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Salminen, Marja Johanna

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J. Salminen, R. Klein, T. Veikkolainen, S. Mertanen, I. Mänttari

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Mesoproterozoic geomagnetic reversal asymmetry in light of new paleomagnetic and geochronological data for the Häme dyke swarm, Finland: Implications for the Nuna supercontinent

Salminen, J.^{1*}, Klein, R.¹, Veikkolainen, T.¹, Mertanen, S.², and Mänttari, I.²

(1) Department of Physics, University of Helsinki, P.O. Box 64, 00014 Finland; Johanna.m.salminen@helsinki.fi; robert.klein@helsinki.fi; toni.veikkolainen@helsinki.fi

(2) Geological Survey of Finland, P.O. Box 96, FI-02151 Espoo, Finland; satu.mertanen@gtk.fi; irmeli.manttari@gtk.fi

* Corresponding author

Abstract

Baltica represents one of the key continents of the Mesoproterozoic supercontinent Nuna forming the core of it together with Laurentia and Siberia. This study presents new geochronological and paleomagnetic data obtained for Häme diabase dyke swarm in southern Finland. New U-Pb (baddeleyite) ages 1642 ± 2 Ma and 1647 ± 14 Ma for two reversely magnetized dykes are acquired. Demagnetization revealed a dual polarity remanent magnetization direction carried by magnetite. The combined normal (N) and reversed (R) polarity direction for 11 dykes (=sites) is $D = 355.6^\circ$, $I = -09.1^\circ$ ($k = 8.6$ and $\alpha_{95} = 16.6^\circ$) yielding a paleomagnetic pole at 23.6°N , 209.8°E ($K = 10.6$ and $A_{95} = 14.7^\circ$) with Van der Voo value $Q = 7$. N and R magnetized units for the Häme dyke swarm show asymmetry in declination values, probably caused by an age difference between the dykes. The Geocentric Axial Dipole (GAD) model indicates that all geomagnetic reversals should be symmetric (in inclination), yet it has been noted that this is not always the case (e.g. 1.57 Ga Satakunta and Åland dykes in Baltica). By analyzing global dual polarity paleomagnetic data we show that the GAD model is a valid assumption at 1.7 – 1.4 Ga and that the asymmetry between some

normal and reversed polarities in global dual-polarity data sets appears randomly over time, and does not follow a global trend. Further, we show that in the case of Åland and Satakunta dykes an unremoved secondary magnetization component could explain the obtained asymmetry. GAD assumption is used to reconstruct the core of Nuna on equatorial latitudes using new data for Häme dykes. Paleomagnetic evidence suggest that maximum assembly of Nuna occurred at 1.5 Ga and the dispersal of the core is proposed to be associated with coeval 1.38 – 1.27 Ga magmatism in its core continents.

Keywords: paleomagnetism, geochronology, Nuna, reversal asymmetry, geocentric axial dipole

1. Introduction

Most cratonic blocks of Earth crust show evidence of collisional events between 2.1 and 1.8 Ga, which has led many researchers to propose that a Mesoproterozoic supercontinent Nuna (a.k.a. Columbia, and Hudsonland; Meert, 2012; Williams et al., 1991, respectively) existed in the Early Proterozoic (e.g. Williams et al., 1991; Hoffman, 1997; Meert, 2002; Rogers and Santosh, 2002; Zhao et al., 2004; Zhang et al., 2012). Many proposed Nuna models differ from each other (e.g., Williams et al., 1991; Hoffman, 1997; Meert, 2002; Rogers and Santosh, 2002; Pesonen et al., 2003; Zhao et al., 2004; Condie, 2004; Zhang et al., 2012; Pisarevsky et al., 2014; Pehrsson et al., in review). However, there is a consensus that Baltica and Laurentia form the core of the Nuna supercontinent in the geologically and paleomagnetically viable North Europe North America (NENA; Gower et al., 1990) connection where northern Norway and Kola Peninsula of Baltica are facing northeastern Greenland of Laurentia between ca. 1.75 and ca. 1.27 Ga (Salminen and Pesonen, 2007; Evans and Pisarevsky, 2008; Lubnina et al., 2010; Pisarevsky and Bylund, 2010; Evans and Mitchell, 2011; Pesonen et al., 2012), but different configurations have also been presented

by Johansson (2009) and Halls et al. (2011). Based on Mesoproterozoic passive margins surrounding Siberia (Pisarevsky and Natapov, 2003) and similar geology between Siberia and Western Greenland from 1.9 Ga onward it was recently proposed that Siberia forms the Nuna core together with Baltica and Laurentia in tight fit between East Siberia and Western Greenland (e.g. Rainbird et al., 1998; Wu et al., 2005; Evans and Mitchell, 2011; Ernst et al., 2016; Evans et al., 2016). This tight fit is supported by 1.8 - 1.38 Ga paleomagnetic data from Siberia, Baltica and Laurentia, but an alternative view has also been presented by Pisarevsky et al. (2008). Adding cratons around the core of Nuna has been a major initiative among paleogeographers in recent years (e.g., Congo-São Francisco, Salminen et al, submitted; India, Pisarevsky et al., 2013; North China, Zhang et al., 2012, Xu et al., 2014; Australia cratons, Payne et al., 2009; Li and Evans, 2011; Pehrsson et al., in review; Amazonia, Johansson, 2009; Bispo-Santos et al., 2012; D'Agrella-Filho et al., 2012; 2016) that has lead to the paleogeographical model of Nuna to take shape.

In this study we use the paleomagnetic method, which is the only quantitative tool for a Nuna reconstruction. The number of good quality paleomagnetic data has increased in recent years enabling more reliable global reconstructions based on paleomagnetic data (e.g Evans and Mitchell, 2011; Zhang et al., 2012). Application of paleomagnetic data to the reconstructions requires not only a high reliability of the remanence directions, but also confidence in accurate dating of the rocks and their acquisition of remanent magnetization. A reliable paleomagnetic pole generally fulfills at least three of the seven quality criteria of Van der Voo (1990). If two of these include an adequately precise geochronological age and a positive paleomagnetic field test, the obtained paleomagnetic pole can be called a “key” pole (Buchan et al., 2000; Buchan, 2013). Paleomagnetic data for different continents allow comparison of lengths and shapes of apparent polar wander paths (APWPs) to test proposed long-lived proximities of the cratons. As long as these landmasses have traveled together they should have identical APWPs. In the absence of well-defined APWPs, pairs of coeval

paleomagnetic poles from these cratons can be used for a rough test (e.g. Buchan et al., 2000; Evans and Pisarevsky, 2008). These pairs of paleopoles should plot on top of each other within their error limits, after Euler rotation to the continents' correct relative configuration.

To further test the NENA connection, the Early Mesoproterozoic 1.65 Ga Häme dykes that are related to rapakivi granite magmatism in southern Finland are studied, since they can provide a good quality coeval paleopole to the 1.63 Ga Melville Bugt pole (Halls et al., 2011) for Laurentia. A paleomagnetic study on the Häme dyke swarm has been carried out earlier, however only a secondary component was obtained (Neuvonen, 1967) most probably due to the inadequate demagnetization method that was used (single-step AF demagnetization at 30 mT). Here we use modern demagnetization techniques combined with U-Pb isotope geochronology and extend the analysis to studying the geomagnetic polarity asymmetry, using early Mesoproterozoic paleomagnetic data.

The axial dipolar field model of Earth's geomagnetic field is the key assumption for interpreting the past latitude and geography of the continents from paleomagnetic data. The model indicates that all geomagnetic reversals should be symmetric, meaning that obtained normal (N) and reversed (R) magnetization directions of dual polarity data are antiparallel. Notable asymmetry has been obtained earlier at 1.1 Ga in the Keweenaw rocks of Laurentia (e.g. Palmer, 1970; Pesonen and Nevanlinna, 1981; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Nevanlinna and Pesonen, 1983; Schmidt and Williams, 2003), but Swanson-Hysell et al. (2009) showed that it was an artefact of the rapid motion of North America during this time. In addition, pronounced asymmetric paleomagnetic results have recently been obtained for Mesoproterozoic diabase dykes in Åland and Satakunta, southern Finland (Salminen et al., 2014; 2015). Several different reasons could explain this: i) unusual behavior of the geomagnetic field, especially permanent non-dipolar field contamination (e.g., Pesonen and Neuvonen 1981; Veikkolainen et al., 2014a,b), ii) an unremoved secondary component (Halls et al., 2011); iii) an age difference

(associated with tectonic drift) between normal and reversed polarity directions (Swanson-Hysell et al., 2009), iv) relative crustal tilting of blocks with N- and R- polarity directions (Halls and Shaw, 1988). A further aim of this study is to test the possibility of a non-dipolar field contamination during the Mesoproterozoic and to analyze the effect of secondary components to dual polarity data.

2. Geological background

The Häme dyke swarm is located in the Svecofennian domain of the Fennoscandian Shield in southern Finland. The Svecofennian domain was formed at 1920-1870 Ma as a result of accretion of several island-arcs and microcontinents against the Archean craton in the NE (Lahtinen et al., 2005). Lithologically the rocks are predominantly composed of Paleoproterozoic metasedimentary and metavolcanic rocks of island-arc type, and granitoids which have intruded into these supracrustal rocks. The sedimentary and volcanic rocks were metamorphosed under high temperature and low-pressure conditions (Korsman, 1977; Korsman et al., 1984). After the stabilization of the Svecofennian domain, about 200 Ma later, the crust in southern Finland was intruded by Mesoproterozoic rapakivi granite batholiths and stocks that sharply cut the surrounding Paleoproterozoic bedrock (e.g. Rämö 1991). The total age range from U-Pb dating for the Fennoscandian rapakivi province is ca. 1.65 – 1.5 Ga. The rapakivi granites are associated with diabase and quartz porphyry dyke swarms that radiate from the rapakivi granite (Fig. 1). We will use the term “Subjotnian” later in this paper when we refer to these Early Mesoproterozoic units in the southern part of Finland. The Häme diabase dyke swarm is associated with the Ahvenisto rapakivi granite which forms a separate satellite complex north of the colossal Wiborg rapakivi granite batholith. The Ahvenisto satellite consists of a horseshoe-shaped gabbro-anorthosite rim and a central rapakivi granite batholith in which the major rock type is biotite granite (Savolahti,

1956). The rapakivi granites in southeastern Finland were emplaced at shallow crustal levels over long periods. The Wiborg, Suomenniemi and Ahvenisto granites have U-Pb zircon ages of 1650-1620 Ma (e.g. Vaasjoki et al., 1991) and according to Rämö et al. (2014) the emplacement of the plutonic rocks of the Wiborg batholith took place 12 Ma at minimum.

Most diabase dykes around the rapakivi areas of southeastern Finland are considered to be coeval or slightly older than the rapakivi granite, but genetically associated. The contemporaneous diabase dykes presumably represent derivatives of the mantle-originated thermal perturbations that caused anatexis of deep parts of the crust and subsequent emplacement of rapakivi granite batholiths in an extensional tectonic setting (Rämö, 1991; Haapala and Rämö, 1992). The rapakivi granites are shown to cut the diabase dykes. Presumably the upward movement of rapakivi granite melt continued after the injection of the diabase dyke magma and eventually the rapakivi massifs cut the diabase dykes (Laitakari, 1969). According to Laitakari and Leino (1989) the Ahvenisto gabbro anorthosite pluton probably formed the magma chamber for two types of diabase dykes (see below) radiating from the Ahvenisto pluton. The intrusion of rapakivi massifs disrupted the bedrock around their margins by forming steep faults and joints that were filled with basaltic magma. However, the diabase magma most likely intruded along structures that best agreed with the direction of the deep faults, and the jointing systems of the bedrock may thus be older than the diabase dykes (Laitakari, 1969).

The Häme dyke swarm extends ca. 150 km NW from the Wiborg and Ahvenisto rapakivi granites. Several earlier geochronological results on the dyke swarms related to Wiborg rapakivi have been reported (Vorma, 1975; Vaasjoki, 1977; Laitakari, 1987; Siivola, 1987; Vaasjoki and Sakko, 1989; Vaasjoki et al., 1991; Heinonen et al., 2010). The most reliable ages are the 1646 ± 6 Ma (U-Pb, zircon) for the Ansio diabase dyke trending N60W (Häme swarm; Laitakari, 1987); the 1635 ± 3 Ma (U-Pb, zircon) for the Nikkari quartz porphyritic dyke (Suomenniemi swarm; Vaasjoki et al., 1991); 1643 ± 5 Ma

(U-Pb, zircon) for the Lovasjärvi diabase dyke (Suomenniemi swarm; Siivola, 1987); and 1636 ± 2 Ma (U-Pb, zircon) for the Leviänlahdenvuoret quartz porphyritic dyke (Ahvenisto swarm; Heinonen et al., 2010). These are coeval with a gabbro-anorthosite surrounding the Mäntyharju rapakivi intrusion (Vaasjoki and Sakko, 1989). Older U-Pb zircon ages of 1690 Ma for the Hyvärilä diabase dyke (Ahvenisto swarm; Vormaa, 1975), and 1667 ± 9 Ma for the N80°NW Virmaila dyke (Häme swarm; Vaasjoki et al., 1991) were obtained. However they are considered unreliable in light of the age determinations shown in this study.

