0.7	13.5	5	Elston *et al.* (2002)
Nova Guarita dykes – R	1419	17/197	16.1 ± 1.2	27.6	6	Bispo-Santos *et al.* (2012)
Midsommersø dolerites	1382	10/100	10.3 ± 1.7	7.4	4	Marcussen & Abrahamsen (1983)
Zig-Zag Dal basalts	1382	17/131	8.1 ± 0.7	10.3	4	Marcussen & Abrahamsen (1983)
Yanliao mafic sills	1323	18/142	9.7 ± 1.0	17.2	5	Chen *et al.* (2013)
Väster-Norrland dolerites	1257	43/252	7.9 ± 0.2	27.0	4	Piper (1979)
Sudbury dykes	1237	37/146	10.8 ± 0.4	0.9	5	Palmer *et al.* (1977)
Nova Floresta formation	1200	16/115	9.0 ± 0.8	41.2	4	Tohver *et al.* (2002)
Harohalli alkaline dykes	1192	10/44	20.4 ± 5.5	77.4	5	Pradhan *et al.* (2008), Dawson & Hargraves (1994), Radhakrishna & Mathew (1996)
Gardar NE–SW dyke swarms	1172	18/102	12.5 ± 1.0	31.6	4	Piper (1992)
Giant gabbro dykes of Tugtutôg	1164	12/68	15.3 ± 1.3	38.0	4	Piper (1977)
Thunder Bay dykes – R	1120	19/94	15.0 ± 1.8	51.8	5	Pesonen (1979)
Central Arizona diabases – R	1116	9/87	28.4 ± 2.1	77.5	4	Donadini *et al.* (2009), Harlan (1993), Elston (1989)
Lowermost Mamainse lavas – N2	1110	11/54	15.3 ± 1.2	30.6	5	Swanson-Hysell *et al.* (2009)
Umkondo dolerites – R	1110	27/247	13.0 ± 0.6	4.5	6	Gose *et al.* (2006), Hanson *et al.* (2004), Seidel (2004), Pancake (2001), Jones & McElhinny (1966)

Table 1 *(Continued.)*

Rock unit	Age	B/N	S(B)	|λ|	MV	References
Lower Mamainse lavas – R2	1108	14/72	15.3 ± 0.9	57.0	5	Swanson-Hysell *et al.* (2009)
Powder Mill volcanics – R	1107	29/180	19.0 ± 0.9	50.8	4	Palmer & Halls (1986)
Uppermost Mamainse lavas – R1	1101	10/52	11.4 ± 1.6	26.3	5	Swanson-Hysell *et al.* (2009)
North Shore traps	1099	33/150	11.1 ± 0.4	27.7	4	Tauxe & Kodama (2009)
Portage lake curved lavas	1095	28/195	13.7 ± 0.5	16.9	5	Hnat *et al.* (2006)
Thunder Bay dykes – N	1094	19/70	11.2 ± 0.6	21.3	5	Pesonen (1979)
Central Arizona diabases – N	1090	29/284	16.7 ± 0.8	26.3	4	Donadini *et al.* (2009), Harlan (1993), Helsley & Spall (1972)
Lake Shore traps	1087	31/354	12.3 ± 0.3	19.0	5	Kulakov *et al.* (2013), Diehl & Haig (1994)
Michipicoten Island formation	1087	13/58	12.5 ± 0.9	12.4	4	Palmer & Davis (1987)
Alcurra dykes and sills	1077	9/47	10.5 ± 1.6	31.5	5	Schmidt *et al.* (2006)
Franklin dykes – N+R	723	12/80	7.1 ± 0.4	0.5	6	Christie & Fahrig (1983)
Dokhan volcanic formation	596	10/49	12.5 ± 1.5	20.5	4	Nairn *et al.* (1987)
Skinner Cove volcanics	551	10/57	13.5 ± 1.4	17.0	5	McCausland & Hodych (1998)

