Baltica during the Ediacaran and Cambrian: A paleomagnetic study of Hailuoto sediments in Finland

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Abstract

We present a new Late Neoproterozoic paleomagnetic pole for Baltica from an inclined 272 m deep oriented sedimentary drill core in Hailuoto, Western Finland. Three components of magnetization were isolated with alternating field (AF) and thermal demagnetization treatments. The ChRM (characteristic remanence magnetization) component is a high coercivity/unblocking temperature dual polarity component of D = 334.4°; I = 57.7° with α_{95} = 5.8° and k = 25.2 (N = 26 samples), that passes a positive reversal test. This corresponds to a paleomagnetic pole of Plat = 60.5°N, Plon = 247.9°E with A95 = 7.6°. As it is a dual-polarity ChRM with high coercivity/blocking temperature, with no resemblance to younger events, we interpret it as a primary component. A paleolatitude for Hailuoto of 38.3° was calculated from the ChRM. Two secondary components were identified. The first is a low coercivity/blocking temperature component with a remanent magnetization of D = 239.0°; I = 67.3°; α_{95} = 8.7° (N = 13 samples), which we interpret as drilling-induced remanent magnetization (DIRM). The second secondary component has a remanent magnetization of D = 49.4°; I = 34.9°; α_{95} = 8.6° (N = 5 samples) and is commonly seen in Fennoscandian formations.

The ChRM Hailuoto pole adds to the scattered Ediacaran paleomagnetic data of Baltica and indicate large distances between similar aged paleomagnetic poles. We present reconstructions of Baltica and Laurentia between 616 and 550 Ma which move Baltica from high latitudes (615 Ma), over the polar region, to low latitudes (550 Ma), and Laurentia from low latitudes (615 Ma) to a polar position (570 Ma) and back to an equatorial position (550 Ma). Low to mid latitude position of Baltica determined by the Hailuoto paleomagnetic pole and the lack of glaciogenic sediments determined in an earlier study of Hailuoto sediments indicate a warm deposition environment.

Keywords: Ediacaran, Paleomagnetism, Paleogeography, Baltica, Laurentia, Hailuoto sediments

1. Introduction

The Late Neoproterozoic to Early Cambrian is a fascinating time interval in Earth's history. It includes global scale glaciations (Kirschvink, 1992; Hoffman *et al.*, 1998; Hoffman and Schrag, 2002), the diversification of early life (Knoll, 1992), and the break-up of the supercontinent Rodinia (e.g. Hoffman, 1991; Dalziel, 1997; Bingen *et al.* 1998; Evans, 2009). Laurentia and Baltica occupied central positions at the core of Rodinia. They became more isolated from other continents as rifting along Rodinia's margins occurred during the mid to late Neoproterozoic. By the start of the Ediacaran (*ca.* 635 Ma) the break-up of Rodinia was near completion. With the exception of Baltica and Siberia, all continents were separated from Laurentia (Li *et al*., 2013). Li *et al*. (2013) further show that by *ca.* 580 Ma the break-up of Rodinia was complete, Gondwana was in its early stages of assembly, and Baltica and Siberia had separated from Laurentia.

The role of paleomagnetism in reconstructing lithospheric blocks in their ancient paleopositions is vital. Paleomagnetism is the only quantitative tool for providing ancient latitudes and azimuthal orientations of continents and it also reveals information of the geomagnetic field in the past. A reliable paleomagnetic pole generally fulfills at least three of the seven quality criteria of Van der Voo (1990). If two of these include adequate geochronology and a positive paleomagnetic field test, the obtained paleomagnetic pole can be called a "key" pole (Buchan et al., 2000; Buchan, 2013). The paleogeography for the Ediacaran–Cambrian is the subject of significant controversy and it has puzzled researchers for the past two decades (Meert *et al.,* 1993, 1994; Kirschvink *et al.*, 1997; Evans, 1998; Meert, 1999; Popov *et al.*, 2002; Meert *et al.*, 2003; Nawrocki *et al.*, 2004; Iglesia Llanos *et al.*, 2005; McCausland *et al.*, 2007; Meert *et al.*, 2007; Pisarevsky *et al.*, 2008; Abrajevitch and Van der Voo, 2010; McCausland *et al.*, 2011). Ediacaran paleomagnetic data is complex since contradictory paleomagnetic results from coeval rocks have been obtained from both Baltica and Laurentia. McCausland *et al.* (2011) show how Laurentia has unclear paleogeographic relations during the Precambrian–Cambrian transition. Published paleomagnetic results from the Ediacaran period positions Laurentia at low paleolatitudes at 615 Ma, shortly after that at high southern latitudes during 590–570 Ma, and then again at low latitudes from 565 to 550 Ma.

A similar phenomenon is observed in high quality paleomagnetic data from Baltica. Higher quality paleomagnetic data for Baltica exist for only three time intervals during the Ediacaran–Ordovician (Meert, 2014). The 616 ± 3 Ma pole from Egersund dykes (Walderhaug *et al.*, 2007) positions Baltica at high latitudes. Between 570 Ma and 550 Ma, poles from Zigan formation (547.6 ± 3.8 Ma; Levashova et al., 2013), Verkhotina sediments (550.2 ± 4.6 Ma; 550 ± 5.3 Ma; Popov *et al.*, 2005), Winter coast sediments (555 ± 3 Ma; Popov *et al.*, 2002), Zolotitca sediments (550.2 ± 4.6 Ma, 550 ± 5.3 Ma; Iglesia Llanos *et al.*, 2005), Chernokamenskay group sediments (557 ± 13 Ma, Fedorova et al., 2014), Basu-Kukkarauk formation (ca. 560 Ma, Golovanova et al., 2011), and recent poles from 560-570 Ma Kurgashlya, Bakeevo and Krivava Luka ormations (Lubnina *et al.*, 2014) positions Baltica at low to equatorial latitudes. The Late Cambrian – Middle Ordovician poles from Narva Limestones (Khramov and Iosifidi, 2009) and St. Petersburg Limestones (Smethurst *et al.*, 1998) positions Baltica at high latitudes again. Based on these high quality data Baltica swayed by ca. 90°, from high latitudes at 616 Ma to low latitudes at 570 - 550 Ma and again to high latitudes from 550 to 500 Ma. The other Late Neoproterozoic poles presented in Table 2 are too questionable or unreliable to include in the APWP. The two poles - equatorial and high latitude - from 584 ± 7 Ma (^{39}Ar-^{40}Ar) Alnö carbonatite complex dikes were artificially derived from a wide spread in magnetic declinations and inclinations (Meert et al., 2007). The 583 ± 15 Ma (^{39}Ar-^{40}Ar) pole from Fen Carbonatite Complex, lack stability tests and is likely a Permo-Triassic overprint (Meert, 2014). The Cambrian Andarum-alum limestone and Tornetrask group poles are questionable. The Tornetrask results (Torsvik and Rehnström, 2001) come from samples taken close to tectonically disturbed regions of Caledonian front, which can cause complications in interpreting those results (Meert, 2014). Pole for Andarum-alum is calculated only from 11 samples yielding thus only virtual geomagnetic pole, which cannot be used in APWP. Both the Tornetrask and Andarum-alum (Torsvik and Rehnström, 2001) have a dual-polarity magnetization, but they fall close to the remagnetized Fen Complex pole.

Explanations for the two almost coeval magnetization components (high and low latitude) include fast continental motions (Meert and Tamrat, 2004), rapid true polar wander (TPW) (Mitchell *et al.*, 2011), or a non-actualistic geodynamo where the geomagnetic field alternated between axial and equatorial configurations during the Ediacaran (Abrajevitch and Van der Voo, 2010). Abrajevitch and Van der Voo (2010) argued that high plate velocities and true polar wander are implausible explanations for such rapid changes in the positions of continents, as both TPW and plate tectonics are speed-limited phenomena. However, as Meert (2014) pointed out, the analysis of Abrajevitch and Van der Voo (2010) analysis relied on problematic poles for Baltica, implying apparent polar wander (APW) rates exceeding 70 cm/yr (Meert, 2007). By taking into account reliable poles only, rapid plate motions can explain individual segments of the apparent polar wander path (APWP).