The widths of the Häme dykes vary from 250 m to a few centimetres (Laitakari, 1969). The dykes comprise two dyke sets with different trends and compositions (Laitakari, 1969, 1987; Luttinen and Kosunen, 2006, Vaasjoki and Sakko, 1989). One dyke set is about 100 km long and have dominating strikes of 45° - 60°NW. The other dykes trend in 85°NW direction and can be followed for about 150 km from the southern part of the Ahvenisto pluton through Orivesi up to Kuru. According to Laitakari (1969) there are several observations where the trend is between these maxima (60 - 85°NW). The dykes dip vertically or subvertically. In general, the widths of the dykes get narrower the longer the distance is from the rapakivi granite. Contacts with the host rock are typically sharp. Non-crystalline glass has been discovered in some of the narrowest dykes (Lindqvist and Laitakari, 1980). In some of the dykes there are amygdoloids which indicate crystallization of magma close to the surface (Laitakari, 1987).

Compositionally the Häme diabase dykes are unmetamorphosed olivine diabases where the main minerals are plagioclase, olivine and clinopyroxene (Laitakari, 1987). The dykes may also contain biotite, potassium feldspar and orthopyroxene. The 80°NW trending diabase dykes are mainly olivine tholeiites and typically contain abundant olivine, but no plagioclase phenocrysts or big plagioclase fragments (Laitakari and Leino, 1989). The 60° NW trending dykes are characterized by phenocrysts, megacrysts (up to < 20

cm) and fragments of plagioclase (Laitakari, 1969) which due to flowage differentiation are concentrated in the central parts of the dyke.

3. Sampling and Methods

3.1 Sampling

Standard 2.5-cm diameter cores were collected with a portable field drill from 22 diabase dykes of Häme swarm for paleomagnetic measurements during field campaigns in 2009 and 2014 (Fig. 1). Host rocks were sampled for baked contact tests (Everitt and Clegg, 1962) at 11 of the dyke sites. Additional unbaked host rock sites were sampled in 2016. Cored samples were oriented using solar and/or magnetic compasses. One new geochronology sample was taken from the more than 60 m wide Torittu dyke (site H17) for U-Pb dating. The Torittu dyke has ophitic texture which is spotted due to plagioclase clusters, 1-2 cm in diameter (Laitakari, 1969). The interstices between the plagioclase laths are filled by olivine and augite. Accessory minerals are titanomagnetite, biotite, apatite, zircon and serpentine as the alteration product of olivine. Another new age dating was done on existing baddeleyite from Virmaila dyke (VR, width 60 m) that was reanalyzed using Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) method.

3.2 Geochronological methods

A ca. 5 kg whole-rock sample from a Torittu dyke (H17) were prepared for dating. Zircon and baddeleyite for Laser Ablation Multicollector Inductively Coupled Plasma Mass Spectrometry (LA-MC-ICPMS) U-Pb dating were selected by hand-picking after heavy liquid (CH_2I_2 , and Clerici solution) and Frantz magnetic separation. The dyke sample yielded 20 baddeleyite grains (width $> 20 \mu\text{m}$) and 15 zircon grains. The chosen grains were mounted

in epoxy resin and sectioned approximately in half and polished. Back-scattered electron images (BSE) and cathodo luminescence (CL) images were taken using SEM (Scanning Electron Microscope) to target the spot analysis sites on the mineral grains. This sample was assigned sample code A2414 for the Finnish Rock Age Database of the Geological Survey of Finland (GTK).

U–Pb dating analyses for Torittu sample were performed using a Nu Plasma HR multicollector ICPMS at the Geological Survey of Finland in Espoo using a technique very similar to Rosa et al. (2009) except that an Analyte G2 193 nm laser laser microprobe was used. The analyses were made in static ablation mode with a beam diameter of 20 μm , pulse frequency of 10 Hz, and beam energy density of 2.07 J/cm². A single U–Pb measurement included 30 s of on-mass background measurement, followed by 60 s of ablation with a stationary beam. Masses 204, 206 and 207 were measured in secondary electron multipliers and 238 in the extra high mass Faraday collector. Ion counts were converted and reported as volts by the Nu Plasma time-resolved analysis software. ²³⁵U was calculated from the signal at mass 238 using a natural ²³⁸U/²³⁵U=137.88. Mass number 204 was used as a monitor for common ²⁰⁴Pb. Raw data were corrected for background, laser induced elemental fractionation, mass discrimination, and drift in ion counter gains and reduced to U–Pb isotope ratios by calibration to concordant reference baddeleyite of known age, using protocols adapted from Andersen et al. (2004) and Jackson et al. (2004). In-house standard baddeleyite A974 (1256.2 $\pm$ 1.4 Ma; Söderlund et al., 2004) was used for calibration. Age related common lead (Stacey and Kramers, 1975) correction was used when the analysis showed common lead contents above the detection limit. The calculations were done off-line, using an interactive spreadsheet program written in Microsoft Excel / VBA by Tom Andersen (Rosa et al., 2009). To compensate for drift in instrument sensitivity and Faraday vs. electron multiplier gain during an analytical session, a correlation of signal vs. time was assumed for the reference zircons. A description of the algorithms used is provided

in Rosa et al (2009). The U–Pb isotopic data were potted and the age calculations were performed using the Isoplot/Ex 3 program (Ludwig, 2003). Ages were calculated with 2σ errors and without decay constants errors. Data-point error ellipses in the Fig. 2 are at the 2σ level.

Another new age dating was done on existing baddeleyite from Virmaila dyke (VR, width 60 m) that was reanalyzed using Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) method. Earlier analyses for Virmaila diabase dyke had yielded an age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1988). We reanalyzed the existing baddeleyite from Virmaila dyke (VR, width 60 m). The decomposition of baddeleyite and extraction of uranium and lead for Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) age determinations mainly follows the procedure described by Krogh (1973, 1982). ^{235}U - ^{208}Pb -spiked and unspiked isotopic ratios were measured using a VG Sector 54 TIMS. The measured lead and uranium isotopic ratios were normalized to the accepted ratios of SRM 981 and U500 standards. The Pb/U ratios were calculated using the PbDat program (vers.1.24; Ludwig, 1993). The concordia plots and the final age calculations were done using the Isoplot/Ex 3.00 program (Ludwig, 2003). The common lead corrections were done using the age related Stacey and Kramers (1975) lead isotope compositions ($^{206}\text{Pb}/^{204}\text{Pb} \pm 0.2$, $^{208}\text{Pb}/^{204}\text{Pb} \pm 0.2$, and $^{207}\text{Pb}/^{204}\text{Pb} \pm 0.1$). The total procedural blank level was 20-50 pg. All the ages are calculated with 2σ errors and without decay constant errors. In Fig. 3, the data-point error are at 2σ level.

3.3 Rock magnetic and paleomagnetic methods

Rock magnetic and paleomagnetic measurements were carried out at the Solid Earth Geophysics Laboratory of the University of Helsinki (UH), Finland. Magnetic mineralogy was investigated by thermomagnetic analysis of selected powdered whole-rock samples.

Temperature dependence of low-field magnetic susceptibility was measured from -192°C to ~700°C (in argon gas) followed by cooling back to room temperature using an Agico CS3-KLY-3S Kappabridge system, which measures the bulk susceptibility of the samples. Curie temperatures were determined using the Cureval 8.0 program (<http://www.agico.com>).

Stepwise alternating field (AF) demagnetizations were done using a three-axis demagnetizer with maximum field up to 160 mT, coupled with a cryogenic 2G (now WSGI) DC SQUID magnetometer to isolate the characteristic remanent magnetization (ChRM) component. Sister specimens were thermally demagnetized using an argon-atmosphere ASC Scientific model TD-48SC furnace. Remanent magnetizations were measured with a DC SQUID magnetometer. Vector components were visually identified using stereographic and orthogonal projections (Zijderveld, 1967) and the directions were calculated by a least squares method (Kirschvink, 1980). Mean remanence directions for the different components were calculated according to Fisher (1953), giving a unit weight to each sample (each specimen has a unit weight within a sample) to compute site mean directions and corresponding virtual geomagnetic poles (Irving, 1964). The paleogeographical reconstructions and apparent polar wander paths (APWP) were done with the GPlates program (Boyden et al., 2011; Gurnis et al., 2011; Williams et al., 2012).

4. RESULTS

4.1 U-Pb geochronology results

Results for LA-MC-ICPMS U-Pb baddeleyite data from the Torittu diabase dyke sample is presented in Table 1 and in Fig. 2. Only one small elongated magmatic zircon grain was dated, yielding an age of 1.90 Ga. This is either an inherited grain or contamination from the rock crushing and separating processes, and will not be discussed further. Eleven baddeleyite domains were dated, all of which were euhedral and translucent. Two analyses showing

major discordance and high common lead contents (2.6% and 3.5%, respectively) were rejected (Table 1). We further ignored the two “youngest” data points with lower error correlation values and moderate common lead contents (1.9% and 2.2%, respectively) and calculated a concordia age of 1647 ± 14 Ma (MSWD=0.03; n=7) (Fig. 2).

Results in the reanalysis of the baddeleyite from the Virmaila dyke two fractions of best existing baddeleyite of diabase sample of Virmaila dyke using the ID-TIMS is presented in Table 2 and in Fig. 3. Two fractions of the best quality baddeleyite grains were air-abraded for 30 minutes. The U-Pb TIMS measurements (Table 2) yielded concordant and nearly concordant ages with a mean of 1642 ± 2 Ma (Fig. 3) being notably younger than the earlier age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1989).

4.2 Rock magnetic results

Examples of representative rock magnetic results are shown in Fig. 4. Temperature dependence of low-field magnetic susceptibility at low temperatures (heating from -192°C to room temperature) shows the presence of nearly stoichiometric magnetite in diabase samples indicated by the Verwey transition (Verwey, 1939) at around -153°C to -146°C . Temperature dependence of low-field magnetic susceptibility at high temperatures (in argon) show the presence of Hopkinson peak and Curie temperatures at range of $570 - 582^{\circ}\text{C}$ being consistent with magnetite and/or low-Ti titanomagnetite as the magnetic carrier. Small irreversibility between heating and cooling curves may be due to slight mineralogical changes during the heating.

4.3 Paleomagnetic results

4.3.1 Primary remanent magnetization

Paleomagnetic results are listed in Table 3, and representative demagnetization behaviors are illustrated in Figs 5 and 6. We measured a total of 125 diabase samples from 22 dykes; total of 44 baked host rock samples for 11 of the dyke sites and 33 unbaked samples. Eleven dykes (45 samples) show presumably primary characteristic remanent magnetization (ChRM) dual-polarity direction with northerly and southerly declinations and shallow up- and downward pointing inclinations (Fig. 7). The remaining sampled Häme dykes were remagnetized or showed unstable directions (see Table 3). In general, samples showing primary ChRM were stable to both AF and thermal demagnetization methods. ChRM was obtained with 30-160 mT AF fields and with unblocking temperatures up to 585°C. We accepted the data from dykes where at least two samples show similar directions. Ideally components with Mean Angular Deviation (MAD) values less than 5° were included, but in some cases a higher MAD was accepted if a clear ChRM was observed. The mean direction for ChRM of five normal polarity Häme dykes is $D = 015.8^\circ$, $I = -09.2^\circ$ (with $k = 10.8$, $\alpha_{95} = 24.3^\circ$) and the palaeomagnetic pole derived from VGPs is at 22.4°N, 187.9°E (with $K = 14.8$ and $A95 = 20.6^\circ$). For six reversed polarity Häme dykes the ChRM mean is $D = 159.3^\circ$, $I = 8.1^\circ$ (with $k = 17.0$, $\alpha_{95} = 16.7^\circ$) (Table 3, Fig. 7) and the palaeomagnetic pole is at 22.3°N, 227.4°E with $K = 33.5$ and $A95 = 11.7^\circ$. Two of the dated dykes, Virmaila (1642 ± 2 Ma) and Torittu (1647 ± 14 Ma), show reversed polarity directions. The combined grand mean normal and reversed polarity ChRM direction for 11 sites is $D = 355.6^\circ$, $I = -09.1^\circ$ (with $k = 8.6$ and $\alpha_{95} = 16.6^\circ$) yielding a paleomagnetic pole at 23.6°N, 209.8°E (with $K = 10.6$ and $A95 = 14.7^\circ$).