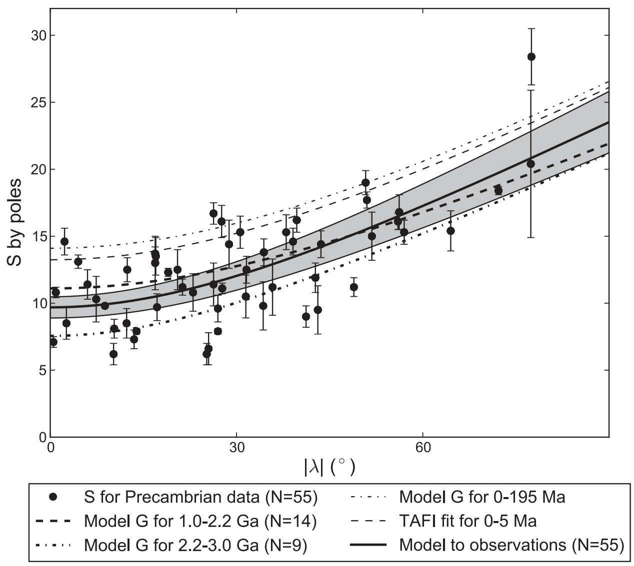

Figure 1. The observed dispersion of virtual geomagnetic poles (S parameter) by the absolute value of paleolatitude (|λ|) for our within-site corrected Precambrian paleomagnetic data set and a corresponding Model G fit ($N = 55$, shown as 'Model to observations'). Each entry is accompanied by its error limits corresponding to 2σ (two standard deviations from the mean). The best statistical fit to our compilation along with its error envelope is also shown. Parametric PSV models of S for timeslots of 0–5 Ma (TAFI fit; $b = 0.25$; $a = 13.24$), 1.0–2.2 Ga ($b = 0.21$; $a = 11.10$) and 2.2–3.0 Ga ($b = 0.22$; $a = 7.56$) by Smirnov *et al.* (2011) and the Model G fit to 5–195 Ma data (Biggin *et al.* 2008a) are also shown. For data used to produce this plot, see Tables 1 and 2 and Supporting Information Appendices A and B.

model and observations are most prominently visible at shallow palaeolatitudes, leading to a readjustment of the equatorial intercept (term b in eq. 3) to a smaller value, while the slope of the curve (term a) seems to remain close to the corresponding TAFI value. For all of our fits, consistent calculation of lower and upper error limits was possible by subtraction, and in turn, addition of jackknife error estimates from their respective *S* parameter values and thus producing new data sets with lowest and highest possible S values.

4 RESULTS AND CONCLUSIONS

The morphology of the geomagnetic field during the Precambrian and the evolution of the solid inner core have been a matter of controversy (Bloxham 2000; Aubert *et al.* 2009; Smirnov *et al.* 2011; Pozzo *et al.* 2012). In a PSV study of Precambrian igneous rocks, the maximum age for the amalgamation of the Earth's solid core was estimated to be 3.5 Ga, meaning that no permanent geodynamo was existent before that (Smirnov *et al.* 2011). Discussion on the earliest evidence of functioning geodynamo has included palaeomagnetic data from the 3.48 Ga Barberton komatiites of South Africa (Yoshihara & Hamano 2004) and from the 3.46 Ga Marble Bar Chert of Pilbara, Australia (Hale 1987; Suganuma *et al.* 2006), but only the Barberton data has a confirmed Paleoarchaean age as proven by Tarduno *et al.* (2010) using the conglomerate test presented by Usui *et al.* (2009).

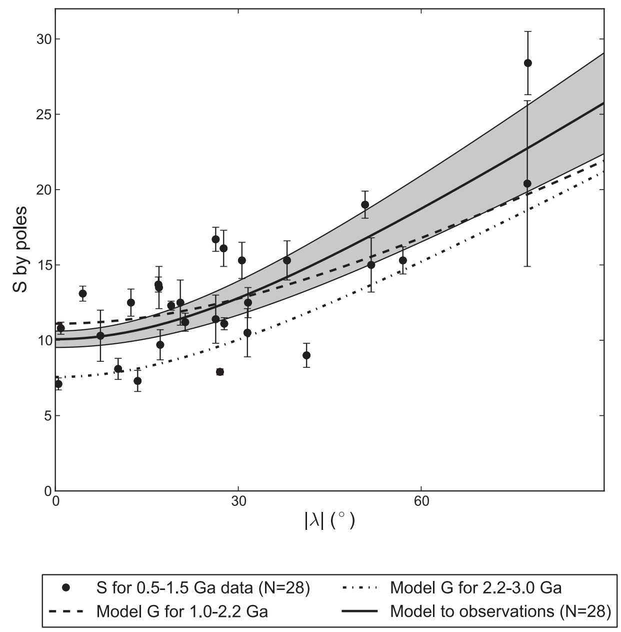

Figure 2. The observed dispersion of virtual geomagnetic poles (S parameter) by the absolute value of paleolatitude (|λ|) for our within-site corrected Meso-Neoproterozoic paleomagnetic data set and a corresponding Model G fit (0.5–1.5 Ga, $N = 55$, shown as 'Model to observations'). Unlike in Fig. 1, TAFI fit and 5–195 Ma fit are not visible, but 1.0–2.2 and 2.2–3.0 Ga fits by Smirnov *et al.* (2011) are plotted.