The Neoproterozoic is also characterized by global scale glaciations. Two major glaciation events occurred during the Ediacaran, namely the Marinoan (*ca.* 635 Ma) and the Gaskiers (*ca.* 580 Ma) (e.g. Hoffman and Li, 2009). The Marinoan glaciation extended from >70° palaeolatitude to the equator (e.g. Hoffman and Li, 2009), implying global glaciation and has been interpreted to represent a Snowball Earth event (Kirschvink, 1992; Hoffman *et al.*, 1998; Hoffman and Schrag, 2002). Marinoan glacial deposits have been found on all major continental blocks except for the North China block (Li *et al.*, 2013). Gaskiers glacial deposits are found on Baltica, Laurentia, North China, Australia, Tarim, Avalonia, Congo-São Francisco and Amazonia (Li *et al.*, 2013), however, feeble evidence of low latitude glaciation (e.g. Hoffman and Li, 2009) suggests that the Gaskiers was not as widespread as the Marinoan. Li *et al.* (2013) point out that on the Avalon Peninsula, Newfoundland, precise TIMS U—Pb zircon dates appear to constrain the duration of the Gaskiers glaciation to ≤2.6 m.y. (Condon and Bowring, 2011). The short duration has commonly been regarded as evidence that it was not a Snowball glaciation (e.g., Halverson, 2006).

The aim of the present work is: 1) to obtain a new paleomagnetic pole for Baltica; 2) to test the Late Neoprpterozoic paleogeoraphic positions of Baltica; and 3) to better understand the environmental conditions during that time. With a new Late Neoproterozoic pole, we further aim to explore if the large sways in the Ediacaran APWP can be explained by high plate velocities alone.

2. Geological background

The assembly of the Baltica began at 2.0 Ga when Sarmatia and Volgo-Uralia joined each other to form the Volgo-Sarmatian protocraton, which existed as a separate unit until ca. 1.8-1.7 Ga when it docked with Fennoscandia and a unified Baltica was created (Bogdanova et al., 2008). After the movements of the Svecofennian orogeny terminated at *ca.* 1750 Ma (Bogdanova et al., 2008) and Baltica and Laurentia formed a joint NENA continent (Gower et al., 1990). This was followed by a geologically passive time of about 150 Ma when no significant deformation occurred and when the crust eroded to a peneplain. The quiet time ended with the intrusion of rapakivi granites and associated bimodal magmas at 1650 - 1500 Ma (Rämö and Haapala, 2005; Bogdanova et al., 2008). This caused instability of the crust and led to the development of intracratonic rift basins between ca. 1600 - 1300 Ma when thick fluvial layers started to fill the basins (Kohonen and Rämö, 2005). The crust was thinned by about 20 km at the Gulf of Finland and about 10 km at Lake Ladoga in Russia and at the Bay of Bothnia (Figure 1; Korja *et al.*, 1993).

Due to the Sveconorwegian orogeny, the southwestern parts of the Fennoscandian shield were uplifted. This resulted in the formation of thick and hundreds of kilometers long foreland sedimentary deposits east of the Svoconorwegian orogeny (Larson *et al.*, 1999). During Neoproterozoic a vast area of the crystalline bedrock was exposed when about 500–2000 meters of sediments were eroded away (Puura *et al.*, 1996). Due to marine transgression, a new sedimentary event began in Baltica at the end of the Neoproterozoic (Puura *et al.*, 1996). Fluvial and shallow marine deposition prevailed in the slowly submerging northwestern part of Baltica (Kohonen and Rämö, 2005). These sedimentary layers covered the entire southern Finland (Puura *et al.*, 1996; Bogdanova *et al.*, 2008). Today, the scattered Meso- and Neoproterozoic deposits of sedimentary rocks in Finland are preserved from erosion in tectonic basins in Satakunta, Muhos, the Gulf of Bothnia (the northern arm of the Baltic Sea) as well as in meteorite impact craters (Kohonen and Rämö, 2005).

The Muhos sedimentary basin forms a SE-NW trending area of about 20 km by 50 km, mainly consisting of Mesoproterozoic clay and silt rocks (Kohonen and Rämö, 2005). This graben-like formation has a maximum depth of about 1 km (Kalla, 1960) and is surrounded by the late orogenic (*ca.* 1800 Ma) granitoid complex. The Muhos formation has been considered to be of the same age as the Mesoproterozoic Satakunta sandstone (Bogdanova et al., 2008). The Muhos sedimentary formation continues to the West and covers almost the entire Bothnian Bay between Finland and Sweden (Wannäs, 1989). The Neoproterozoic Hailuoto formation was deposited on top of the Muhos formation and occurs West of the Muhos on the island of Hailuoto. Hailuoto Island is situated off the coast of the city of Oulu, Northern Finland (65°N 25°E) (Fig. 1). The island arose from the Bothnian Bay during the Quaternary post-glacial land uplift (Tynni and Donner, 1980). The sedimentary bedrock of Hailuoto is covered by a Quaternary layer of loose sand of up to *ca.* 70 m thick. The bedrock consists of interbedded conglomerate, sandstone and mudstone up to a maximum depth of *ca.* 560 m. The dominant rock type of the Hailuoto formation is a medium-grained, pale pink or light greenish subarkose (Veltheim, 1969; Tynni and Donner, 1980).

INSERT FIGURE 1 HERE

Tynni and Donner (1980) point out that the sedimentary rocks of Hailuoto resemble glacial clay, but are in fact not glaciogenic. The depositional age of the Hailuoto sediments is poorly constrained. The micro-fossils, such as filamentous blue-green algae (eg. *Oscillatoriopsis magna*), in the upper part of the formation (Tynni and Donner, 1980) indicates a late Neoproterozoic age. The micro-fossils diversity also indicates a deposition environment of low oxygen content and warm temperatures. Tynni and Donner (1980) correlated the upper part of the Hailuoto sequence with the upper part of the Visingö formation in Sweden. Subsequent dating of shales and clay from the the upper part of the Visingö formation gave a Rb/Sr age of 663 - 703 Ma (Bonhomme and Welin, 1983). Yet the fossil species suggest younger ages for Hailuoto, since among the Acritarch fossils found in Hailuoto sediments are species *Kildinella sinica* and *Churia cicularis* (Tynni and Donner, 1980). Both are closely related to the genus *Leiospaeridia* (Lindgren, 1981, Tynni and Donner, 1980) found in the Kotlin formation of the Vendian sediments in Estonia (Mens and Pirrus, 1997). The age of the lower Kotlin boundary in the White Sea area has been confined between 551 Ma and 548 Ma based on U-Pb zircon dates of volcanic tuffs (Grazhdankin *et al.*, 2011). So far the Hailuoto depositional age remains poorly constrained as Late Neoproterozoic, if the correlation with Visingö is correct then age of deposits is older than 663 - 703 Ma, but based on obtained old fossil record the sediments might be as young as the Precambrian-Cambrian boundary. Although the exact age of deposition of the

Hailuoto sediments is not known, in this paper we will assume an age range of 570-600 Ma suggested by Tynni and Donner (1980).

3. Methods

3.1 Sampling and sample preparation

The 272 m deep Hailuoto drill core M52-HAIL-04-005 mainly consists of conglomeratic sandstone and sandy conglomerates with smaller layers of mudstone, and sandstone (Solismaa, 2008; Fig. 4). The conglomeratic sediments are poorly sorted, and the granule and small pebble sized clasts are sparsely dispersed in a finer grained matrix. The drilling was done with an angle of 70° to the direction of 226°. In the core layers inclined to the borehole axis can be seen. Orientation was done by using the Ballmark orientation system, in which an indent marking is made on the drill core at the time the core is broken. This is one of the first times when paleomagnetic data is obtained of the rocks from inclined deep drill core, which use simple geometry for sample orientation instead of cumbersome and imprecise in-door studies for the same purpose.