Since the primary remanent magnetization of the dykes has thermal origin by default, the baked contact test (Everitt and Clegg 1962) can be used to verify the primary nature of magnetization. Baked-contact tests were performed at eleven dyke sites of the Häme region, but due to unstable remanent magnetization in the host rocks it was successful only for the dated 1642 ± 2 Ma Virmaila dyke. Reversely magnetized Virmaila dyke (sites VR, HR, H8) provides a full positive baked contact tests (Fig. 6, Table 3). The dyke and the

baked host rock (Figs 1 and 6, Table 3) show a shallow southerly (reversed polarity) pointing ChRM directions. The unbaked rocks at the sampling area were not the best targets for paleomagnetic study, but we were able to obtain a typical Svecofennian type direction ($D=341^\circ$, $I=49^\circ$) from a Svecofennian aged host granodiorite, 2.5 km and ca. 1.0 km from the Virmaila dyke (Fig. 6, Table 3).

4.3.2 Secondary remanent magnetization

In addition to the primary ChRM most of the measured samples show secondary magnetization components. In many cases a viscous component resembling the Present Earth's Field (PEF) direction at the sampling site ($D = 7^\circ$, $I = 73^\circ$) was removed with fields ≤ 10 mT and temperatures $< 200^\circ\text{C}$. The majority of the samples also show another secondary component, here after called component B following Mertanen (1995), with an easterly declination and an intermediate to steep downward inclination (Fig. 7). In some cases component B was obtained from the coarser grained samples taken from the interior of the dyke and it was removed with a field of ≤ 20 mT and temperatures $< 350^\circ\text{C}$ indicating its secondary nature. In these cases a shallow primary component was obtained for samples from the finer grained margins of dyke. In other cases higher AF field (30 -160 mT) and higher temperatures $500\text{--}580^\circ\text{C}$ were needed to demagnetize component B and an original remanent magnetization direction was not obtained at all in the dyke (e.g. dyke H11). Due to the positive baked contact test the dual-polarity shallow component was interpreted to represent a primary magnetization and thereafter the component B is interpreted to represent a secondary magnetization, being similar to the one obtained in several other intrusions and shear zones in Fennoscandia (e.g. Mertanen, 1995; Preeden et al., 2009; Salminen et al. 2014; 2015).

Three dykes (H16, H19 and H22) show high coercivity remanent magnetization directions with southerly declinations and steep downward inclinations (Fig. 7)

and two dykes (H3 and H7) show high coercivity WNW declinations with downward shallow inclinations. The origins of these directions are not known and they are not further discussed here.

5. DISCUSSION AND CONCLUSIONS

5.1 Geochronology

Sampled Häme dykes are associated with Wiborg rapakivi within the Fennoscandian Shield. Age succession of Wiborg rapakivi granite from oldest in the North (1646 ± 4 Ma Värtö tirilite) and youngest in the South (1627 ± 3 Ma Ristisaari dark wiborgite) suggest that the overall locus of magmatic activity may have shifted southward during the build-up of the Wiborg batholith (Rämö et al., 2014). According to Laitakari (1969) and Vaasjoki and Sakko (1989) two sets of Häme diabase dykes occur, trending either 60°NW or 80°NW . Dykes trending 60°NW are more differentiated and they often contain megacrysts of plagioclase. The $\text{N}80^\circ\text{W}$ striking diabase set is more uniform, consisting mainly of medium grained ophitic rock.

This study provides a new U-Pb (baddeleyite) age of 1647 ± 14 Ma for a diabase dyke in 80°NW Torittu (Häme swarm) and reanalysing the existing sample from Virmaila dyke provides a new age of 1642 ± 2 Ma being clearly younger than the previous U-Pb (baddeleyite) age of 1667 ± 9 Ma (Vaasjoki and Sakko, 1988) for Virmaila. One reason for this discrepancy could be the baddeleyite fraction might have included some inherited zircons. The new age narrows down the emplacement time of Häme dyke swarm with slightly different trend of strikes and now the petrologically different dyke sets show coeval ages (Vaasjoki and Sakko, 1989). See as well Rämö et al. (2014) and Rämö and Mänttari (2015).

5.2 Häme paleomagnetic data

5.2.1 Quality of the Häme paleomagnetic pole

Earlier only magnetization direction close to present Earth's field (PEF) direction has been obtained for Häme dykes (Neuvonen, 1967). Since the 1960s demagnetization techniques have developed, and with more extensive sampling, a primary magnetization component has now been obtained for the newly dated Häme dyke swarm (1642 ± 2 Ma, 1647 ± 14 Ma U-Pb on baddeleyite). The new pole for the Häme swarm lies at 23.6°N , 209.8°E (with $K = 10.6$ and $A95 = 14.7^\circ$). This new pole fulfils seven of the Van der Voo (1990) reliability criteria providing a new key pole (e.g. Buchan and Halls, 1990; Buchan et al., 2000; Buchan, 2013) for Baltica. This stresses the limits for the required statistics, and for the baked contact test the Svecofennian aged (1.9-1.8 Ga) host rock in SE Finland is a poor carrier of remanence. The pole has: (1) well-determined U-Pb (baddeleyite) ages of 1642 ± 2 Ma and 1647 ± 14 Ma for reversely magnetized Virmaila and Torittu dykes, respectively. This is also interpreted to represent the age of magnetization; (2) The statistics of this new combined R- and N - polarity pole is just within the limits of Van der Voo (1990) criteria #2 (more than 24 samples, $K > 10$, and $A95 < 16^\circ$), but fulfils them (45 samples, $K = 10.6$, and $A95 = 14.7^\circ$). The statistics of the reversed polarity pole are better than the statistics for normal polarity pole. The reversed polarity pole fulfils criteria #2 (25 samples, $K = 33.5$, and $A95 = 11.7^\circ$), but the normal polarity pole does not (20 samples, $K = 14.8$, and $A95 = 20.6^\circ$); (3) Adequate AF and thermal demagnetization methods were used; (4) Reversely magnetized Virmaila dyke demonstrates a full positive baked contact test. Subjotnian R-polarity directions for Virmaila dyke itself are scattered with $\alpha_{95} = 38.2^\circ$, but they are clearly Subjotnian. Since there are only four specimens taken from the rather coarse grained part of this 60 m wide dyke showing Subjotnian direction, the directions can be more scattered and α_{95} larger than when more than four specimens show a similar direction. The baked host rock samples (seven samples) with

$\alpha_{95} = 18.3^\circ$ show better statistics than the dyke itself. For unbaked host rock samples the quality of data is not as good as for baked host samples, because the majority of the samples show rather unstable behaviour during demagnetization. It has been noted earlier that the Svecofennian aged (1.9-1.8 Ga) host rock in SE Finland is a poor carrier of remanence (e.g. Mertanen, 1995). Although only two unbaked host rock samples for the reversely magnetized Virmaila dyke show Svecofennian direction, their existence clearly demonstrates that the host rock carries different remanence than the dyke and that this remanence has a Svecofennian origin. Moreover, the Svecofennian component was also obtained for an unbaked host rock sample for the N polarity H1 dyke (Table 3). No stable direction for the baked host rock for H1 dyke was obtained, but the unbaked host rock sample demonstrates clearly the presence of a Svecofennian aged component. Despite the scatter of data the full positive baked contact test is demonstrated; (5) Structural control; and (6) the presence of reversals, although they do not pass the reversal test (McFadden and McElhinny, 1990). There are several case of primary remanences from a single magmatic event where reversals are not antipodal because emplacement extends over a significant time interval during which the continent was in motion (e.g. Buchan, 2013). For example, the 2.13-2.10 Ga Marathon dyke swarm of North America was emplaced over 25 Ma and shows polarity asymmetry (Halls et al., 2008). Our interpretation is that to give $Q_6 = 1$ (presence of reversals) does not require the presence of antipodal reversals. The criteria number (7) - a similarity to younger poles is satisfied as well. The pole is similar to the poles obtained from Carboniferous and Permian formations in Sweden and Norway (e.g. Torsvik et al., 1992). However, these areas are more than 500 km to the southwest of our sampling sites, and considering the positive baked contact tests for the dykes it is very unlikely that Carboniferous-Permian geological events generated the ChRM in the Häme dykes.

5.3 Asymmetry in Mesoproterozoic paleomagnetic record and geomagnetic field properties

Site mean remanent magnetization values for Häme dyke intrusions show asymmetry, i.e. the mean normal (N) and reversed (R) directions are not antiparallel at 95% confidence level in declination values (Fig. 7). Mean declination for normal polarity dykes is $D_N = 009.0^\circ$ and for reversed polarity dykes declination is $D_R = 159.3^\circ$ (339.3° when polarity is reversed to normal polarity). Inclination agrees for both polarities. Non-antiparallel directions have already been obtained earlier in other Subjotnian studies of Fennoscandia (e.g. Pesonen and Neuvonen, 1981; Salminen et al., 2014; 2015), but the asymmetry was shown in inclination values as in case for Åland and Satakunta dykes (Fig. 8). A classical example of asymmetry is the 1.1 Ga Keweenaw rocks in North America (e.g. Palmer, 1970; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Schmidt and Williams, 2003), which was first interpreted to be due to anomalous behavior of the geomagnetic field (e.g. Palmer, 1970; Halls and Pesonen, 1982; Pesonen and Halls, 1983; Schmidt and Williams, 2003), but was later shown to be a result of the rapid motion of North America during this time (Swanson-Hysell et al., 2009). We further studied the asymmetry in the Häme and other Subjotnian dyke swarms in Finland, and discuss possible reasons that could explain the data. First, the effect of secondary magnetization to distort the remanence directions was studied.

5.3.1 Secondary magnetizations

Some of the Häme dykes show secondary magnetizations, either a viscous component in the present Earth's field (PEF) direction or the NE directed intermediate inclination component B, which is common in several units in Fennoscandia (e.g. Mertanen, 1995; Salminen et al., 2014; 2015). We tested the effect of an unremoved secondary component through vector analysis, by subtracting the measured secondary component vector from the obtained normal

(N) and reversed (R) polarity ChRM directions for three cases: Åland, Satakunta, and Häme dyke swarms (Fig. 8). In the case of Åland (Salminen et al., 2015), Häme, and Satakunta (Salminen et al., 2014), a secondary component with an easterly declination and an intermediate to steep downward inclination was typically removed with a field of ≤ 20 mT. However in some cases the secondary component occurred in higher coercivities or it had even totally overprinted the sample. For our test the intensity of the secondary component was adjusted to give the most symmetric resultant N and R vectors. In the case of Åland, and Satakunta dyke swarms, an unremoved secondary component, with an intensity of 20 - 30% of the ChRM component, gave rise to a resultant vector that lies within the 95% confidence of the measured ChRM, suggesting that the asymmetry in the N and R ChRM components may be caused by an unremoved secondary component. By subtracting the secondary component the inclination gets shallower. In the case of the Häme dyke swarm, the asymmetry is only in the declination. Therefore in the vector subtraction, the inclinations for the resultant N and R vectors move further away from the observed ChRM as the declinations of the resultant N and R vectors move closer to the observed ChRM. This may suggest other factors contributing to the asymmetry, such as an age difference between the dykes.

5.3.2 Reversal test for global Mesoproterozoic paleomagnetic dual-polarity data

We carried out a reversal test (McFadden and McElhinny, 1990) on a global dataset. The results are shown in Table 4, and the units that do not pass the reversal test are outlined with black in Fig. 9. Data for Satakunta, Åland, and Häme dyke swarms of Baltica do not pass the reversal test. The large asymmetry for the 1576 Ma shallow-inclination data of the Åland and Satakunta dykes (Salminen et al., 2014; 2015) could indicate a possibility of a small axial quadrupole field (see section 5.3.3). However, some other 1.5-1.6 Ga dual polarity poles for Baltica, such as Sipoo dykes (Mertanen and Pesonen, 1995), Ragunda formation (Piper,

1979a) and Nordingrå gabbro-anorthosite (Piper, 1980) do not show asymmetric N and R polarities, indicating that other sources for obtained asymmetry is possible. Similarly, ca. 1600 Ma Satakunta and Dala sandstones (Klein et al. 2014; Piper and Smith, 1980) do not show asymmetry between N and R magnetized units. In the case of Laurentia, the Tobacco Root dykes (1448 ± 49 Ma, Harlan et al., 2008), Harp Lake complex (1450 Ma, Irving et al. 1977), and Melville Bugt dykes (1630 Ma, Halls et al. 2011) do not pass the reversal test. As with Baltica, other coeval Laurentia poles, such as Sherman granite (1431 ± 6 Ma; 1412 ± 13 Ma, Harlan et al., 1994), Snowslip formation (1443 ± 7 Ma; 1454 ± 9 Ma; Elston et al. 2002), McNamara Formation (1401 ± 6 Ma, Elston et al., 2002), Upper Belt-Purcell supergroup (1401 ± 6 Ma; 1443 ± 7 Ma; Evans et al., 1975) and upper member of Mt. Shields Formation (1401 ± 6 Ma; 1443 ± 7 Ma, Elston et al., 2002) do not show asymmetric N and R polarities. Other poles that do not pass the reversal test are Lakhna dykes from India (1466.4 ± 2.6 Ma, Pisarevsky et al., 2013), Pandurra formation from South Australia (ca. 1440 Ma, Schmidt and Williams, 2011), and Fomich River mafic intrusions from Siberia (1513 ± 51 Ma, Veselovskiy et al., 2006). Because there are no other dual polarity paleomagnetic poles for these cratons, it is not possible to determine if the asymmetry of these poles are representative of their respective cratons. One dual polarity pole for North Australia for Mt. Isa dykes (1500-1550 Ma, Tanaka and Idnurm, 1994); one for North China for Yangzhuang Formation (1437 ± 21 Ma; 1560 ± 5 Ma Wu et al., 2005); two for Amazonia for Nova Guarita dykes (1418.5 ± 3.5 Ma, Bispo-Santos et al., 2012) and for Salto de Ceu intrusives (1439 ± 4 Ma, D'Agrella-Filho et al., 2016) pass the reversal test. Although the global dual polarity dataset is relatively small, the asymmetry between some N and R polarities seem to occur randomly over time, and does not follow a global trend.