According to Smirnov *et al.* (2011), the Neoarchaean and the Palaeoproterozoic were characterized by a strongly dipolar field, which transformed to a less dipolar state *ca.* 2 Ga ago and prevailed as such till the end of the Mesoproterozoic. However, this hypothesis is weakened by the high uncertainty in the 1.0–2.2 Ga data

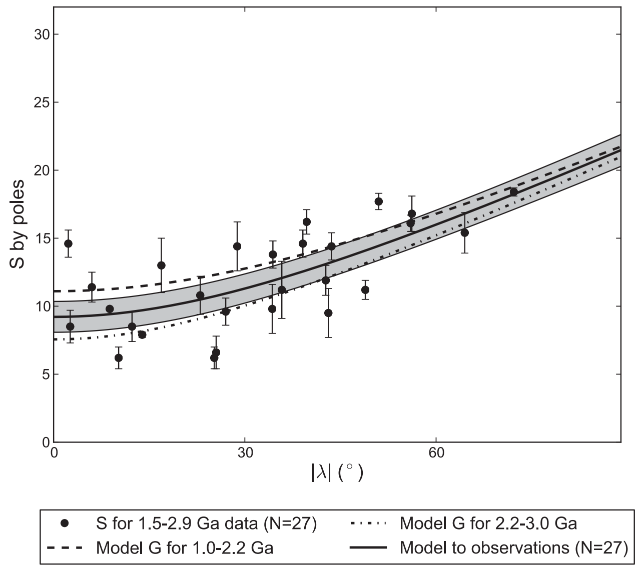

Figure 3. The observed dispersion of virtual geomagnetic poles (S parameter) by the absolute value of paleolatitude (|λ|) for our within-site corrected Mesoarchaean–Mesoproterozoic paleomagnetic data set and a corresponding Model G fit (1.5–2.9 Ga, $N = 27$, shown as 'Model to observations'). Unlike in Fig. 1, TAFI fit and 5–195 Ma fits are not visible, but 1.0–2.2 and 2.2–3.0 Ga fits by Smirnov *et al.* (2011) are plotted.

applied by Smirnov *et al.* (2011), which is most prominently visible in their estimate of the antisymmetric field contribution (term a) of Model G. A substantial effort in analyzing data from 2.82 to 2.45 Ga African, Australian and North American rocks was made by Biggin *et al.* (2008a) who also argued that the geomagnetic field switched its polarity less frequently than in the Phanerozoic and the deviation of the long-term field from the axial symmetry was small. This conclusion supported the originally unexpected result of the geodynamo simulation of Roberts & Glatzmaier (2001), who found that a large inner core triggers instability to the fluid motion in the outer core. The possibility of radioactive matter in the core, notably potassium, has been suggested since the present-day heat flow rates at the core–mantle boundary do not allow a primordial geodynamo to have taken place, but an additional, physically controversial heat source must have been operating (Rama Murthy *et al.* 2003; Pozzo *et al.* 2012).

Due to the potential bias caused by poorly constrained ages, overprint magnetizations and tectonic alteration, it remains obvious that Precambrian observations on average provide less reliable PSV estimates than Phanerozoic ones (Table 2), but still, the Model G fit to our 1.5–2.9 Ga data (Fig. 3) yields values of 9.21 ± 1.14 for *b* and 0.22 ± 0.02 for a, leading to a curve which falls below the corresponding 0–5 Ma model (Smirnov *et al.* 2011) and even more substantially below the 5–195 Ma curve (Biggin *et al.* 2008a). However, the actual difference of Precambrian and 5–195 Ma G curves may be smaller than that visible in Fig. 1, since Biggin *et al.* (2008a) applied a constant value to correct site-level data, a possible source of error especially at shallow palaeolatitudes. Applying latitudinally dependent within-site correction to the data of Biggin *et al.* (2008a) would provide a solution to the problem.