Oriented samples of sedimentary rock were prepared from the drill core. Samples with a maximum length of about 6 cm were first cut from the drill core and from these 7 - 8cm^3 cubic specimens parallel to the core, were prepared for paleomagnetic measurements. The sample numbering corresponds with the depth of the sample in the drill core (in decimeter), and in many cases more than one specimen were prepared from a sample. Samples from the entire sedimentary part of the drill core were measured (Fig. 4). There are three large gaps of more than 10 m each between measured samples at depth ranges 69.5 - 79.7 m, 87.2 - 101.2 m (largely lost core), and 170.2 – 181.8 m (conglomeratic with clasts up to 5cm). When we exclude these gaps the average distance between measured samples is 1.6 metres, with distances as small as 10 cm (Fig. 4).

3.2 Laboratory measurements

The measurements were carried out at the Geophysics laboratory of the Geological Survey of Finland (GTK) and the Solid Earth Geophysics Laboratory of the University of Helsinki (UH). Paleomagnetic measurements with progressive alternating field (AF) and thermal demagnetizations were performed using a 2G-DC SQUID magnetometer (UH), 2G-RF SQUID magnetometer (GTK) (both equipped with Model 2G600 automatic sample degaussing systems) and an ASC Scientific TD-48SC (UH) thermal demagnetizer using an argon atmosphere to minimize oxidation during thermal demagnetization. Prior to demagnetizing part of the samples were immersed in liquid nitrogen in a non-magnetic field (Borradaile, 2004) and after that they were heated up to 125 °C to demagnetize remanent magnetization carried by goetithe. Magnetic susceptibility was measured after each heating step to monitor possible chemical alterations.

Natural remanent magnetization (NRM) components were visually identified using stereographic and orthogonal projections (Zijderveld, 1967) and the directions were calculated by a least squares method (Leino, 1991), with a mean angular deviation (MAD) equal or less than 6° (with a few coarser grained samples a higher MAD was accepted). Mean remanence directions for the different components were calculated according to Fisher (1953). APWP and paleogeographic reconstructions were plotted with the GMAP programme (Torsvik and Smethurst, 1999). The magnetic carriers were identified by thermomagnetic analysis of selected specimens along the depth profile, using Agico's KLY-3S Kappabridge system (UH), which measures the bulk susceptibility (k) of the samples while heating the samples from room temperature to 700 °C and cooling back to room temperature (in Argon gas).

4. Results

4.1. Paleomagnetic and rock magnetic results of Hailuoto

A total of 148 specimens from 90 samples were measured, of which 48 specimens from 36 samples gave stable results. Results from rest of 90 samples were unstable and will not be further discussed. From accepted 36 samples we identified three prevailing paleomagnetic components, which are summarized in Table 1 and examples of thermal demagnetization behaviour are shown in Fig. 2.The mean paleomagnetic directions are plotted in Fig. 3. As shown in Table 1 and Fig. 4, the results were obtained from various lithologies along the drill core, however the coarser-grained parts of the drill core yielded largely unstable magnetizations, and were therefore unusable. It must be noted that the stratigraphic log in Fig. 4 is simplified from a detailed stratigraphic log by Solismaa (2008), and that the indicated lithologies are interbedded with many layers that are too small to be detected on the scale of the figure.

INSERT TABLE 1 HERE

INSERT FIGURE 2 HERE

We employed the standard convention of assigning “normal” polarity to north-directed ChRM vectors from Proterozoic rocks in Baltica; relating this arbitrary definition to the absolute sense of Precambrian geomagnetic polarity is uncertain due to the lack of continuity in the Baltica APWP prior to Phanerozoic. The dual polarity ChRM componentwas obtained from 26 samples. Normal polarity component shows an intermediate downward NNW direction (7 samples) and reversed polarity component shows an intermediate upward SSE direction (19 samples). Based on thermal demagnetization curves, the magnetic carriers of the ChRM are magnetite and hematite. Presence of both magnetite and hematite is supported by the thermomagnetic curves (susceptibility vs. temperature) of samples at 79.9 m and 170.2 m depths respectively (Fig. 5). As can be seen in Figure 2a the reversed polarity ChRM is carried by both magnetite and hematite. This indicates that hematite is produced soon after deposition and the chemical remanent magnetization in hematite has recorded the magnetic field direction essentially contemporaneous with magnetite and is regarded as a primary magnetization. The case is slightly different for normal polarity samples, where two almost similar directed components were obtained (Fig. 2b). The normal (D = 323.5°; I= 40.7°; α_{95}=14.5°; k=18.8, n = 7) and reversed (D = 158.5°; I = -45.5°; α_{95} = 8.5°; k = 16.5, n = 19) observed ChRM directions pass the McFadden and McElhinny (1990) reversal test with classification C (observed angle, γ = 11.93° and, critical angle, γ_c = 18.17°). To confirm primary remanence of ChRM, a conglomerate test was attempted on 0.5 – 2 cm clasts, prepared from conglomeratic samples. Although the remanence of the clasts show a random distribution, they proved to have unstable magnetizations, and the conglomerate test is therefore unreliable.

The other two magnetization components obtained from Hailuoto samples are overprinted magnetization components. The first one with low coercivities/blocking temperatures (Fig.2) is carried by 13 of the measured samples. It is directed to the SW and has a steep positive inclination similar to the drilling direction. We interpret it as a drilling-induced remanent magnetization (DIRM), which is a common phenomenon in multi-domain magnetic grains in sedimentary rocks (Audunsson and Levi, 1989; De Wall and Worm, 2001). It has no geological meaning and will not be further discussed. The present earth field (PEF) direction for Hailuoto is D_{PEF} = 8.2°; I_{PEF} = 75.9° (Fig. 4.) and was not detected in any of the measured samples, as it was possibly overprinted by the DIRM. The second secondary magnetization component was obtained from five samples and is directed to the NE with a moderately steep positive inclination. It is isolated from low to intermediate coercivities and temperatures. This component often occurs in Fennoscandian formations, and it was recently obtained from Subjotnian Satakunta and Åland diabase dykes (Salminen *et al.*, 2014, Salminen *et al.*, 2015). It is considered to represent a late Paleozoic secondary magnetization event (Preeden *et al.* 2009).

INSERT FIGURE 3 HERE

INSERT FIGURE 4 HERE

INSERT FIGURE 5 HERE

4.2 Inclination shallowing

It is important to make inclination corrections for paleomagnetic results obtained from sedimentary rocks since inclination shallowing can be as much as 20° at mid-latitudes (Kodama, 2012). Examples of sedimentary inclination shallowing are numerous in a variety of natural settings (e.g. Bilardello and Kodama, 2010) and by several laboratory re-deposition experiments (e.g. Tauxe and Kent, 1984). As Domeier *et al.* (2012) point out due to the great diversity of sediment characteristics and depositional conditions, in addition to the assortment of specific mechanisms by which a magnetization may acquire a shallow inclination bias, the magnitude of the bias in sedimentary rocks is variable. So far there are four ways to correct this: 1) the DRM tensor approach of Jackson *et al.* (1991); 2) the Elongation/Inclination (E/I) method of Tauxe and Kent (1984) using the statistical field model TK03.GAD; 3) the E/I method of Tauxe and Kent (2004) using samples from a single horizon and assuming a Fisher (1953) distribution; and 4) a blanket correction based on assumed values for flattening factor f using an inclination correction equation by King (1955):

$$\tan(I_o) = f \tan(I_f)$$

where I_o is the observed inclination, I_f is the field inclination and f is an empirically derived 'flattening factor' (Tauxe *et al.*, 2008). According to the compilation of Bilardello and Kodama (2010), f values from magnetite dominated sedimentary rocks range from 0.54 to 0.79, with a mean of 0.65, whereas hematite-dominated sedimentary rocks have yielded f values from 0.4 to 0.83, with a mean of 0.59. We corrected the ChRM using the flattening estimate since the DRM tensor is difficult to quantify; the first E/I method requires a large data set (>100), which we do not have; the second E/I method has never been tried in practice. We followed the suggestion made by Domeier *et al.* (2012) and

Torsvik *et al.* (2012) and used a flattening estimate of f = 0.6. The results are shown in Table 1 and Fig. 3 together with the observed results. The normal (D = 323.1°; I= 54.2°; α_{95}=12.0°; k=26.3, n = 7) and reversed (D = 158.5°; I = -58.7°; α_{95} = 6.7°; k = 26.1, n = 19) inclination corrected ChRM directions also pass the reversal test with classification C (observed angle, γ = 9.84° and, critical angle, γ_c = 16.46°). Moreover the statistics of bot normal and reversed directions get better after the inclination correction.