5.3.3 Geomagnetic field properties at the Mesoproterozoic

The geocentric axial dipolar (GAD) field model (Hospers, 1954) of Earth's geomagnetic field is the key assumption for interpreting the past latitude and geography of the continents from paleomagnetic data and its validity has been tested extensively in various geological timescales (e.g. Evans, 2006; Donadini et al., 2007; Biggin et al., 2008; Smirnov et al., 2011). In the case of GAD the antipodality of normal and reversed magnetization direction is expected, whereas steepening or shallowing of inclinations can result from the contamination of GAD by zonal multipolar fields. Declination data, however, remains unchanged as long as non-zonal multipolar fields are not present. So the GAD model indicates that all geomagnetic reversals should be symmetric, yet it has been noted that this is not always the case. Pronounced asymmetry in inclination was previously obtained for Early Mesoproterozoic (Subjotnian) 1.57 Ga paleomagnetic data for Åland and Satakunta, Finland. Paleomagnetic results from this study for Häme dykes show symmetric inclination, but slightly different declination values for normal and reversed polarity dykes, which cannot be tested with GAD model.

Here we aim to test the validity of the GAD model at Late Paleoproterozoic - Mesoproterozoic between 1.7 and 1.4 Ga. For this period most of the global paleomagnetic data has been produced from Baltica and Laurentia, and both continents occupy shallow and moderate latitudes establishing the geologically and paleomagnetically valid NENA (Northern Europe – North America) connection (e.g. Gower et al., 1990; Salminen and Pesonen, 2007; Evans and Pisarevsky, 2008; Salminen et al., 2014). Because of this, the global Early Mesoproterozoic paleomagnetic record is characterized by the relatively large amount of low-inclination data causing an apparent overrepresentation, which indeed is not a direct signal of a strongly non-dipolar geomagnetic field (Meert et al., 2003). Therefore the commonly used database-wide inclination test “the global inclination frequency analyses” (e.g. Evans, 1976) is not a feasible method to test the validity of GAD but alternative ways need to be considered. As a viable alternative, we applied the quantity called inclination

asymmetry (Veikkolainen et al., 2014b) for global dual-polarity paleomagnetic data. In actuality, zonal multipoles higher than octupole, and all non-zonal multipoles are beyond detection in the Precambrian, and therefore we focus our analysis on axial dipole, quadrupole and octupole fields only, using the dependence of paleocolatitude θ (90° -paleolatitude λ) and inclination I (Equation 1, Merrill and McElhinny, 1983):

$$\tan I = \left[2\cos\theta + \frac{g_2^0}{g_1^0} \left(\frac{9}{2} \cos^2\theta - \frac{3}{2} \right) + \frac{g_3^0}{g_1^0} (10\cos^3\theta - 6\cos\theta) \right] \times \left[\sin\theta + \frac{g_2^0}{g_1^0} (3\cos\theta\sin\theta) + \frac{g_3^0}{g_1^0} \left(\frac{15}{2} \cos^2\theta\sin\theta - 3\sin\theta \right) \right]^{-1} \quad (1)$$

In Equation 1, the ratio of quadrupole to dipole (g_2^0/g_1^0) and octupole to dipole (g_3^0/g_1^0) are usually referred to as G2 and G3, respectively. For pure GAD, the equation reduces to $\tan I = 2 \cot \theta$, analogous to the dipole equation $\tan I = 2 \tan \lambda$. Using this information, the difference between the observed inclination and GAD inclination, here referred to as inclination asymmetry, can be solved. It follows that the same (co)latitude can be represented by different inclinations, depending on the ratios G2 and G3. If GAD is contaminated by a multipolar field, inclinations at a certain latitude are shallowed or steepened when compared to the situation where GAD is the only field component. Because the field direction affects this behavior, dual-polarity paleomagnetic data are of particular importance for analyzing the polarity asymmetry.

For analysis, we gathered dual-polarity 1.7 - 1.4 Ga paleomagnetic data (Table 4, Fig. 9) from the PALEOMAGIA database available at <http://www.helsinki.fi/paleomagia>. We filtered the data using $Q_{1-7} \geq 3$ quality criterion for the combined polarity. (Van der Voo, 1990; Veikkolainen et al. 2014c). The resulting dataset incorporates 7 sedimentary and 16 igneous units from seven cratons: Amazonia, Baltica, India, Laurentia, North China, Siberia and South Australia. Since nearly all polarity pairs in the compilation had α_{95} error parameters determined for both polarities either from site- or sample-level statistics, we were able to

investigate whether GAD is a reasonable geomagnetic field model within error limits of observations (Fig. 10). The locations of continents have been selected to be either at northern hemisphere (positive latitudes) or at southern hemisphere (negative latitudes) based on our Nuna reconstructions (see chapter 5.4). While most cratons occupy one hemisphere, equatorial Baltica, Laurentia and Siberia are exceptions. The difference between dot-dashed and dashed lines in Fig. 10 reflects a situation where the dominant dipolar field remains unchanged, but a supplementary geocentric axial quadrupole changes its polarity. Models consisting of standing GAD and a reversing octupole were also tested, but they appeared to fit the data poorly and no further investigation was carried out. Because almost all of our observational data are between our dipole-quadrupole model curves, we may safely conclude that GAD is a relatively good fit between 1.7 Ga and 1.4 Ga, and if g_{20} is needed, its strength is less than 25% of GAD. However, the directions showing largest deviations from GAD have α_{95} errors so large that their error bars actually overlap the GAD curve in Fig. 10. These data are from Sipoo (Mertanen and Pesonen, 1995) and Ragunda dykes (Piper, 1979a) of Baltica.

5.3.4. Age difference between N and R polarity and crustal tilting

A small but significant age difference between N and R magnetized dykes could explain the asymmetry as for Keweenaw case (e.g. Swanson-Hysell et al., 2009; 2014), but the actual age span for the Subjotnian dykes for Baltica, especially for the Häme, Åland and Satakunta dyke swarm, awaits further precise dating.

Relative crustal tilting between N and R polarity “dominated” blocks is another possible explanation to the asymmetric paleomagnetic results (Halls and Shaw, 1988). However in the case of Åland and Satakunta there are no support for this since the majority

of the dips of the R- and N-polarity dykes are vertical to subvertical and the dykes with different polarity do not occur in separate blocks.

We have shown that the GAD model is a valid assumption at 1.7 – 1.4 Ga and that the asymmetry between some normal and reversed polarities in global dual-polarity data sets seem to occur randomly over time, and does not follow a global trend. One possible explanation for the asymmetry in Mesoproterozoic paleomagnetic results, seen in Häme dykes include a possible age difference between the dykes. In addition to this, an unremoved secondary magnetization component is a possible explanation for asymmetric results in Åland and Satakunta dykes. However more precise age data and more detailed paleomagnetic work on these dykes and on coeval 1.63 Ga Sipoo, 1.63 Ga Suomenniemi and 1.59 Ga Breven-Hällefors dyke swarms would be needed.

5.4 New Häme pole and its implications to Nuna supercontinent

5.4.1 Apparent polar wander path for the core of Nuna

The new 1642 ± 2 Ma; 1647 ± 1 Ma (U-Pb, baddeleyite) combined R- and N- polarity Häme paleomagnetic pole 23.6°N , 209.8°E (with $K = 10.6$ and $A95 = 14.7^\circ$) fulfils seven Van der Voo criteria, but stressing the limits of criteria #2, and adds to the Subjotnian poles for Baltica. The N-polarity ($Q=3$), R-polarity ($Q=6$) and combined pole ($Q=7$) for Häme are plotted in Figure 11 together with selected poles for Laurentia and Siberia. The N-polarity pole with lower quality plots on the younger part of the apparent polar wander path than the better quality R-polarity pole and the combined pole. The R-polarity pole and the combined pole are coeval with the pole for the Melville Bugt dyke swarm (1622 ± 3 Ma, 1635 ± 3 Ma) in Greenland (Halls et al., 2011). Including the Häme pole, there are now four Subjotnian aged key poles (Buchan et al., 2000; Buchan, 2013) for Baltica (Table 5), the other ones

being the 1575.9 ± 3.0 Ma pole for Åland dykes (Salminen et al., 2015); the 1575.9 ± 3.0 Ma pole for Satakunta N-S & NE-SW dykes (Salminen et al., 2014), and the 1452 ± 12 Ma, 1459 ± 3 Ma, 1457 ± 2 Ma (U-Pb baddeleyite; Söderlund et al., 2005) pole for Lake Ladoga mafic rocks (Salminen and Pesonen 2007; Shcherbakova et al. 2008; Lubnina et al., 2010) (Table 5, Fig. 11). Many other Subjotnian units in Baltica show a large scatter in the paleomagnetic data (recently reviewed by Salminen et al. 2014). Table 5 and Fig. 11 show the best quality data. Good quality Subjotnian, non-key poles for Baltica are the 1469 ± 9 Ma pole for Bunkris-Glysjön-Öje dykes (Pisarevsky et al., 2014), the 1633 ± 10 Ma (U-Pb, zircon, Törnroos, 1984 Ma) pole for Sipoo quartz porphyry dykes (Mertanen and Pesonen, 1995) and the 1631 Ma pole for quartz porphyry dykes in SE Finland (Neuvonen, 1986). Both in SE Finland and in Sipoo the dykes are associated with rapakivi granites. In addition to quartz porphyry dykes there are also diabase dykes which are mainly separate but in some cases form composite dykes (Laitala, 1984; Rämö, 1991). It has been suggested that the fissures in the Sipoo area were first intruded by a diabase magma, and later, during rapakivi intrusion, the central part of the diabase crack was intruded by granite porphyry magma (Laitala, 1984; Törnroos, 1984). Paleomagnetism of the Sipoo dykes show that the presumably older diabase dykes record an opposite polarity remanence compared with the younger normal polarity quartz-porphyry dykes that show a NE pointing shallow to moderate inclination component (Mertanen and Pesonen, 1995). Similar behavior was obtained in coeval Melville Bugt dykes (U-Pb, baddeleyite 1622 ± 3 Ma, 1635 ± 3 Ma) in Greenland (Halls et al., 2011), where U-Pb data shows that a NE-directed normal polarity component is younger than the reversed component. Halls et al. (2011) suggest that the NE-directed component may have a steeper negative inclination because of difficulty in removing the strong PEF-like component.

The temporal and compositional overlap of the anorogenic and orogenic magmatism between west Baltica and east Laurentia, as well as the paleomagnetic data

available, suggest that these continents coexisted during the Mesoproterozoic. In the geologically paleomagnetically viable NENA (Gower et al., 1990) connection Baltica is in a somewhat “upside-down” position relative to Laurentia, such that the Timanide margin from northern Norway to the polar Urals restores parallel to the East Greenland margin of Laurentia (Fig. 11). Recent data have appeared to support, within the analytical uncertainties, a single NENA juxtaposition between Baltica and Laurentia between ca. 1750 and ca. 1270 Ma (Salminen and Pesonen, 2007; Evans and Pisarevsky, 2008; Lubnina et al., 2010; Pisarevsky and Bylund, 2010; Evans and Mitchell, 2011; Pesonen et al., 2012), and perhaps lasting as long as ca. 1120 Ma (Salminen et al., 2009; Raiskila et al., 2011). Since the amalgamation of various blocks in Siberia at ca. 1.9 Ga (Pisarevsky et al., 2008) is broadly coeval with the assemblies of Laurentia (Hoffman, 1997) and Baltica (Bogdanova et al., 2005) and the Siberian craton is nearly surrounded by Mesoproterozoic passive margins (Pisarevsky and Natapov, 2003) we favor the proposed tight fit between East Siberia and Western Greenland from 1.9 Ga on (e.g. Rainbird et al., 1998; Wu et al., 2005; Evans and Mitchell, 2011; Ernst et al., 2016; Evans et al., 2016) to form the core of Nuna. This is supported by 1.8 - 1.38 Ga data from Siberia, Baltica and Laurentia (Figs 11 and 12; Table 5), although an alternative looser fit, ca. 1500 km away from the northern margin of Laurentia, has been proposed (Pisarevsky and Natapov, 2003; Pisarevsky et al., 2008).