Model G fits for our 0.5–1.5 Ga subset ($b = 10.07 \pm 0.54$ and $a = 0.26 \pm 0.04$, Fig. 2) and for our entire Precambrian data compilation ($b = 9.67 \pm 0.79$ and $a = 0.24 \pm 0.03$, Fig. 1) seemed to have a much closer similarity with each other, than the fits for our 1.5–2.9 Ga subset and Phanerozoic data sets had. However, when using sign test in trying to analyse whether the 0.5–1.5 Ga model fit can be used to explain our 1.5–2.9 Ga data, we obtained a two-tailed p value of 0.0357 which is against the null hypothesis. Accordingly, the corresponding p value for 1.5–2.9 Ga model and 0.5–1.5 Ga data was 0.0522, which also remains below our threshold value of 0.1. Put together, these results are in favour of statistically significant difference between the *S* parameter in Mesoarchaean–Mesoproterozoic and Meso-Neoproterozoic data and an overall difference between the S parameter in the Precambrian and in the most recent eras.

The latitude dependence of the observed scatter has been thought to be merely a manifestation of an invalid conversion from directions to poles rather than an inherent character of the geomagnetic field (Linder & Gilder 2012). An underlying assumption of this theory, however, is a latitudinally invariant scatter of directions which is in contradiction with our data and in previously published results of Keweenawan (Halls & Pesonen 1982) and the entire globe (Smirnov *et al.* 2011) alike. To fit an arbitrary model to our directional data, we applied the original parametric equation for the scatter of directions with respect to latitude (Irving & Ward 1964), referred to as Model A:

$$s = 46.8\sigma(1 + 3\sin^2\lambda)^{-1/2}. \tag{4}$$

Our best-fitting value for the term 46.8σ was $s = 15.07 \pm 1.30$ which envelops the value $S = 13.83$ obtained from the data of

Table 2. Values of symmetric (*b*) and antisymmetric (*a*) Model G terms, corresponding *b*/*a* ratios in different PSV studies of Phanerozoic and Precambrian geomagnetic field, and reversal rates calculated for the Precambrian using the PALEOMAGIA database (Veikkolainen *et al.* 2014a). N corresponds to the total number of entries in the corresponding time interval, with the number of mixed- and combined-polarity entries given in brackets after filtering out the individual N and R subentries in combined (*c*) data. Although the notations of *a* for antisymmetric terms and *b* for symmetric terms are commonly employed, Biggin *et al.* (2008a) used *a* to represent symmetric and *b* to represent antisymmetric terms. These have been converted here to match our definition. References: [1] Smirnov *et al.* (2011), [2] Biggin *et al.* (2008a) and [3] this compilation.

Age	*b*	*a*	*b*/*a*	*N*	Rev. rate	Ref.
0–5 Ma (TAFI)	13.24 ± 0.81	0.25 ± 0.03	53.0 +10.9/−8.6	–	–	[1]
5–195 Ma	14.10 ± 1.24	0.25 ± 0.04	56.4 +16.6/−12.1	–	–	[1]
1.0–2.2 Ga	11.10 ± 1.46	0.21 ± 0.09	52.9 +53.8/−20.7	2072 (394)	0.22 / Ma	[1]
2.2–3.0 Ga	7.56 ± 0.84	0.22 ± 0.02	34.4 +7.6/−6.4	348 (67)	0.19 / Ma	[1]
0–5 Ma	11.9 ± 0.7	0.26 ± 0.02	45.8 +6.7/−5.8	–	–	[2]
0–195 Ma	15.5 +1.9/−3.3	0.27 +0.10/−0.05	54.7 +21.7/−20.7	–	–	[2]
2.45–2.82 Ga	5.9 ± 2.1	0.30 +0.05/−0.08	19.7 +16.7/−8.8	184 (42)	0.23 / Ma	[2]
0.5–1.5 Ga	10.07 ± 0.54	0.26 ± 0.04	38.7 +9.5/−7.0	1536 (427)	0.27 / Ma	[3]
1.5–2.9 Ga	9.21 ± 1.14	0.22 ± 0.02	41.9 +9.9/−8.2	1332 (287)	0.20 / Ma	[3]