5. Discussion

5.1 The Hailuoto palemagnetic pole

The combined observed ChRM component of the Hailuoto sediments (D = 334.2°; I = 44.4°; α_{95} = 7.2°; k = 16.5) yields a paleomagnetic pole of Plat = 48.7°N and Plon = 241.1°E with A95 = 8.1°. The inclination corrected direction of D = 334.4°; I = 57.7°; α_{95} = 5.8°; k = 25.2 yields a paleomagnetic pole of Plat = 60.5N° and Plon = 247.9E° with A95 = 7.6°. As it is a dual-polarity ChRM with high coercivity/blocking temperature, with no resemblance to younger events, we interpret it as a primary component. Furthermore, the confidence limit (α_{95}) of the ChRM improves after the inclination correction, suggesting that it is the primary DRM. The paleolatitude of Hailuoto implied by the inclination corrected ChRM is 38.3°.

The Hailuoto paleomagnetic pole passes five of the seven Van der Voo reliability criteria (Van der Voo, 1990) for paleomagnetic data: a sufficient number of samples, adequate demagnetization (including vector subtraction), structural control and tectonic coherence with Baltica, the presence of reversals, and no resemblance to younger (i.e. Phanerozoic) paleopoles (Torsvik *et al.* 2012). It lacks the critical first and fourth criteria, i.e. a well determined age and a field stability tests, and can therefore not be regarded a key paleomagnetic pole (Buchan, 2013). Moreover the Hailuoto depositional age remains poorly constrained as Late Neoproterozoic (above in section 2). Although the exact age of deposition of the Hailuoto sediments is not known, in this paper we will assume an age range of 570-600 Ma suggested by Tynni and Donner (1980), which is also supported by the fact that the inclination corrected Hailuoto pole plots close to the well defined 560-570 Ma pole from Krivaya Luka formation (Lubnina et al., 2014).We wish to include the Hailuoto paleomagnetic pole in the discussion of the Ediacaran APWP and reconstructions due to its crucial age range.

5.2 Ediacaran apparent polar wander path for Baltica

We plotted the ChRM pole of Hailuoto with selected Neoproterozoic – Cambrian poles, listed in Table 2 (Fig. 6.). Although the exact deposition age of the Hailuoto sediments is unsure, for the purpose of this discussion we assume the age range proposed by Tynni and Donner (1980) of 600 - 570 Ma, which places the age of the Hailuoto pole between the Egersund (616 Ma) and the Verkhotina (Popov *et al.* 2005) and Zolotitca (Iglesia Llanos *et al.* 2005) poles (ca. 550 Ma). The polarity was chosen to minimize the distances between the older Egersund pole (616 Ma) and younger poles (570-560 Ma) from Southern Urals. In Table 2 Ediacaran poles are classified into highly reliable A grade poles (well determined age and positive field stability test), seemingly reliable B grade poles (well determined age or positive field test), and questionable to unreliable C and D grade poles (neither well determined age nor positive field test). Considering the Hailuoto pole together with A and B grade poles, we propose a loop in the APWP that starts at the Egersund pole (616 Ma), jumps to the Hailuoto pole (ca. 600 - 570 Ma), and follows the poles of Krivaya, Kurgahlya and Bakeevo formations (570 – 560 Ma) (Lubnina *et al.* 2014), continues to the 550 Ma Verkhotina (Popov *et al.* 2005) and Zolotitca (Iglesia Llanos *et al.* 2005) poles, and ends at the St. Petersburg limestone pole (Smethurst *et al.* 1998) close to the Egersund pole (illustrated with black dotted line in Fig. 6A). The Ediacaran poles of Laurentia (Fig.6B) starts with the Long Range dykes pole (615 Ma) (Murthy *et al.* 1992) followed by the poles of Baie de Moutons A (583 Ma) (McCausland *et al.* 2011), Callandar alkaline complex (Symons and Chiasson, 1991) and Catoctin basalts (575 Ma) (Meert *et al.* 1994), and Skinner cove (550 Ma) (McCausland and Hodych, 1998), and completing the loop with the Florida Mountains auriole pole (503 Ma).

The angle of the Baltica APWP segment between the Egersund dyke pole (616 Ma) and the Hailuoto pole is 98° for the uncorrected pole and 86° for the inclination corrected pole (f=0.6), which infers a plate motion of between *ca.* 24 and 21 cm/yr respectively with the lower Hailuoto age limit of 570 Ma. This is consistent with the upper speed limit for plate tectonics (between 20 and 25 cm/yr) reported by Meert *et al.*, (1993) and Gurnis & Torsvik (1994). The APWP of Baltica between the 550 Ma mean pole and the St Peterburg (478 Ma) (Smethurst *et al.* 1998) pole (Fig. 6A) yields an 87° angle which results in a velocity of *ca.* 16 cm/yr when considering only A and B grade poles.

INSERT TABLE 2 HERE

INSERT FIGURE 6 HERE

5.3 True polar wander and a non-actualistic geogynamo during the Ediacaran

The APWP of Baltica presented in Fig. 6 is plausible as a result of high plate velocity alone, however this does not negate the TPW (Mitchell *et al.*, 2011) and non-actualistic geodynamo (Abrajevitch and Van der Voo, 2010) hypotheses for the large sway in the APWP, especially if we consider the higher age limit of 600 Ma for the Hailuoto formation.

True polar wander (TPW) (Evans, 2003; Mitchell *et al.*, 2011) and a non-actualistic geodynamo (Abrajevich and Van der Voo, 2010) are two hypotheses given as possible mechanisms for the rapid continental motion during the Ediacaran. True polar wander is the rotation of the Earth with respect to its spin axis causing the geographic locations of the North and South poles to change or "wander". Such a shift in poles results in a systematic change in the APWP's of all cratons. Although the APWP's for Baltica and Laurentia presented in Fig. 6 both form a loop starting at ca. 615 Ma and ending at ca. 500 Ma, the angles between individual coeval poles differ, and therefore negate TPW as a main cause for large APWP segments.

In the case of Laurentia, the large angle of 66° of the APWP segment between Catoctin Basalts (572 Ma) and Sept-Iles intrusions (564 Ma), which equates to 1.05 m/yr, cannot be explained by high plate velocity alone. However, the non-actualistic geodynamo hypothesis by Abrajevitch and Van der Voo (2010), which postulates an alternation of the geomagnetic dipole axis between a co-axial and an equatorial alignment, partly relies on problematic poles for both Laurentia and Baltica. The Grenville dykes yield both a shallow and steep pole, however only the steep pole can be considered a primary pole as it is supported by a positive field test (Murthy 1971, Halls, *et. al.* 2015). In the case of Baltica the model is based on the Alnö carbonate complex (584 ± 7 Ma) (Meert *et al.* 2007, Piper 1981) in which both a high and a low latitude component co-exist in the same unit. However the primary nature has not been proven for either. Furthermore, the Fen carbonate complex pole (Meert *et al.* 1998), another pole in the model, lacks stability tests and is likely a Permo-Triassic overprint (Meert 2014). When we omit these problematic poles the model resembles the APWP presented here, i.e. a back and forth sway from shallow to steep back to shallow poles.

5.4 Reconstruction of Baltica and Laurentia between 615 Ma and 550 Ma

The reconstructions illustrated in Fig. 7 show a possible scenario for Baltica and Laurentia between 615 Ma and 550 Ma. In the Baltica reconstructions we consider an age of 600-570 Ma for Hailuoto, but as mentioned above, the Hailuoto sedimentary formation may extend this age rage. The poles used for the reconstructions are listed in Table 3. In these reconstructions Baltica moves from a high latitude position at 615 Ma to low and mid latitudes between 600 Ma and 550 Ma. The orientation of Baltica is very similar between 600 Ma and 550 Ma with the west coast of Norway directed towards the pole. In the 615 Ma reconstruction however, the west coast of Norway points away from the pole, implying that Baltica moved across the polar region between 615 Ma and 600 Ma without significant rotation. This motion of Baltica coincides with the final break-up of Baltica and Laurentia from Rodinia configuration and the opening of the Iapetus Ocean, which is supported by the fact the Long Range (Laurentia) and Egersund (Baltica) dykes are rift-related (Bingen *et al.* 1998, Puffer 2002). Laurentia moved from low latitudes at 615 Ma to a polar position at ca. 570 Ma and back to an equatorial position at 550 Ma (Fig 7). The reconstructions presented here are consistent with recent reconstructions by Meert (2014), Li *et al.* (2013) and McCausland (2011).