Selected high-quality paleomagnetic poles for Baltica and Laurentia are rotated in the NENA reconstruction and poles for Siberia are rotated according to the tight fit between East Siberia and Western Greenland (Fig. 11; Tables 5 and 6). In our reconstructions the Siberian craton is first restored to its configuration prior to ~20-25° Devonian rotation in the Vilyuy graben (Pavlov et al., 2008; Evans, 2009; Table 6). Greenland and its poles are rotated to Laurentia after Roest and Sirvastava (1989). In the Fig. 11 the continents are shown on actual paleolatitudes based on new 1.65 Ga paleomagnetic data for Häme dykes. Used Euler rotations are listed in Table 6. Paleomagnetic data position the core of Nuna on

equatorial or on low latitudes at Subjotnian times. In this NENA reconstruction the 1.59 Ga Western Channel diabase pole of Laurentia is in-between 1.65 Ga Häme pole and 1.57 Ga Åland and Satakunta poles, so that its error limits are overlapping with error bars of these poles for Baltica. 1.63 Ga Melville Bugt pole of Greenland (rotated to Laurentia, see Table 6) overlaps with coeval Häme pole and thus supports the reconstruction (Halls et al., 2011; see also discussion in Salminen et al., 2014). For Siberia the 1473 ± 24 Ma paleomagnetic pole from Sololi-Kyutingde (Wingate et al., 2009) is of high quality, whereas the 1503 ± 5 Ma Kuonamka dykes pole (Ernst et al., 2000) fails to show a positive field stability test and is marked by considerable scatter in directions with only one intrusion dated. One recent high quality 1503 ± 2 Ma paleomagnetic pole for West Anabar and one moderate quality 1483 ± 17 Ma pole for North Anabar (Evans et al., 2016; Table 5) support the proposed reconstruction. There are no other data of Nuna interval for Siberia to test how long this connection existed, but coeval 1.38 – 1.27 Ga magmatism at Laurentia, Baltica and Siberia indicates that this core of Nuna started to disperse at 1.38 – 1.27 Ga (Evans and Mitchell, 2011).

5.4.2 Implications to Nuna

Adding cratons around the equatorial core of Nuna (Baltica-Laurentia-Siberia) has been a major initiative among paleogeographers in recent years (e.g., Congo-São Francisco, Salminen et al., 2016; India, Pisarevsky et al., 2013; North China, Zhang et al., 2012, Xu et al., 2014; Australia cratons, Payne et al., 2009; Li and Evans, 2011; Amazonia, Johansson, 2009; Bispo-Santos et al., 2012; D’Agrella-Filho et al., 2012; 2016). Here we briefly discuss the Nuna reconstruction we favor (Fig. 12). Used global paleomagnetic poles are listed in Table 5 and used Euler rotations are listed in Table 6.

The large Congo-São Francisco (C-SF) (e.g. Trompette, 1994; Teixeira et al., 2000) craton has little Mesoproterozoic paleomagnetic data. As Evans (2013) shows, based

on U-Pb ages, the craton experienced prominent magmatic events at 1.73 Ga (Danderfer et al., 2009), 1.50 Ga (Silveira et al., 2013; Ernst et al., 2013), 1.38 Ga (Mayer et al., 2004; Drüppel et al., 2007; Maier et al., 2007, 2013), and 1.11 Ga (Ernst et al., 2013). In spite of C-SF the mafic magmatism of 1.5 Ga is met only in Siberia, suggesting a direct connection of that block with C-SF (Ernst et al., 2013; Evans, 2013). For the Nuna reconstructions in Fig. 12 the C-SF craton is placed adjacent to Baltica close to Siberia so that SW Congo is reconstructed against S-SE Baltica at 1.5 Ga using a recent good quality pole for 1.5 Ga Bahia dykes (Salminen et al., 2016). C-SF has only a few poles for the Nuna interval, but the poor quality 1.38 Ga Kunene pole (Piper, 1974) supports this configuration (Fig. 12; Table 5). We propose that C-SF joined the core of Nuna between 1.65 Ga and 1.5 Ga and that this reconstruction lasted at least until 1.38 Ga as it is also supported by the coeval 1.38 Ga magmatism in Congo craton (Kunene anorthosite complex; Piper, 1974; Mayer et al., 2004; Drüppel et al., 2007; Maier et al., 2013); SE-E Baltica in the Volgo Uralian region (Mashak event; Puchkov et al., 2013); and in the eastern Anabar shield in Siberia (Chieress dykes, Ernst et al., 2000). There are no Mesoproterozoic paleomagnetic data for Kalahari, but in Fig. 12 it is shown juxtaposed with C-SF craton close to its Rodinia connection (Li et al., 2008).

India's position within Nuna varies considerably in the recent literature (e.g. Meert et al., 2011; Zhang et al., 2012; Pisarevsky et al., 2013; Evans, 2013). We show two different position for India. For India 1 (Fig. 12, 1647 Ma reconstruction) the southern part of Baltica with the Archean Dharwar and Sarmatia cratons are shown next to each other (Fig. 12). This reconstruction is supported by the 1.46 Ga pole for Lakhna dykes (Pisarevsky et al., 2013), but not by the 1650 ± 10 Ma Tiruvannamalai pole (Radhakrishna and Joseph, 1996). Using the opposite polarity for Lakha pole the India 2 position is shown in the reconstruction where it is juxtaposed against the southeastern part of Laurentia (juxtaposing 1.65 – 1.55 Ga Mazatzal orogeny in North America and 1.8-1.3 Ga orogeny in northern Dharwar). We favor this position because it leaves enough space for Congo-São Francisco to collide with

Baltica between 1.65 and 1.5 Ga and it is close to India's position in Rodinia as suggested by Li et al. (2008). For India there are no other good quality Mesoproterozoic paleomagnetic poles to test the connection and we keep its relative position to Laurentia -Baltica fixed in 1.65 Ga and 1.5 Ga in the shown reconstructions (Fig. 12, Table 6).

North China is placed in various positions in different Nuna models. India and North China cratons have been linked together by Zhao et al. (2002, 2004) and Hou et al. (2008), either indirectly or directly joined to western Laurentia. We show North China in-between Siberia and C-SF (Fig. 12). Ca. 1.78 Ga, 1.5 – 1.44 Ga, 1.3 Ga paleopoles for North China plot on coeval poles on the APWP of the core of Nuna (Fig. 12) support its rather fixed position juxtaposed to Siberia and C-SF (Fig. 12) as is also suggested based on coeval magmatism on these cratons (Piper, 1974; Mayer et al., 2004; Drüppel et al., 2007; Maier et al., 2013; Puchkov et al., 2014; Ernst et al., 2000).

Australian cratons (western, northern and southern; Myers et al., 1996; Evans, 2009) and Mawsonland (Australian Gawler craton s.s. into Terre Adélie in Antarctica, and possibly as far south as the Transantarctic Mountains near the Miller Range; Goodge et al., 2001; Fitzsimons, 2003; Cawood and Korsch, 2008; Payne et al., 2009; Evans, 2009) are shown in proto-SWEAT juxtaposition with the western margin of Laurentia (Hamilton and Buchan, 2010; Evans and Mitchell, 2011; Pisarevsky et al., 2014; Pehrsson et al., in review). Hamilton and Buchan (2010) demonstrated that this is both geologically and paleomagnetically plausible between 1.75 and 1.59 Ga (Fig. 12; Table 5). Moreover, Goodge et al. (2008) identified matches between 1.4 Ga granitic rocks of the North American autochthon and granitic clasts found within the central Transantarctic Mountains, which could be interpreted in either a SWEAT or proto-SWEAT framework (Evans, 2013). These cratons were still on their way to Nuna at 1.65 Ga (Fig. 12; Table 6).

West Africa and Amazonia formed a rigid continent since the Mesoproterozoic (e.g. Trompette, 1994). The long-lived connection of Amazonia + West

Africa and Baltica (SAMBA) as a coherent entity from 1.8 Ga terrane amalgamation until the late Proterozoic breakup of Rodinia was suggested by Johansson (2009). The reconstruction of Amazonia and Baltica in the SAMBA connection is based on a continuation of Paleo-Mesoproterozoic orogens and magmatic belts from one continent to the other, including distinctive Mesoproterozoic rapakivi granite suites. However this reconstruction was later questioned by Fuck et al. (2008) noting that the Ventuari-Tapajós and Rio Negro-Juruena provinces are truncated by younger Grenville-age orogeny in their northern parts, which questions the continuity of Baltica and Amazonian accretionary belts (Pisarevsky et al., 2014). In our Nuna reconstruction Amazonia + West Africa are kept separate from Nuna as in Pisarevsky et al. (2014). We interpolated between 1789 ± 7 Ma Colider Volcanics paleopole (Bispo-Santos et al., 2008) and 1439 ± 4 Ma Salto do Ceu mafic intrusions poles (D'Agrella-Filho et al., 2016) to reconstruct Amazonia's paleopositions at 1.65 Ga and 1.5 Ga (Fig. 12; Table 5; Table 6).

Geological and paleomagnetic evidence indicate that the assembly of the core of Nuna was fixed at 1.65 Ga and occupied equatorial to low latitudes. Paleomagnetic poles from different Nuna cratons in Fig. 12 (Table 5) against the APWP of the core of Nuna show that the supercontinent was still assembling at 1.65 Ga, however similar geology of Baltica and India supports that India was joined the core of Nuna before 1.65 Ga. Congo-São Francisco and North China were still on their way to Nuna at 1.65 Ga, but they had joined it at 1.5 Ga. Combined Amazonia and West Africa is excluded from our Nuna model. Paleomagnetic evidence support that maximum packing of Nuna occurred at 1.5 Ga (Fig. 12) and dispersal of core is proposed to be associated with coeval 1.38 – 1.27 Ga magmatism in Baltica, Laurentia and Siberia. 1.38 Ga magmatism in C-SF and 1.3 Ga in North China support this dispersal timing as do the diverging younger 1.3 – 1.2 Ga paleomagnetic poles from majority of Nuna continents. Zhang et al. (2012) pointed out that the long break-up time

of Nuna is similar to Rodinia in length: it started at 1.6 – 1.4 Ga as manifested by large igneous provinces and reached its final stage at 1.28 – 1.26 Ga.

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Figure 1. Geological maps of the study areas with the location of sampling sites. Upper map shows the simplified geological map of the whole Southern Finland. Lower map shows the sampling sites for Häme dykes. Ahv. refers to Ahvenisto.

Two columns figure

Figure 2. Concordia plot for baddeleyite LA-MC-ICPMS U-Pb data, A2414 Torittu diabase dyke, Häme, Finland (Table 1).

1.5 columns figure

Figure 3. Mean age of the two analysed baddeleyite fractions from sample A1135 Virmaila dyke, Häme, Finland (Table 2). The data point errors are at 2σ level.

1.5 columns figure

Figure 4. Low and high temperature thermomagnetic curves showing variation in normalized magnetic susceptibility for the Häme dyke swarm samples. Curves were corrected for furnace effects.

Two columns figure

Figure 5. Example of AF and thermal demagnetization of Häme dyke showing primary remanent magnetization direction. Figures show stereographic projection and orthogonal projections of remanent magnetization directions. For stereonet: black (white) represents downward (upward) directions. Orthogonal projections: black (white) represents horizontal (vertical) plane.

Two columns figure

Figure 6. Positive baked contact test for samples for the dated Virmaila dyke (Häme swarm), and its baked and unbaked host rock. (a) Dyke sample, (b) Baked host ca. 20 m from the contact, and (c) Unbaked host 2.5 km from contact. Symbols as in Fig. 5.

Two columns figure

Figure 7. Site mean paleomagnetic directions for primary remanent magnetization (left); for secondary remanent magnetization (B component, middle); for magnetization of unknown origin. Closed (open) symbols represent downward (upward) directions, black stars represent Present Earth Field (PEF) direction on the sampling area.

Two columns figure

Figure 8. Vector analyses to study the effect of an unremoved secondary component for asymmetric Subjotnian dual polarity paleomagnetic data for Baltica, subtracting the measured secondary component (component B) from the obtained normal and reversed polarity characteristic remanent magnetization (ChRM) directions (ChRM (N) and ChRM (R) for three cases: Åland (Salminen et al., 2015), Satakunta (Salminen et al., 2014), and Häme dyke (this work) swarms.

Two columns figure

Figure 9. Global dual polarity paleomagnetic directions (Table 4). Colors refer to continents as explained in the legend.