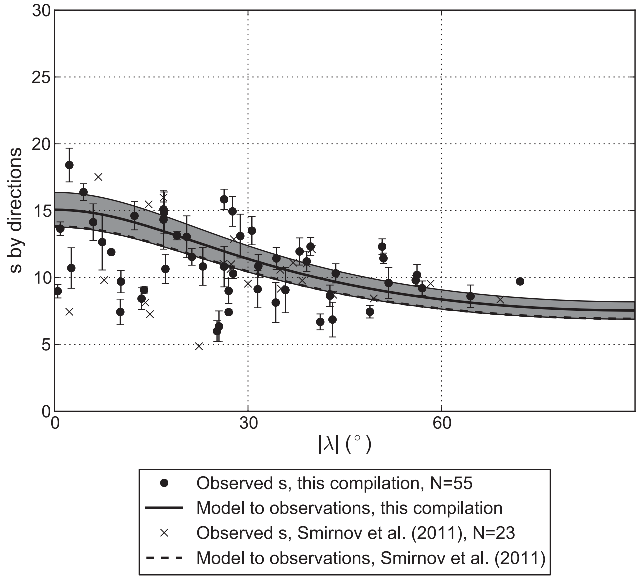

Figure 4. The observed dispersion of palaeomagnetic directions by the absolute value of paleolatitude (|λ|) for our pan-Precambrian paleomagnetic data set and a corresponding model fit ($N = 55$). Each entry is accompanied by its error limits corresponding to 2σ (two standard deviations from the mean) after transforming the within-site corrected data from pole space to direction space. The best statistical fit along with its error envelope is also shown. For comparison, data used by Smirnov *et al.* (2011) and a model fit to their data are visible too.

Smirnov *et al.* (2011) as seen in Fig. 4. Although the concept of directional scatter is effectively outdated in studies of PSV (Tauxe & Kent 2004; Deenen *et al.* 2011), we can note that a slight dependence of S with respect to palaeolatitude is visible. It is also evident that a transformation of the theoretical directional scatter curves presented by Linder & Gilder (2012) to pole space produces S curves with shapes deviating from the TAFI fit of the last 5 Ma, especially at high palaeolatitudes where the TAFI fit renders an increasing growth of S values by λ according to the standard $S = \sqrt{(a\lambda)^2 + b^2}$ function.

By taking the within-site scatter S_w into account, the total scatter of VGPs as derived from individual PSV entries decreased slightly, 1.2° in average, but with an observable dependence on palaeolatitude. Hence it is neither allowed to neglect the S_w in any PSV study, nor to use a constant value for site-level data (Biggin *et al.* 2008b). There is close correspondence with the 1.0–2.2 and 2.2–3.0 Ga models of PSV (Smirnov *et al.* 2011) in our compilations of 0.5–1.5 and 1.5–2.9 Ga Precambrian data, but the TAFI fit for the last 5 Ma (Smirnov *et al.* 2011) clearly shows a statistical difference from our data sets. Values of S_B from the Mesoarchaean–Mesoproterozoic (1.5–2.9 Ga) show generally smaller values of S, and also a smaller latitudinal increase via parameter a, than the values of the Mesoproterozoic–Neoproterozoic (0.5–1.5 Ga) do, though this assumption be treated with caution due to the paucity of high-quality data in the early Precambrian, especially at steep palaeolatitudes. Despite being an plausible hypothesis in explaining several paleogeographic problems in Meso- and Neoproterozoic, such as the occasional large paleolatitudinal shifts of Rodinia and the tendency of continents to occypy low palaeolatitudes (Evans 2003), true polar wander (TPW) does not offer any robust explanation for the difference in our Mesoarchaean–Mesoproterozoic and Mesoproterozoic–Neoproterozoic PSV data sets and is not further discussed. The study of reversal rate is, however, a point of interest since it provides a robust estimate of the stability of the field (Coe & Glatzmaier 2006).

With the absence of long continuous polarity patterns in the Precambrian considered, no accurate estimates for the reversal rate of the geomagnetic field can be given, though Coe & Glatzmaier (2006) presented a value of 0.2 reversals/Ma, compared to the value of 1.7 reversals/Ma as observed for the last 150 Ma. While Coe & Glatzmaier (2006) calculated their value by averaging over several well-known magnetostratigraphic records (e.g. Gallet *et al.* 2000; Elston *et al.* 2002; Strik *et al.* 2003), our approach is based on the database-wide calculation of the percentage of results with the MV(6) criterion (i.e. the presence of reversals) fulfilled, following the workflow of Roberts & Piper (1989) and acknowledging the crude nature of this approach. For the combined entries with both polarities present, the individual ‘N’ and ‘R’ subentries were not considered but the superseded ‘C’ result, with both N and R subsets included, was taken into account. After this procedure, 2868 records remained, and 719 of them (25.1 per cent) satisfied the MV(6) condition, all of them of combined (‘C’) or mixed (‘M’) polarity. With the timespan of our database (2942 Ma) considered, we ended up to the rate of 0.24 reversals/Ma, which is surprisingly close to the value by Coe & Glatzmaier (2006).