INSERT FIGURE 7 HERE

INSERT TABLE 3 HERE

The polar position of Laurentia at 570 Ma (Fig 7) and the high latitude model of early Gondwana land at 575 Ma (Pisarevsky *et al.* 2008), as well as the low latitude of Baltica presented in this study and warm deposition environment of Hailuoto sediments (Tynni and Donner, 1980), suggest that no global scale glaciation event, as presented in the snowball (Kirschvink, 1992) model, took place during the time of the Hailuoto deposition. These conditions therefore indicate zonal surface temperature gradients similar to today, i.e. warm equatorial areas and cold polar areas, unlike the high obliquity model presented by Williams (1975, 1993, 2008).

Conclusion

We present a new Neoproterozoic pole for Baltica obtained from a drill core from Hailuoto sedimentary rock with an estimated age of 570 - 600 Ma. The paleomagnetic pole Plat = 48.7°N, Plon = 241.1°E, A95 = 8.1° (inclination corrected: Plat = 60.5°N, Plon = 247.9°E, A95 = 7.6°) suggests rapid drift of Baltica between the intrusion of Egersund diabase dykes (616 Ma), and the time of deposition of the Hailuoto sediments (600 - 570 Ma). High plate velocity is a plausible cause for the rapid drift of Baltica during the Ediacaran, however considering the higher age limit of Hailuoto sediments (600 Ma), a non-actualistic geomagnetic field cannot be completely dismissed as a contributing factor in the large distances between Ediacaran poles. Our reconstructions of Baltica and Laurentia between 615 Ma and 550 Ma, move Baltica from high latitudes (615 Ma), over the polar region, to an equatorial position (550 Ma) and move Laurentia from low latitudes (615 Ma) to a polar position (570 Ma) and back down to an equatorial position. A warm deposition environment and the lack of glaciogenic sediments determined in an earlier study of Hailuoto sediments, as well as the low to mid latitute position of Baltica determined by the Hailuoto paleomagnetic pole, indicate warm equatorial areas and cold polar areas, unlike in high obliquity models proposed for the Ediacaran.

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FIGURE CAPTION:

Fig.1. Location of Hailuoto and the Muhos formation. From Veltheim (1969).

Table 1. Mean paleomagnetic results from Hailuoto sediments.

Fig. 2. Examples of thermal (H5_1887A and H5_1471A) demagnetization of Hailuoto sediment specimens. On stereoplots (A), solid symbols represent lower hemisphere vectors, and open symbols represent upper hemisphere vectors. B is the intensity plots (J/J_0) of thermal demagnetization. On orthogonal projection diagrams (C), solid symbols show projections onto the horizontal plane, and open symbols show projections onto the vertical plane. H5_1887A = normal polarity ChRM, H5_1471A = reversed polarity ChRM, H5_1334A = steep secondary component (DiRM), and H5_1012A = secondary component B.

Fig. 3. Left: Stereoplot showing the paleomagnetic results of the ChRM (observed) component. Right: Mean directions, including the secondary components; normal and reversed directions of the ChRM are combined in a single mean direction. Solid symbols represent lower hemisphere vectors, and open symbols represent upper hemisphere vectors. PEF indicates the present day magnetic field direction for Hailuoto.

Fig. 4. Stratigraphy of the drill core (M52-HAIL-04-005), and magnetic inclination of ChRM along the depth profile. Geological log by Solismaa (2008).Colors in the inclination column: black (white) indicates normal (reversed) polarity. Grey indicates the lack of stable paleomagnetic data. Black dots indicate all the measured samples.

Fig. 5. Thermomagnetic curves (susceptibility vs. temperature) for Hailuoto sediment samples, indicating magnetite and hematite as the magnetic carriers. Solid line indicates heating curve and the dotted line indicates the cooling curve.

Table 2. Selected Ediacaran poles and Ordovician poles for Baltica

Fig. 6. APWP of (A) Baltica and (B) Laurentia. The Hailuoto ChRM pole is marked in green. A and B grade poles are indicated by red and orange respectively. Numbers in bold indicate age in Ma.

Fig. 7. Reconstructions of Baltica and Laurentia between 615 -550 Ma. Baltica shown observed (obs.) and inclination corrected (i.c.) positions in 550 Ma reconstruction. Euler poles provided in Table 3.

Table 3. Paleomagnetic poles used for the reconstruction in Fig 7.

Sample	Lith.	N/n	Dec (°)	Inc (°)	α_{95} (°)	k	Plat (°N)	Plon (°E)	A_{95} (°)	K
Hailuoto ChRM component. normal polarity										
H5_626	Ms	1/1	301.1	24.2			24.3	271.2		
H5_655	Ms	1/4*	293.9	48.1	16.1	33.4	36.2	286.5	18.6	25.4
H5_797	sCSs	1/1	341.2	31.7			40.5	228.7		
H5_1259	SC	1/1	318.0	33.8			35.8	256.3		
H5-1468	MSs	1/1	342.3	58.6			62.3	235.2		
H5_1471	MSs	1/1	336.6	41.1			45.8	236.3		
H5-1489	CSs	1/1	335.4	37.7			43.2	237.0		
Mean ChRM (N)		***7/10**	**323.5**	**40.7**	**14.5**	**18.8**	**43.0**	**252.0**	**15.4**	**16.4**
Mean (f = 0.6)			**323.1**	**54.2**	**12.0**	**26.3**	**53.9**	**259.2**	**14.8**	**17.6**
Hailuoto ChRM component. reversed polarity										
H5_612	Ms	1/1	196.2	-35.2			-43.1	3.7		
H5_1334	CSs	1/1	161.0	-44.0			-48.9	51.3		
H5_1420	MSs	1/1	171.6	-32.7			-42.4	35.6		
H5-1439	MSs	1/1	171.6	-25.4			-38.0	35.2		
H5_1531	sCSs	1/1	132.3	-33.8			-33.8	82.4		
H5_1551	SC	1/1	178.5	-60.3			-66.2	27.6		
H5_1581	SC	1/1	127.8	-50.2			-43.5	93.8		
H5-1598	SC	1/1	154.9	-40.3			-44.9	58.2		
H5-1650	SC	1/1	149.9	-39.0			-42.7	64.0		
H5_1656	SC	1/1	150.1	-67.9			-69.1	86.7		
H5_1678	SC	1/1	187.8	-33.1			-42.7	14.7		
H5_1693	SC	1/1	158.5	-64.7			-68.2	67.5		
H5_1702	SC	1/2	116.2	-39.1			-30.9	100.5		
H5_1842	SC	1/1	156.4	-35.9			-42.2	55.3		
H5_1852	SC	1/1	191.7	-40.3			-47.2	8.8		
H5_1863	SC	1/1	166.5	-36.4			-44.3	42.6		
H5_1887	SC	1/2	157.4	-43.2			-47.5	55.8		
H5_1930	SC	1/1	122.9	-53.9			-44.6	101.2		
H5-1979	SC	1/1	139.3	-47.4			-45.6	79.7		
Mean ChRM (R)		***19/21**	**158.5**	**-45.5**	**8.5**	**16.5**	**-49.5**	**55.1**	**9.7**	**12.8**
Mean (f = 0.6)			**158.5**	**-58.7**	**6.7**	**26.1**	**-61.6**	**60.4**	**9**	**14.9**
Mean ChRM (C)		***26/31**	**334.2**	**44.4**	**7.2**	**16.5**	**48.7**	**241.1**	**8.1**	**13.2**
Mean (f = 0.6)			**334.4**	**57.7**	**5.8**	**25.2**	**60.5**	**247.9**	**7.6**	**15.0**
Hailuoto secondary component. steep										
H5_605	Ms	1/1	240.4	42.7			11.0	331.3		
H5_609	Ms	1/2	219.3	48.8			9.6	350.9		
H5_626	Ms	1/1	257.3	75.1			49.2	340.3		
H5_645	Ms	1/1	198.8	78.2			43.1	15.0		
H5_664	Ms	1/1	209.7	59.5			17.9	1.4		