1.5 columns figure

Figure 10. Inclination (I) vs. paleolatitude (λ) for Subjotnian data according to hemispheric constraints given in Table 4. Colors refer to continents as follows: magenta for Amazonia, green for Baltica, purple for India, cyan for Laurentia, orange for South Australia, red for North China and gray for Siberia. Error bars for inclination are also given. White symbols refer to directions with too few data for calculation of α_{95} error, regardless of continent. Solid line indicates the relation of I vs. λ in GAD field. A geomagnetic field model with a standing geocentric axial dipole and a reversing geocentric axial quadrupole ($G_2=\pm 0.25$) is also given. Dot-dashed line indicates the situation where g_{20} is parallel to GAD, and dashed line indicates the situation where it is antiparallel. All data fulfill the $Q1-7 \geq 3$ quality criterion for both polarities.

Two columns figure

Figure 11. Apparent polar wander path for the core of Nuna (see Table 5 for codes and Table 6 for used Euler rotations). The key poles are shown using full colors with black outline. Non-key poles are shown with transparent colors. Separate N- and R-polarity poles for Häme are shown with transparent color and with dashed outline.

1.5 columns figure

Figure 12. Nuna reconstructions. Upper left: reconstruction at 1647 Ma; upper right: poles for different cratons color coded with craton against APWP of the Nuna core (dashed grey). Poles are rotated together with the continents. At 1647 Ma India is shown in both options 1 and 2 (India 1, 2). Lower left: reconstruction at 1500 Ma (maximum packing on Nuna); lower right: respective poles for different cratons. Poles are rotated together with the continents. At

1500 Ma India is shown in option 2 (India 2). Codes for poles in Table 5 and used Euler rotations in Table 6. For pole figure the key poles are shown using non-transparent colors with black outline.

Two columns figure

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Research highlights

- A new paleomagnetic pole at 23.6°N, 209.8°E has been obtained for Häme dykes
- Precise U-Pb ages of 1642 ± 2 Ma and 1647 ± 14 Ma for two reversely magnetized dykes are acquired.
- The pole is used to fix the core of the Nuna supercontinent on shallow latitudes
- The Geocentric Axial Dipole model of the geomagnetic field is supported at 1.7 – 1.4 Ga

Table 1. LA-MC-ICPMS U-Pb baddeleyite (euhedral) data for diabase sample of Torittu dyke, Finland.

Sample/ spot #	²⁰⁷ Pb ²⁰⁶ Pb	±σ	²⁰⁷ Pb ²³⁵ U	±σ	²⁰⁶ Pb ²³⁸ U	±σ	²⁰⁷ Pb ²⁰⁶ Pb	±σ %	²⁰⁷ Pb ²³⁵ U	±σ %	²⁰⁶ Pb ²³⁸ U	±σ %	ρ ¹⁾	Disc. % ²⁾	U ppm	²⁰⁶ Pb ppm	²⁰⁶ Pb/ ²⁰⁴ Pb	f ₂₀₆ % ³⁾
H17F_bad-01a	1637	23	1643	45	1649	78	0.1007	0.0013	4.045	0.224	0.2914	0.0157	0.97	0.8	191	50	9.51E+03	0.00
H17F_bad-02a	1520	43	1626	52	1710	90	0.0946	0.0023	3.962	0.256	0.3038	0.0183	0.93	14.2	155	42	1.98E+03	1.90
H17F_bad-03a	1623	11	1623	47	1622	84	0.1000	0.0006	3.944	0.231	0.2862	0.0167	1.00	-0.1	219	52	1.01E+04	0.14
H17F_bad-03b	1599	70	1576	57	1558	83	0.0987	0.0037	3.720	0.263	0.2735	0.0164	0.85	-2.9	139	33	1.71E+03	2.20
H17F_bad-03c	1703	26	1664	52	1633	89	0.1043	0.0015	4.147	0.263	0.2883	0.0178	0.97	-4.6	116	28	4.87E+03	0.64
H17F_bad-03d	1658	19	1647	64	1639	113	0.1019	0.0011	4.066	0.320	0.2895	0.0225	0.99	-1.3	219	52	1.35E+04	0.00
H17F_bad-04a	1663	41	1672	64	1680	110	0.1021	0.0023	4.192	0.327	0.2977	0.0222	0.96	1.1	619	149	1.60E+04	0.56
H17F_bad-04b	1678	13	1701	65	1720	119	0.1030	0.0008	4.341	0.343	0.3058	0.0241	1.00	2.8	708	173	2.93E+04	0.01
H17F_bad-04c	1653	37	1676	72	1694	127	0.1016	0.0021	4.209	0.368	0.3006	0.0255	0.97	2.8	592	139	4.58E+04	0.89
H17F_bad-05a*	1631	77	1798	91	1944	168	0.1004	0.0043	4.873	0.529	0.3521	0.0352	0.92	22.3	200	53	5.60E+03	2.6
H17F_bad-06a*	1477	40	1287	56	1176	79	0.0925	0.002	2.551	0.196	0.2002	0.0148	0.96	-22.2	717	125	4.78E+02	3.5

All errors are in 1 sigma level. * rejected. Sample was assigned a code A2414 in GTK database. 1) Error correlation in conventional concordia space. 2) Age discordance at closest approach of 1 sigma error ellipse to concordia. 3) Percentage of common ²⁰⁶Pb in measured ²⁰⁶Pb, calculated from the ²⁰⁴Pb signal assuming a present-day Stacey and Kramers (1975) model terrestrial Pb-isotope composition.

Table 2. ID-TIMS U-Pb data on baddeleyite, diabase sample of Virmaila dyke, Finland.

Sample information Analysed mineral and fraction	Sample	U	Pb	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁸ Pb/ ²⁰⁶ Pb	ISOTOPIC RATIOS ¹⁾							APPARENT AGES (Ma±2σ)		
	mg	ppm	ppm	measured	radiogenic	²⁰⁶ Pb/ ²³⁸ U	±2σ%	²⁰⁷ Pb/ ²³⁵ U	±2σ%	²⁰⁷ Pb/ ²⁰⁶ Pb	±2σ%	ρ	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb
VR (A1135)															
Badd.#1 abr 30 min	0.42	246	77	1121	0.11	0.2856	0.24	3.9773	0.28	0.1010	0.11	0.92	1620	1630	1642±2.0
Badd.#2 abr 30 min	0.25	150	47	812	0.09	0.2882	0.38	4.0094	0.41	0.1009	0.13	0.95	1633	1636	1641±2.4

1) Isotopic ratios corrected for fractionation, blank (30 or 50 pg), and age related common lead (Stacey and Kramers, 1975; ²⁰⁶Pb/²⁰⁴Pb±0.2, ²⁰⁷Pb/²⁰⁴Pb±0.1, ²⁰⁸Pb/²⁰⁴Pb±0.2). 2) ρ:Error correlation between ²⁰⁶Pb/²³⁸U and ²⁰⁷Pb/²³⁵U ratios. All errors are 2 sigma (σ).

Table 3. Paleomagnetic results for Häme dykes, Finland.

Site	Site name	Age (Ma)	Str (°)	Dip (°)	w (m)	Lat (°N)	Lon (°E)	D (°)	I (°)	a95	k	pol	Plat (°N)	Plon (°E)	A95	K	B/N
H1	Orivesi		090	90	5	61.660	24.258	015.4	-35.7	14.9	69.4	N	07.4	189.9			3
H12	Hirtniemi		275	90	12	61.340	25.395	031.8	-02.7			N	22.8	170.5			2
H15	Harmoistenkaivo		320	90	45	61.490	25.125	342.7	-07.9	14.4	22.7	N	24.2	219.9			6
H18	Tuomasvuori		315	-	50	61.361	25.290	028.8	-02.4			N	24.5	169.0			2
H25	Muorinkallio		310	-	80	61.420	24.987	018.7	04.1	10.4	34.7	N	29.9	179.3			7
	Mean for N-polarity dykes: Häme							015.8	-09.2	24.3	10.8	N	22.4	187.9	20.6	14.8	5*/20
H13	Hirtniemi		315	90	60	61.340	25.410	148.9	-03.7	27.9	8.5	R	26.2	240.8			5
H17	Torittu	1647 ± 14	270	90	60	61.452	25.143	151.4	-00.4			R	25.0	237.1			2
H20	Koukkujärvi		290	90	3	61.441	25.217	148.7	11.7	29.6	7.6	R	17.8	238.5			5
H23	Partakorpi		290	-	30	61.457	25.055	169.2	22.9	19.8	22.5	R	15.5	216.3			4
H24	Romo		310	-	50	61.404	25.032	167.8	30.4	13.2	34.4	R	11.5	216.7			5
VR+HR	Virmaila	1642 ± 2	275	90	60	61.428	25.313	171.6	-12.6	38.2	6.8	R	34.8	215.9			4
	Mean for R-polarity dykes: Häme							159.3	08.1	16.7	17.0	R	22.3	227.4	11.7	33.5	6*/25
		1642 ± 2; 1647 ± 14						355.6	-09.1	16.6	8.6	M	23.6	209.8	14.7	10.6	11*/42
		GRAND MEAN: Häme															
<i>Häme baked host rock</i>																	
H8 and H9	Orvokkila, Virmaila (for site VR+HR)					61.429	25.312	153.0	17.8	18.3	11.8	R	15.2	233.9			7
<i>Unbaked host rock</i>																	
H1	Orivesi					61.660	24.258	333.2	45.2				50.7	243.8			1
VR+HR	Virmaila							340.6	48.8				56.0	235.9			2

Häme diabase dyke excluded from mean calculations

H2	Orivesi		100	-	1.5	61.663	24.265	UNSTABLE						
H3	Kasiniemi/Ansio	1646 ± 6	120	-	20	61.453	24.899	325.6	14.8	17.0	30.2	30.6	245.1	4
H4	Kasiniemi/Ansio	1646 ± 6	120	-	30	61.440	24.919	UNSTABLE						
H6	Karivuori		100	-	70	61.494	24.782	UNSTABLE						
H7	Vehkajärvi		100	90	3.1	61.507	24.837	311.4	16.6	37.3	12.0	26.4	261.1	3
H11	Hirtniemi		315	90	6.0	61.334	25.399	017.1	38.2	17.6	50.4	48.4	181.0	3
H16	Torittu		100	90	0.6	61.452	25.143	183.3	66.1	13.2	26.7	-22.5	204.1	6
H19	Koukkujärvi		310	-	24	61.442	25.218	165.8	44.1	11.0	126.2	01.9	218.1	3
H21	Koukkujärvi		40	45	0.5	61.442	25.217	UNSTABLE						
H22	Koukkujärvi		290	90	12	61.441	25.218	169.5	43.8			00.7	211.2	2

str/dip., Structural orientation (strike and dip) of dykes; w, width of dykes; Lat/Lon, site latitude and longitude; D, declination; I, inclination; a95, the radius of the 95% confidence cone in Fisher (1953) statistics. k, Fisher (1953) precision parameter; pol., polarity of the isolated direction: N(R), normal (reversed) polarity; Plat/Plong, pole's latitude/longitude of the pole. A95, radius of the 95% confidence cone of the pole; K, Fisher (1953) precision parameter of pole. (B)/N, number of analyzed (sites)/samples; * The number used to calculate mean value.

Table 4. Paleomagnetic data of dual-polarity formations of the Mesoproterozoic (1.4 – 1.7 Ga).