Because our temporal intervals were not of the same length, and the number of observations was greater in our 0.5–1.5 Ga subset when compared to that of 1.5–2.9 Ga, we did not calculate absolute reversal rates for both intervals separately, but rather, determined which interval had a higher percentage of combined- and mixed polarity data, and then normalized these percentages with the number of entries within each of the two intervals. We observed that 427 out of 1536 entries (27.8 per cent) in the 0.5–1.5 Ga subset were of ‘M’ or ‘C” polarity, compared to 287 out of 1332 entries (21.2 per cent) in the 1.5–2.9 Ga subset. These correspond to 0.27 reversals/Ma between 0.5 and 1.5 Ma and 0.20 reversals/Ma between 1.5 and 2.9 Ga, thus supporting the theory of a lower reversal frequency in the farther past (Biggin *et al.* 2008a). When using the 1.0–2.2 and 2.2–3.0 Ga time intervals of Smirnov *et al.* (2011), we could also find a small decrease in the reversal rate with respect to time (Table 1). It is evident that repetition, including studies multiply performed on same rocks, is visible in Precambrian data, and the correlation of coeval or nearly coeval reversal records from different parts of the globe is in most cases impossible, but very large changes to our database would be needed to account for a reversal rate which is closer to the 0–150 Ma value than our Precambrian result given in this study. Although we can truthfully estimate only the lower limit of the reversal rate, it must be emphasized that any proof of duplicate observations of the same reversal as recorded by rocks in different parts of the globe would lead us to adjust to our estimate of the reversal frequency to an even lower value.

The major finding of our PSV study is that the geomagnetic field of the longest geological timescales has been subject to a lesser degree of temporal variation than the field of the last 5 Ma due to the smaller rate of PSV especially before 1.5 Ga. This is most strongly visible in the symmetric terms in the field, with our pan-Precambrian model systematically showing lower values but a nearly similar shape of the latitudinal variation curve when compared to the TAFI fit. The question of the dipolarity of the average field is generally more problematic to answer using PSV data alone, since the contribution of the temporally averaged field in the palaeomagnetic VGP scatter has been estimated to be not more than 2–5°, almost independent of palaeolatitude (Hulot & Gallet 1996). However, the estimated values of the axial quadrupole (G2) for the last 5 Ma range from 2 to 4 per cent of GAD and those of the axial octupole

(G3) are as small as 1–5 per cent of GAD (Johnson *et al.* 2008). For the Precambrian, the corresponding values are 0 per cent for G2 and 6 per cent for G3 (Veikkolainen *et al.* 2014c), so while studies of PSV are useful in analyzing the changes of the geomagnetic field in small timescales, these changes are almost entirely averaged out in the long-term field which can be approximated using the GAD alone.

Although the different patterns of PSV in the Mesoarchaean–Mesoproterozoic (1.5–2.9 Ga) and Meso-Neoproterozoic (0.5–1.5 Ga) data cannot be directly attributed to estimating the validity of the GAD hypothesis or any other model of the time-averaged field, they are coincident with the decrease of statistically significant asymmetric field reversals from the Neoproterozoic to the Archaean (Veikkolainen *et al.* 2014b). This is supplemented by the overall decrease of reversal rate and previous theories of a smaller or nonexistent inner core (Coe & Glatzmaier 2006; Biggin *et al.* 2008a; Pozzo *et al.* 2012). The assumption of a higher stability of geodynamo, coupled with fewer reversals, gains additional support.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Our complete PSV data compilation is provided as two appendices (A and B) in the online version of the article. In Appendix A, the general arrangement of PSV records used in the analysis is shown, and in Appendix B, the complete site-level data are listed. Both appendices are in the form of Excel spreadsheets. (http://gji.oxfordjournals.org/lookup/suppl/doi:10.1093/gji/ggu348/-/DC1).

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