H5_797	sCSs	1/1	291.2	51.4			37.6	290.9		
H5_1183	SC	1/2	210.4	61.8			20.6	1.5		
H5_1259	SC	1/1	206.7	74.2			37.1	8.7		
H5_1307	sCSs	1/1	215.7	70.8			33.2	1.3		
H5_1334	CSs	1/1	285.8	67.9			50.9	310.6		
H5_1345	CSs	1/1	251.8	65.8			35.9	333.2		
H5_2328	CSs	1/3*	246.5	74.4	23.9	27.6	45.3	345.4	43.6	9.0
H5_2428	WB	1/2	255.8	66.8			38.4	331.2		
Mean secondary. steep		**13*/18**	**239.0**	**67.3**	**8.7**	**23.9**	**35.2**	**344.5**	**12.5**	**12.0**
Hailuoto secondary component B										
H5_645	Ms	1/3*	45.7	28.1	33.8	14.3	31.2	150.8	40.6	10.3
H5_655	Ms	1/1	53	47.2			40.9	136.5		
H5_797	sCSs	1/1	47.3	32.5			33.2	148.0		
H5_1012	SMs	1/1	45.7	28.1			31.2	150.8		
H5_1396	CSs	1/1	56.7	38.4			33.3	136.4		
Mean secondary. B		**5*/7**	**49.4**	**34.9**	**8.6**	**80.7**	**34.1**	**144.7**	**6.9**	**122.8**

Sample is the sample number (H5 refers to the drill core, and the number following it to the depth in decimetre); N/n are number of samples/specimens; Lith. is lithology (Ms – mudstone, SMs – sandy mudstone, MSs – muddy sandstone, sCSs – slightly conglomeratic sandstone, CSs – conglomeratic sandstone, SC – sandy conglomerate, WB – weathered bedrock); Dec and Inc are magnetic declination and inclination; α_{95} and k are the confidence limit (95%) and precision parameter respectively of the magnetic component; Plat and Plong are the paleo-latitude and longitude; A95 and K are the confidence limit (95%) and precision parameter of the paleopole.

Formation	Inc	Plat (N)	Plon (E)	dp	dm	A95	Q(1-6)	Class	Age (Ma)	Ref
Narva limestones, comp C1		18	55	5	7	5.9	011011 = 4	C	475	Khramov and Iosifidi (2009)
St Petersburg limestone, comp. Pr		34.7	59.1	5.7	6.4	6.0	011110 = 4	B	478	Smethurst *et al.* (1998)
Narva sediments, comp. C2	obs.	22	87	5	6	5.5	011011 = 4	C	500	Khramov and Iosifidi (2009)
	i.c.	33.6	77.2							
Andarum-alum shale	obs.	52	111	7	10	8.4	001010 = 2	D	500	Torsvik & Rehnström (2001)
	i.c.	60.4	89.4							
Tornetrask formation	obs.	56	116	12	15	13.4	001011= 3	C	535	Torsvik & Rehnström (2001)
	i.c.	66.1	96.9							
Nekso sandstone	obs.	40	170			5.1	011010 = 3	C	545	Lewandowski and Abrahamsen (2003)
	i.c.	43.7	169							
Zigan formation clastic rocks -C	obs.	16.2	318.4			4.1	111011 = 5	B	547.6 ± 3.8	Levashova *et al.* (2013)
Verkhotina sediments, comp. Z	obs.	31.7	292.9	1.6	2.7	2.1	111111 = 6	A	550.2 ± 4.6 550 ± 5.3	Popov *et al.* (2005)
	i.c.	42.5	293.0							
Zolotitca sediments, comp. B	obs.	28.3	290.0	2.5	4.4	3.3	111011 = 5	B	550.2 ± 4.6 550 ± 5.3	Iglesia Llanos *et al.* (2005)
	i.c.	41.4	299.0							
Volhyn lavas - combined		20.0	4.4			28.3	001011 = 3	C	561-580	Elming *et al.* (2007)
Winter coast sediments, com. Z	obs.	25.3	312.2	2.3	3.7	2.9	111011 = 5	B	555 ± 3	Popov *et al.* (2002)
	i.c.	36.1	319							
Chernokamenskay group sediments -C	obs.	17.3	306.7			6.0	111011 = 5	B	557 ± 13	Fedorova *et al.* (2014)
Podolia, comp. V	obs.	40.0	276.5			7.5	011011 = 4	C	545-570	Iosifidi *et al.* (2005)
	i.c.	47.9	285.5							
Basu-Kukkarauk formation -C	obs.	-1.1	7.3			5.8	111011 = 5	B	560	Golovanova *et al.* (2011)
	i.c.	3.5	22.4							
Krivaya Luka formation		58.6	275.8	4.8	7.7	6.1	111011 = 5	B	560-570	Lubnina *et al.* (2014)
Kurgashlya formation		50.9	314.5	4.4	6.4	5.3	111011 = 5	B	560-570	Lubnina *et al.* (2014)
Bakeevo formation		42.3	299.1	4	6.9	5.3	111111 = 6	A	560-570	Lubnina *et al.* (2014)
Fen Complex		53.2	145.7			5.6	111001 = 4	C	583 ± 15	Meert *et al.* (1998); Luleå working group new mean
Alnö carbonate complex (steep)		62.7	101.2	12.3	14.3	13.3	111010 = 4	C	584 ± 7	Meert *et al.* (2007)
Alnö carbonate complex (shallow)		3.5	269.0	16.5	32.5	23.2	001010 = 2	D		Meert *et al.* (2007)
Hailuoto sediments	obs.	48.7	241.1			8.1	011011 = 4	C	600-570	This study.
	i.c.	60.5	247.9			7.6				
Egersund dykes		31.4	44.1	14.4	17	15.6	111110 = 5	A	616 ± 3	Walderhaug *et al.* (2007), Bingen *et al.* (1998)
Nyborg formation	obs.	24.3	88.5	17.1	24.9	20.6	101110 = 4	B	653 ± 7	Torsvik *et al.* (1995)
	i.c.	38.3	82.1							

Inc. indicates inclination status for sedimentary formation where obs. = observed, and i.c. = inclination correction of f=0.6 following Torsvik et al. (2012); Plat/Plong = pole latitude/longitude; A95 = confidence circle of the pole; Q_{1-6} =

Van der Voo reliability criteria (1 to 6) of paleomagnetic poles (Van der Voo, 1990), Class = quality of pole (Meert, 2014).

Formation		Plat (°N)	Plon (°E)	Euler poles Lat	Euler poles Long	Euler poles angle	Age (Ma)	Reference
Baltica								
Egersund dykes		-31	224	-13	109	127	616 ± 3	Walderhaug *et al.* (2007), Bingen *et al.* (1998)
Hailuoto	*obs.*	-49	61	-12	294	147	600 - 570Ma	This study
sediments	*i.c.*	-61	68	-10	296	159		
Mean 570 Ma		-52	118	18	96	166	570	*[1] This study
Mean 550 Ma	*obs.*	-30	111	16	51	129	550	*[2] This study
	i.c.	-42	116	13	56	139	550	
Laurentia								
Long Range dykes		-19	175	0	265	-109	615 ± 2	Murthy *et al.*1992
Baie de Mouton (A)		-43	153	0	243	-133	583 ± 2	McCausland *et al.* 2011
Baie de Mouton (B)		34	142	0	52	56		
*Callandar alkaline complex		-46	121				575 ± 5	Symons and Chiasson (1991)
Catoctin Basalts		-42	117	0	207	-132	572 ± 5	Meert *et al.* 1994
Sept Iles intrusion		20	142	0	52	70	565 ± 4	Tanczyk *et al.* 1987
Skinner Cove		16	175	0	85	74	550 ± 3	McCausland and Hodych, 1998
*Florida Mountains auriole		-6	169				503 ± 10	Geissman *et al.* (2012)

Plat, Plon is latitude and longitude of the paleomagnetic pole. The Euler poles are used in the reconstruction in Fig.7. *[1] The 570 Ma mean pole was calculated from the Krivaya Luka, Kurgashlya and Bakeevo poles (Lubnina et al., 2014). *[2] The 550 Ma mean pole was calculated from the Verkhotina comp. Z (Popov et al., 2005) and Zolotitca comp. B (Iglesia Llanos *et al.*, 2005) poles. Laurentia poles marked with * were included in the APWP of Laurentia (Fig. 6B), but not in the reconstruction (Fig . 7), therefore the euler poles are not provided for them.