Rock unit	Code	Slat	Slon	Age Ma	Age ref	P	B	N	D	I	a95	k	λ	Plat	Plon	A ₉₅	RT	Q ₁₋₇	HS	Pmag ref
Amazonia																				
Nova Guarita dykes	Ng	-10.3	304.7	1418.5±3.5 (Ar-Ar, biot.) Used: 1419	Bispo-Santos et al. 2012	C	19*	268	220.5	45.9	6.5	27.7	27.3	-47.9	245.9	6.6	I	7	N	Bispo-Santos et al. 2012
						N	2*	29	030.3	-41.3			-23.7	-58.1	243.6			5		
						R	17*	239	221.8	46.3	7.0	27.2	27.6	-47.1	244.5	7.1		6		
Salto de Ceu intrusives	Sd	-15.2	301.9	1439±4 (U-Pb, bad.) Used: 1439	D'Agrella- Filho et al. 2016	C	18*	161	197.4	62.9	5.7	38.0	44.3	-56.0	278.5	7.9	C	6	N	D'Agrella-Filho et al. 2016
						N	5*	31	015.8	-54.8	6.2	152.0	-35.3	-65.5	269.5	7.8		5		
						R	13*	130	198.3	66.0	7.2	34.0	48.3	-52.2	280.7	10.3		5		
Baltica																				
Sortavala B dyke RA+RI	Sv	61.7	030.7	1452±12 (U-Pb, bad.) Used: 1452	Lubnina et al. 2010	C	2	24*	028.6	-23.3	5.6	28.9	-12.2	14.4	182.2	4.2	C	5	S	Lubnina et al. 2010
						N	1	19*	029.9	-21.1	6.0	32.0	-10.9	13.9	180.4	4.6		4		
						R	1	5*	202.8	29.7	15.5	25.2	15.9	10.5	188.4	12.8		4		
Salmi basalts	Sl	61.4	031.9	1499±68 (Sm-Nd) Used: 1499	Bogdanova et al. 2003	C	2	25*	214.1	15.5	6.7	19.5	7.9	15.8	176.6	4.9	C	5	S	Lubnina et al. 2010
						N	1	3*	047.5	-24.1	19.0	43.1	-12.6	07.1	165.0	14.9		3		
						R	1	22*	212.5	13.6	7.1	20.1	6.9	17.2	178.0	5.2		3		
Ragunda group 1 dykes	Rd	63.3	016.1	1505±12; 1514±5 (U-Pb, zr) Used: 1506	Persson 1999	C	12*	64	001.7	-23.5	14.3	10.0	-12.3	60.4	241.2	10.8	I	5	S	Piper 1979a
						N	3*	17	001.0	-32.2	30.0	17.9	-17.5	55.7	242.0	22.3		3		
						R	9*	47	181.9	20.4	18.6	8.6	10.5	62.1	240.9	14.2		4		
Ragunda Fm.	Rf	63.3	016.1	1514±5; 1505±5 (U-Pb, zr) Used: 1514	Persson 1999	C	15*	87	014.5	45.7	6.9	31.0	27.1	51.6	166.6	7.1	C	4	S	Piper 1979b
						N	11*	61	018.2	44.9	8.9	28.0	26.5	51.3	169.5	8.9		3		
						R	4*	26	201.6	-48.2	17.5	29.0	-29.2	53.2	163.7	18.5		3		
Satakunta dykes N-S & NE-SW	Sk	62	021.5	1575.9±3 ^a (U-Pb, zr) Used: 1576	Salminen et al. 2015	C	13*	77	010.8	3.8	11.9	13.1	1.9	29.5	188.8	8.8	F	7	S	Salminen et al. 2014
						N	9*	64	013.9	11.1	12.5	18.1	5.6	33.2	184.6	10.4		6		
						R	4*	14	183.8	12.9	25.0	14.5	6.5	20.8	197.2	15.6		5		
Åland dykes	Ål	60.0	020.0	1575.9±3.0 (U-Pb, zr) Used: 1576	Salminen et al. 2015	C	96*	366	008.6	-11.0	4.0	13.7	-5.6	23.7	191.4	2.9	F	7	S	Salminen et al. 2015
						N	40*	150	012.4	10.5	6.9	12.9	5.3	34.6	185.5	4.9		6		
						R	56*	216	186.2	24.9	3.5	31.1	13.1	16.2	194.2	2.8		6		

Nordingrå granite ^d	Nr	62.9	018.6	1578±19 (U-Pb, zr) Used: 1578	Welin and Lundqvist 1984	C	28*	28	221.3	-25.1	13.3	5.0	-13.2	2.5	149.0	10.5	I	3	S	Piper 1980
						N	9*	9	34.5	31.7	17.4	10.0	17.2	38.4	154.9	14.6		2		
						R	19*	19	229.7	-21.5	16.5	5.0	-11.1	27.5	141.1	12.7		3		
Dala sandstone ^d	Dl	61.1	013.2	1530-1630 (APWP) Used: 1580	Piper and Smith 1980	C	6*	27	014.1	10.0	13.4	26.0	5.0	32.9	176.4	9.7	I	4	S	Piper and Smith 1980
						N	3*	12	007.4	6.9	29.1	19.0	3.5	32.1	184.5	20.8		2		
						R	3*	15	200.8	-12.9	17.8	49.0	-6.5	33.3	168.2	12.9		2		
Satakunta sandstone	Ss	61.2	022.0	1450-1650 (APWP) Used: 1600	Klein et al. 2014	C	13*	37	025.2	3.9	9.0	29.0	2.0	28.0	173.0	7.0	I	5	S	Klein et al. 2014
						N	10*	30	023.7	2.6	12.0	20.0	1.3	28.0	175.0	8.0		4		
						R	3*	7	213.7	-15.6	23.0	31.0	-7.9	31.0	162.0	17.0		3		
Sipoo dykes	Sp	60.3	025.3	1633* Used: 1633	Mertanen and Pesonen 1995	C	5*	21	197.5	-7.3	28.3	8.3	-3.7	31.8	184.6	20.2	I	4	S	Mertanen and Pesonen 1995
						N	1	3*	015.5	9.9	52.9	6.5	5.0	33.4	186.8	38.0		3		
						R	4*	18	198.0	-6.5	40.1	6.2	-3.3	31.6	183.6	28.5		3		
Häme	Hm	61.4	025.1	1646±6 1642±2; 1647±14 (U-Pb, bad.) Used: 1644	this work	C	11*	42	355.6	-09.1	16.6	8.6	-4.6	23.6	209.8	14.7	F	7	N	This study
						N	5*	20	015.8	-09.2	24.3	10.8	-4.6	22.4	187.9	20.6		3		
						R	6*	25	159.3	08.1	16.7	17.0	4.1	22.3	227.4	11.7		6		
India																				
Lakhna dykes	Lk	20.8	082.7	1466.4±2.6 (U-Pb, zr) Used: 1466	Pisarevsky et al. 2013	C	10*	79	048.5	68.7	13.2	14.4	52.1	41.3	120.5	20.5	F	6	N	Pisarevsky et al. 2013
						N	8*	67	47.2	63.1	11.0	26.1	44.6	44.6	129.9	15.4		5		
						R	2*	12	009.6	-85.9			-81.8	12.7	081.3			4		
Laurentia																				
McNamara Fm.	Mc	46.9	246.4	1401±6 (U-Pb, zr) Used: 1401	Evans et al. 2000	C	10*	195	217.2	36.6	7.5	42.8	20.4	-13.5	208.3	6.7	C	7	N	Elston et al. 2002
						N	3*	37	228.7	40.8	13.6	83.1	23.3	-07.2	202.4	12.9		6		
						R	5*	90	032.9	-35.6	9.0	55.9	-19.7	-16.6	213.9	9.6		6		
Upper Belt-Purcell supergroup	Pu	49.4	245.5	1401±6; 1443±7 (U-Pb, zr) ^c	Evans et al. 2000	C	15	44	208.3	34.7	6.2	39.4	19.1	-17.1	217.8	5.4	C	5	N	Evans et al. 1975
						N	5*	15	212.6	35.6	8.1	90.6	19.7	-15.1	213.8	7.1		3		

Mt. Shields Fm., upper member	Mt	48.1	245.8	Used: 1423 1401±6 and 1443±7 (U-Pb, zr) ^c Used: 1425	Evans et al. 2000	R	10*	29	026.1	-34.3	8.9	30.4	-18.8	-17.9	219.5	7.7		4	Elston et al. 2002
						C	14*	742	221.8	29.7	5.3	74.1	15.9	-15.6	202.3	4.3		7 N	
						N	6*	106	219.7	35.6	4.8	192.6	19.7	-12.6	207.5	4.3	B	6	
						R	8*	636	043.3	-27.8	6.7	68.6	-14.8	-16.5	201.8	5.4		6	
Sherman granite	Sh	41.2	254.6	1431±6; 1412±13 (U-Pb, zr) Used: 1425	Aleinikoff 1983	C	12*	79	238.3	48.7	7.1	38.9	29.6	-01.1	207.2	9.0		5 N	Harlan et al. 1994
						N	2*	18	226.2	46.3			27.6	-9.0	214.2	10.1	C	3	
						R	10*	61	060.9	-49.1	7.9	38.6	-30.0	00.6	205.8	10.1		4	
Tobacco Root dykes A	Tr	47.4	247.6	1448±49 Sm-Nd Used: 1448	Harlan et al. 2008	C	14*	157	225.0	61.8	7.7	27.9	43.0	08.7	216.1	10.5		5 N	Harlan et al. 2008
						N	6*	80	248.7	63.5	8.0	71.6	45.1	19.6	204.0	11.3	F	4	
						R	8*	77	030.0	-58.2	9.7	33.8	-38.9	00.3	222.8	12.3		4	
Harp Lake complex	Hl	55.0	297.7	1450 Used: 1450	Krogh and Davis 1973	C	24*	110	273.0	0.8	6.1	25.0	0.4	01.6	206.3	4.3		5 N	Irving et al. 1977
						N	10*	50	272.1	8.5	7.0	49.0	4.3	04.7	208.4	5.0	F	4	
						R	14*	60	093.7	4.8	8.7	22.0	2.4	-00.2	203.3	6.2		4	
Snowslip Fm.	Sn	47.9	245.9	1443±7; 1454±9 (U-Pb, zr) Used: 1457	Evans et al. 2000	C	9*	295	210.9	20.6	4.6	128.8	10.6	-24.9	210.2	3.5		7 N	Elston et al. 2002
						N	2*	49	225.0	27.8			14.8	-16.3	201.3	29.6	C	5	
						R	9*	246	032.1	-20.6	5.5	81.9	-10.6	-24.7	259.2	4.1		6	
Melville Bugt dykes	Mb	75.0	305.0	1622±3; 1635±3 (U-Pb, bad.) Used: 1630	Halls et al. 2011	C	9*	54	213.1	31.2	11.7	20.4	16.9	05.0	273.7	8.7		6 N	Halls et al. 2011
						N	4*	27	211.0	42.9	11.4	65.4	24.9	12.2	276.6	10.9	F	5	
						R	5*	27	034.4	-21.7	17.3	20.5	-11.2	-00.8	271.5	13.3		5	
North Australia																			
Mt. Isa dykes ^d	Mi	-20.6	139.7	1500-1550 (APWP) Used: 1525	Tanaka and Idnurm 1994	C	9*	70	186.1	49.0	6.9	56.5	29.9	-79.0	110.6	6.7		3 N	Tanaka and Idnurm 1994
						N	3*	25	356.3	-45.6	12.5	98.9	-27.0	-81.1	162.2	12.7	C	2	
						R	6*	45	191.6	50.4	8.7	60.1	31.1	-75.1	097.4	9.6		2	
South Australia																			
Pandurra Fm.	Pn	-33.0	137.6	1440 (APWP) Used: 1440		C	18*	90	249.2	45.2	5.3	44.2	26.7	-33.6	064.5	5.4		5 N	Schmidt and Williams 2011
						R	12*	61	251.9	42.3	6.6	44.6	24.5	-27.6	060.2	6.4	F	4	
						N	6*	29	062.2	-52.1	8.5	63.2	-32.7	-38.5	065.4	9.6		4	

North China

Yangzhuang Fm. (Jixian)	Yz	40.2	117.6	1437±21; 1560±5 ^c (U-Pb) Used: 1499	Wu et al. 2005	C	15*	112	072.2	11.5	7.9	24.6	5.8	17.3	214.5	5.7		6	S	Wu et al. 2005
						N	7*	44	071.6	19.1	9.9	37.9	9.8	20.3	211.7	7.5	C	5		
						R	8*	68	252.7	-4.8	11.7	23.5	-2.4	14.7	216.8	8.3		5		

Siberia

Fomich River mafic intrusions	Fi	71.5	106.5	1513±51 (Sm-Nd) Used: 1513	Veselovskiy et al. 2006	C	1	52*	27.0	5.6	5.9	12.3	2.8	19.2	257.8	4.2	F	6	S	Veselovskiy et al. 2006
						N	1	40*	024.6	7.8	5.4	19.5	3.9	20.6	260.2	3.8		5		
						R	1	12*	216.0	2.8	19.5	5.9	1.4	13.5	249.3	14.8		4		

Code, code used in Fig. 9.; Slat/Slong, site latitude/longitude; Age, age of rock unit (U-Pb, zr – zircon, bad. – baddeleyite, biot. – biotite; APWP – interpreted from apparent polar wander path); P, polarity of the isolated direction: N/R/C, normal/reversed/combine polarity; * level of average; B, number of sites. N, number of samples; D, declination; I, inclination; a95, the radius of the 95% confidence cone in Fisher (1953) statistics. k, Fisher (1953) precision parameter; λ, paleolatitude; Plat/Plong, pole’s latitude/longitude of the pole. A95, radius of the 95% confidence cone of the pole; RT, McFadden and McElhinny (1990) reversal test: pass (classification B/C/I), fail (F); Q₁₋₇, Van der Voo (1990) reliability criteria for paleomagnetic data; HS, hemisphere of site, based on reconstructions (Fig. 12.).

^aage not from quartz porphyry dyke
^bage from Åland dykes
^cUnit is stratigraphically between the dated intrusions
^dOmitted from final GAD model

Table 5. Selected global 1.79 – 1.3 Ga paleomagnetic poles.