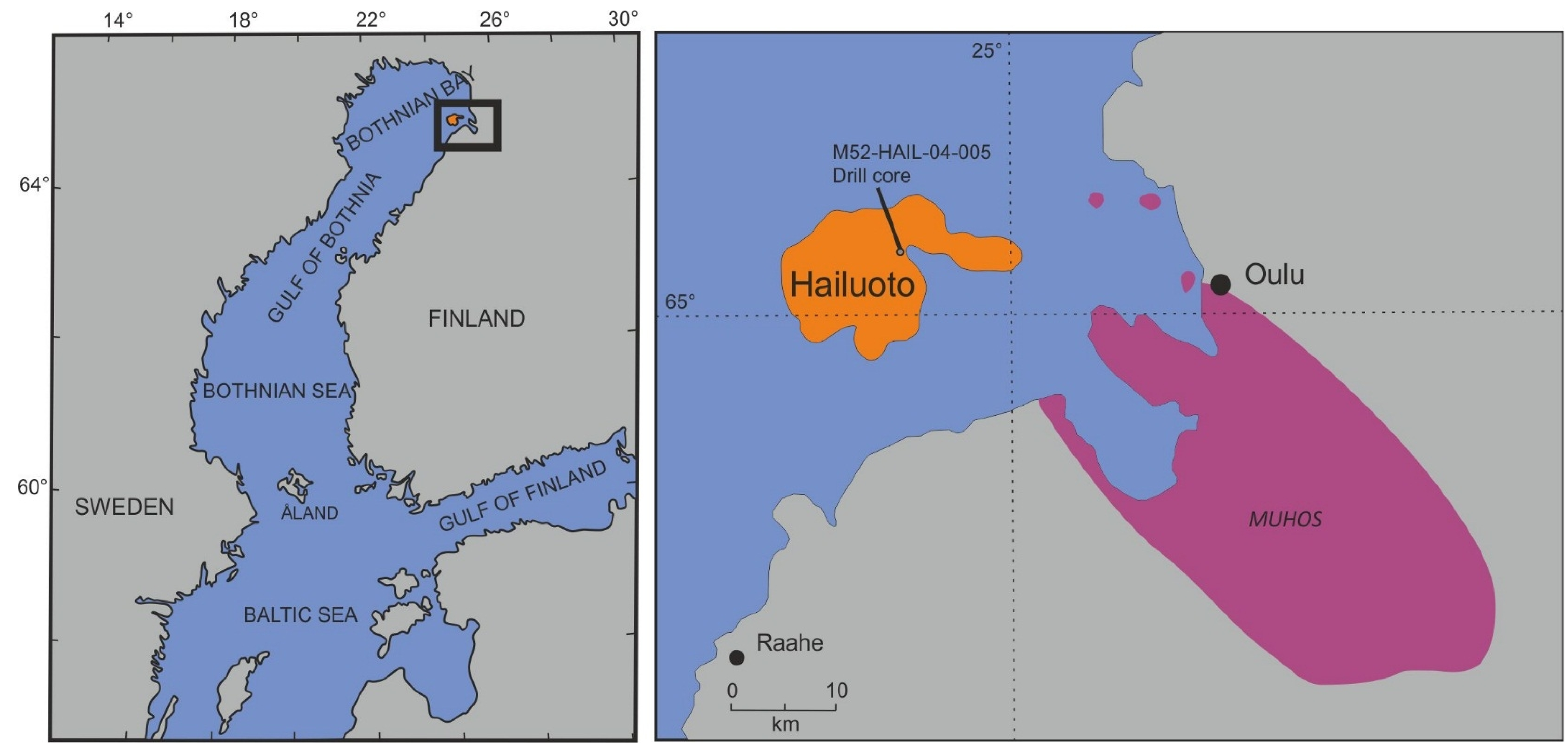
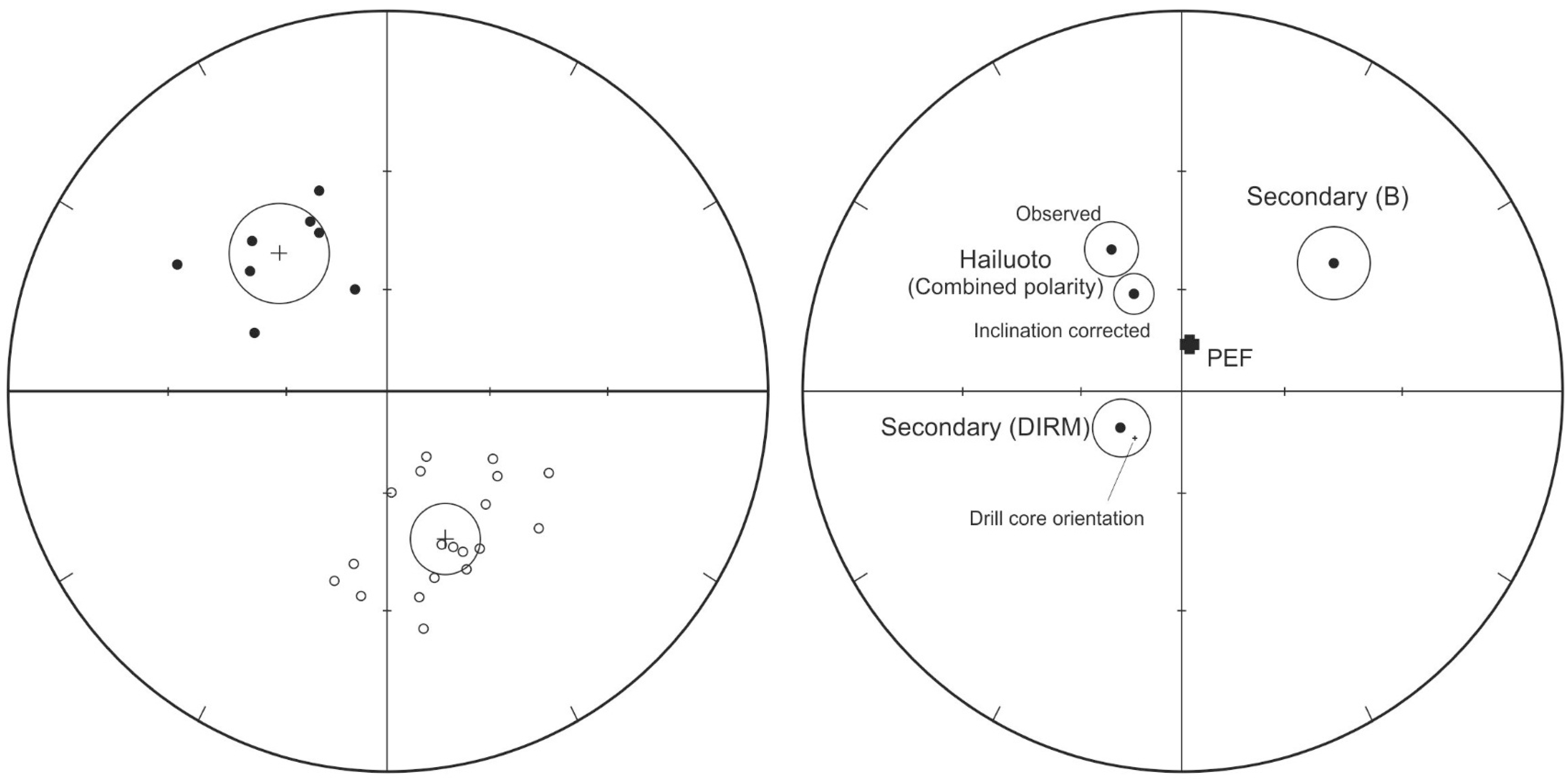


Fig 1.