Code	Rock unit	Plat (°N)	Plong (°E)	A ₉₅ (°)	1234567 Q	Age (Ma)	References
BALTICA-FENNOSCANDIA							
B1	Hoting babbro	43.0	233.3	10.9	1111001 5	1786 ± 3.0	Elming et al. 2009
B2	Småland intrusives	45.7	182.8	8.0	1111110 6	1780 ± 3, 1776 +8/-7	Pisarevsky and Bylund 2010
B3H	Häme	23.6	209.8	14.7	1111111 7	1642 ± 2; 1647 ±14	This work
	Häme N-polarity	22.4	187.9	20.6	0010101 3		This work
	Häme R-polarity	22.3	227.4	11.7	1111101 6	1642 ± 2; 1647 ± 14	This work
B4	Sipoo Quartz porphyry dykes	26.4	180.6	9.4	1110101 5	1633 ± 10	Mertanen and Pesonen 1995
B5	Quartz porphyry dykes	30.2	175.4	9.4	1010101 4	1631	Neuvonen 1986
B6	Åland dykes	23.7	191.4	2.8	1111111 7	1575.9 ± 3.0	Salminen et al. 2015
B7	Satakunta N-S & NE-SW dykes	29.3	188.1	6.6	1111111 7	1575.9 ± 3.0	Salminen et al. 2014
B8	Bunkris-Glysjön-Öje dykes	28.3	179.8	13.2	1010101 4	1469 ± 9	Pisarevsky et al. 2014
B9	Lake Ladoga mafic rocks	11.8	173.3	7.4	1111110 6	1452 ± 12, 1459 ± 3, 1457 ± 2	Salminen and Pesonen 2007; Shcherbakova et al. 2008; Lubnina et al., 2010
B10	Mashak suite	01.8	193.0	14.8	1011111 6	1386 ± 5; 1386 ± 6	Lubnina et al. 2009
B11	Mean Post Jotnian intrusions	04.0	158.0	4.0	1111101 6	1265	Pesonen et al. 2003
B12	Mean Post Jotnian intrusions	-01.8	159.1	3.4	1111101 6	1265	Pisarevsky et al. 2014
AMAZONIA							
A1	Avanavero mafic rocks	48.4	207.5	9.2	1111101 6	1785.5 ± 2.5	Bispo-Santos et al. 2014
A2	Colider volcanics	63.3	118.8	11.4	1110111 6	1785 ± 7	Bispo-Santos et al. 2008
A3	Salto do Ceu mafic intrusions	56.0	081.5	7.9	1111111 7	1439 ± 4	D'Agrella-Filho et al. 2016
A4	Nova Guarita dykes	47.9	065.9	6.6	1111111 7	1418.5 ± 3.5	Bispo-Santos et al. 2012
A5	Indiavaí gabbro	57.0	069.7	8.9	1110001 4	1416 ± 7	D'Agrella-Filho et al. 2012 Tohver et al. 2002

LAURENTIA+ GREENLAND							
L1	Cleaver dykes	19.4	276.7	6.1	1111101 6	1740 +5/-4	Irving et al. 2004
L2	Melville Bugt dykes ^a	02.7	261.6	9.0	1110111 6	1622 ± 3, 1635 ± 3	Halls et al. 2011
L3	Western Channel diabase dykes	09.0	245.0	7.0	1101101 5	1592 ± 3 1590 ± 4	Irving et al. 1972 Hamilton and Buchan 2010
L4	St. Francois Mnt acidic rocks	-13.2	219.0	6.1	1111101 6	1476 ± 16	Meert and Stuckey 2002
L5	Michikamau intrusion	-01.5	217.5	4.7	1111011 6	1460 ± 5	Emslie et al. 1976
L6	Mean Rocky Mnt intrusion	-11.9	217.4	9.7	1110011 5	1430 ± 15	Elming and Pesonen 2010
L7	Purcell lava	-23.6	215.6	4.8	1111101 6	1443 ± 7	Elston et al. 2002
L8	Spokane Formation	-24.8	215.5	4.7	1111101 6	1457	Elston et al. 2002
L9	Zig-Zag Dal intrusions ^a	13.3	209.3	3.8	1111111 7	1382 ± 2	Evans and Mitchell 2011
L10	Mackenzie dykes	04.0	190.0	5.0	1111101 6	1267 ± 2	Buchan and Halls 1990; Buchan et al. 2000
L11	Sudbury dykes	-02.5	192.8	2.5	1111101 6	1235 +7/-3; 1238 ± 4	Palmer et al. 1977
SIBERIA							
S1	Lower Akitkan	30.8	278.7	3.5	1111101 6	1878±4	Didenko et al. 2009
S2	Upper Akitkan	22.1	277.5	5.2	1111101 6	1863±9	Didenko et al. 2009
S3	Kuonamka dykes	06.0	234.0	19.8	1010100 3	1503 ± 5	Ernst et al. 2000
S4	West Anabar intrusions	25.3	241.4	4.6	1111101 6	1503 ± 2	Evans et al. 2016
S5	North Anabar intrusions	23.9	255.3	7.5	1110101 5	1483 ± 17	Evans et al. 2016
S6	Sololi - Kyutinge intrusions	33.6	253.1	10.4	1111100 5	1473 ± 24	Wingate et al. 2009
S7	Chieress dykes	05.0	258.0	6.7	1010100 3	1384 ± 2	Ernst and Buchan 2000
CONGO-SF							
C1	Curaçá dykes ^b	41.6	041.4	15.8	1111100 5	1506.7 ± 6.9	Salminen et al. 2016
C2	Kunene Anorthosite complex	03.3	075.3	18.0	1000010 2	1371 ± 2.5,	Piper 1974

1376 ± 2

NORTH CHINA BLOCK							
N1	Yinshan dykes	35.5	245.2	2.4	1111101 6	1780	Xu et al. 2014
N2	Xiong'er Group	50.2	263.0	5.4	1110111 6	1780 ± 10	Zhang et al. 2012
N3	A1 dykes	44.1	247.4	15.4	1111111 7	Ca. 1780	Piper et al. 2011
N4	A2 dykes	54.6	283.4	6.5	1111101 6	Ca. 1780	Piper et al. 2011
N5	Taihang dykes	36.0	247.0	4.2	1111111 7	1769 ± 2.5	Halls et al. 2000
N6	TH + XR	41.6	246.3	4.3	1111111 7	Ca. 1780	Xu et al. 2014
N7a	Yangzhuang formation	17.3	214.5	5.7	0111111 6	1437 ± 21; 1560 ± 5	Wu et al. 2005
N7b	Yangzhuang formation	2.4	190.4	11.9	0111111 6	1437 ± 21; 1560 ± 5	Pei et al. 2006
N8	Yanliao (Liaoning) mafic sill	-5.9	179.6	3.6	1111101 6	1323 ± 4	Chen et al. 2013
NORTH AUSTRALIA							
Au1	Peters Creek volcanics	-26.0	221.0	4.8	1111111 7	1725-1729	Idnurm 2000
Au2	Fiery Creek Frm.	-23.9	211.8	10.4	1110101 5	1709 ± 3	Idnurm 2000
Au3	West Branch Volcanics	-15.9	200.5	11.3	1111101 6	1709 ± 3	Idnurm 2000
Au4	Mallapunyah Frm.	-35.0	214.3	3.1	1111111 7	1645-1665	Idnurm et al. 1995
Au5	Tooganine	-61.0	186.7	6.1	1111110 6	1648 ± 3	Idnurm et al. 1995
Au6	Emmerugga dolomite	-79.1	202.6	6.1	1111101 6	1645	Idnurm et al. 1995
Au7	Balbirini dolomite (lower)	-66.1	177.5	5.7	1110111 6	1613 ± 4	Idnurm 2000
Au8	Balbirini dolomite (upper),	-52.0	176.1	7.5	1110110 5	1589 ± 3	Idnurm 2000
SOUTH AUSTRALIA							
Au9	Gawler Range volcanics	-60.4	50.0	6.2	0110100 3	Ca. 1590	Chamalaun et al. 1978
INDIA							
I1	Tiruvannamalai	-19.0	235.0	9.0	1110100 4	1650 ± 10	Radhakrishna and Joseph 1996

I2	Lakhna	41.3	120.5	20.5	1110111 6	1466.4 ± 2.6	Pisarevsky et al. 2013
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Code – code in Figures 11 and 12; Plat – pole latitude; Plong – pole longitude; A95 – 95% confidence circle of the pole; Q – van der Voo (van der Voo, 1995) quality factor. a – Greenland rotated to Laurentia reference frame using Euler parameters (67.5°, 241.5°, -013.8°) from Roest and Sirvastava (1989). b – São Francisco rotated to Congo reference frame using Euler parameters (46.8°, 329.4°, +055.9°) from McElhinny et al. (2003).

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Table 6. Euler rotation poles (Elat, Elong, Erot) for Nuna reconstruction at 1.64 Ga and 1.5 Ga.**1.64 Ga**

Laurentia to absolute: (37.94 -151.15 -236.51), this work (based on Häme pole)
 Greenland to Laurentia: (67.5, 241.5, -013.8) after Roest and Sirvastava (1989)
 Baltica to Laurentia: (47.5, 1.5, +49), after Evans and Pisarevsky (2008)
 Amazonia to absolute: (8.95, 9.95, -146.27), interpolating using Colider volcanics (Bispo-Santos et al. 2008) pole
 West Africa to Amazonia: (30.27, -25.42, -88.65)
 Congo-São Francisco (C-SF) to Laurentia: (-7.59, -123.26, -233.65), interpolating using Bahia 1.5 Ga pole (Option A, Salminen et al. 2016)
 SF to Congo: (46.8, 329.4, +055.9°) from McElhinny et al. (2003)
 Kalahari to C-SF: (8.9, 50.3, -69.8) close to Rodinia connection (Li et al. 2008)
 India to Baltica: (30.7, 54.3, +175.1) India 1 option, Dharwar – Sarmatia juxtaposition (Pisarevsky et al. 2013)
 India to Baltica: (14.87, 148.09, +243.28) India 2 option, close to Rodinia position
 North China to Laurentia: (54.25, 117.73, -159.63), interpolating between group of ca. 1.79 Ga and Yangzhuang formation (Wu et al. 2005) poles
 North Australia to Laurentia:
 (39.76, 100.0, +102.59) after Payne et al. (2009); Evans and Mitchell (2011)
 Mawson block to North Australia:
 South Australia to North Australia (-20.0, 135.0, +40.0), after Li and Evans, 2011
 West Australia to North Australia (-20.0, 135.0, +40.0), after Li and Evans, 2011
 Mawson block, E Antarctica to present Australia (-2, 38.9, +31.5), after Powell and Li (1994)
 Siberia to Laurentia:
 Anabar block to Laurentia (77.0, 98.0, +137.0), after Evans et al. (2016)
 Aldan to Anabar before Devonian (60, 115, +25), after Evans (2009)

1.5 Ga

Laurentia to absolute: (38.8, -151.5, -200.3), this work
 Greenland to Laurentia: (67.5, 241.5, -013.8) after Roest and Sirvastava (1989)
 Baltica to Laurentia: (47.5, 1.5, +49), after Evans and Pisarevsky (2008)
 Amazonia to absolute: (-3.92, 7.4, -142.43), interpolating between Colider volcanics (Bispo-Santos et al. 2008) and Salto de Ceu (D'Agrella-Filho et al. 2016) poles
 W-Africa to Amazonia: (30.27, -25.42, -88.65)
 C-SF to Laurentia: (-22.05, -139.8, -204.75), Bahia dykes (Option A, Salminen et al., 2016)
 SF to Congo:
 Kalahari to C-SF: (0.94, 34.83, -115.83) this work
 India 1: India to Baltica: (30.7, 54.3, +175.1), Dharwar – Sarmatia juxtaposition (Pisarevsky et al. 2013)
 India 2: India to Baltica: (14.87, 148.09, +243.28), close to Rodinia position
 North China to Laurentia: (52.95, 114.18, -167.08), interpolating between group of ca. 1.79 Ga and Yangzhuang formation (Wu et al. 2005) poles
 North Australia to Laurentia:
 (39.76, 100.0, +102.59) after Payne et al. (2009); Evans and Mitchell (2011)
 Mawson block to North Australia:
 South Australia to North Australia (-20.0, 135.0, +40.0), after Li and Evans, 2011

West Australia to North Australia (-20.0, 135.0, +40.0), after Li and Evans, 2011

Mawson block, E Antarctica to present Australia (-2, 38.9, +31.5), after Powell and Li (1994)

Siberia to Laurentia:

Anabar block to Laurentia (77.0, 98.0, +137.0), after Evans et al. 2016

Aldan to Anabar before Devonian (60, 115, +25), after Evans (2009)

1.58 Ga Åland dykes

1.58 Ga Satakunta N-S and NE-SW dykes

1.64 Ga Häme dykes

1647 Ma

1500 Ma