Primary (observed)

Mean directions

Fig. 3

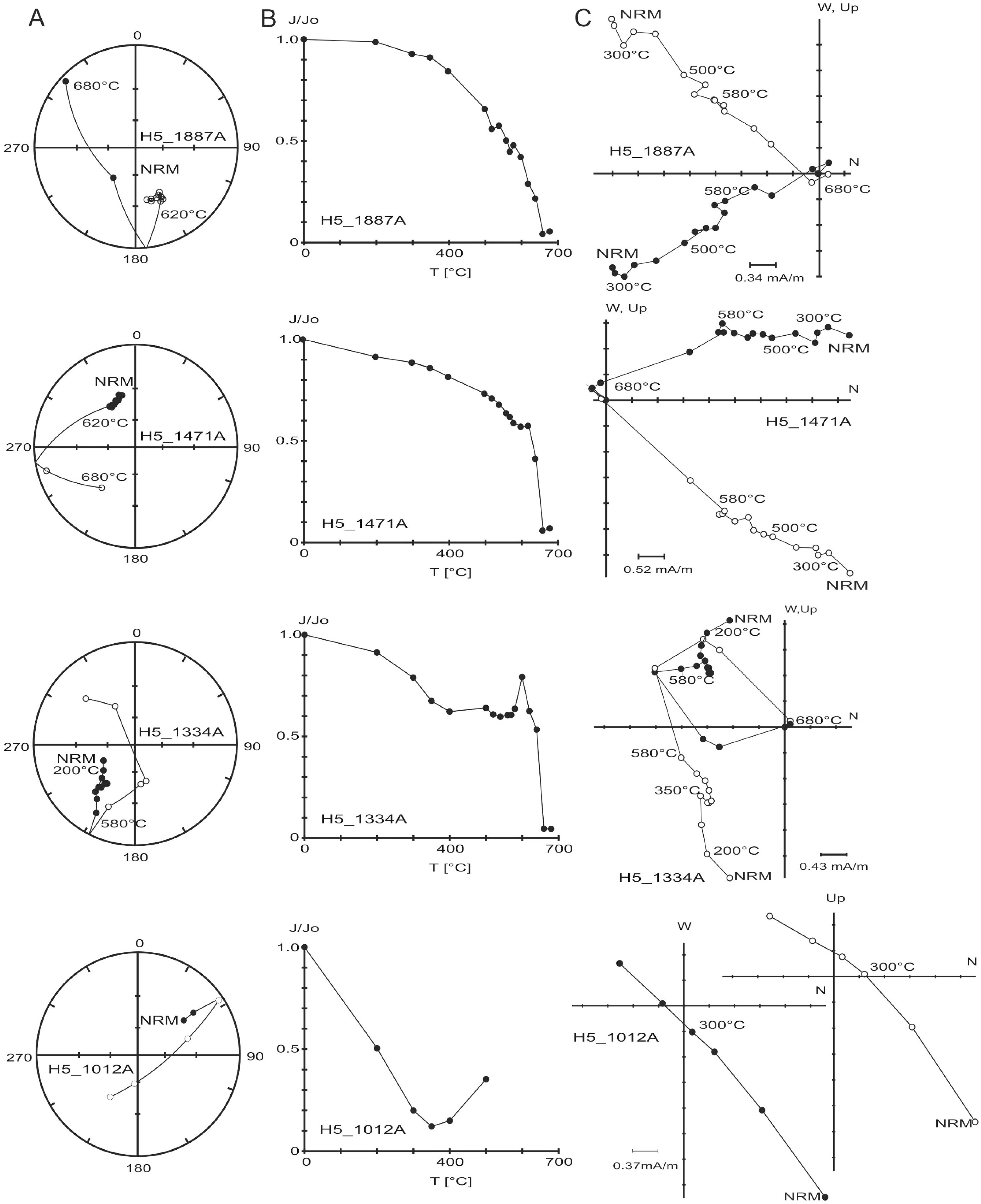

A
B
C
H5_1887A
H5_1471A
H5_1334A
H5_1012A
J/Jo
T [°C]
NRM
0.34 mA/m
0.52 mA/m
0.43 mA/m
0.37mA/m
W, Up
N

Fig. 2

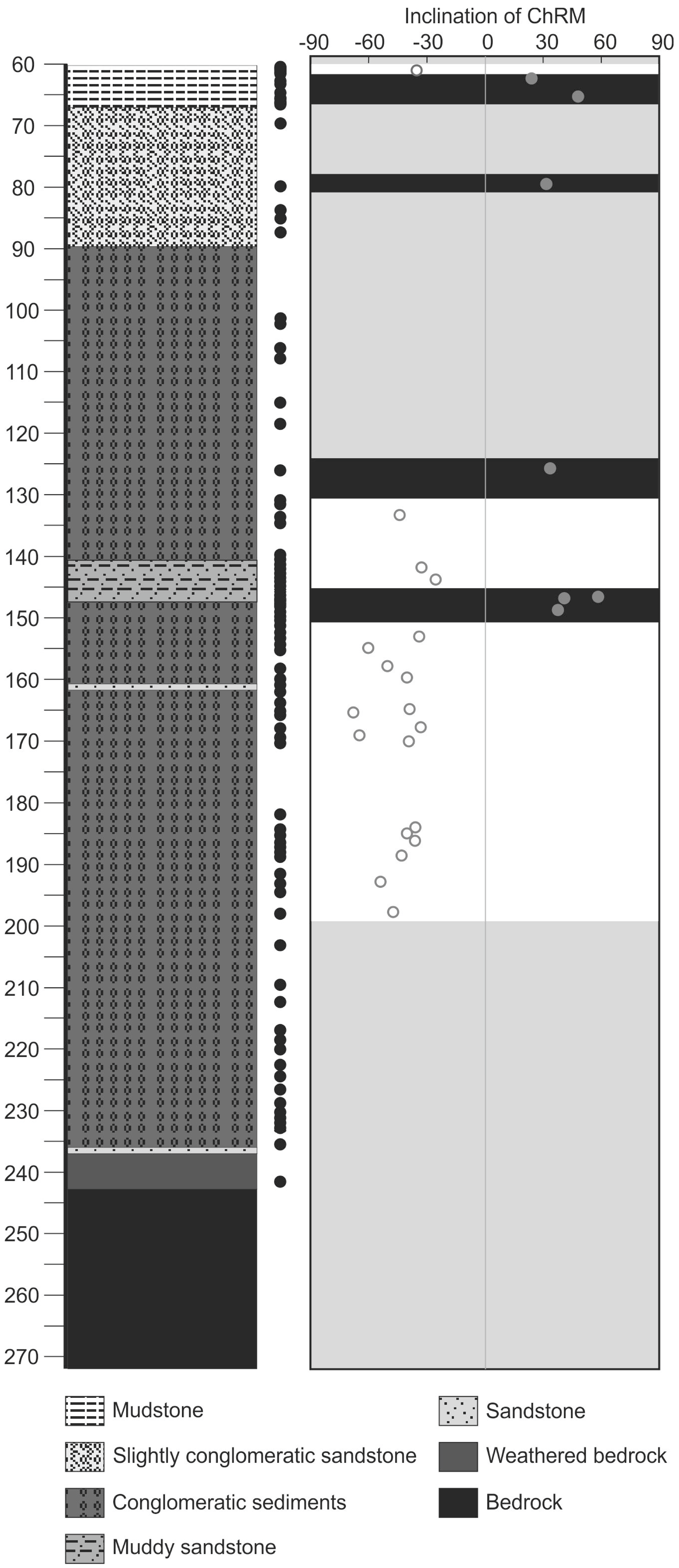

Inclination of ChRM
-90
-60
-30
0
30
60
90
60
70
80
90
100
110
120
130
140
150
160
170
180
190
200
210
220
230
240
250
260
270
Mudstone
Slightly conglomeratic sandstone
Conglomeratic sediments
Muddy sandstone
Sandstone
Weathered bedrock
Bedrock

Fig. 4

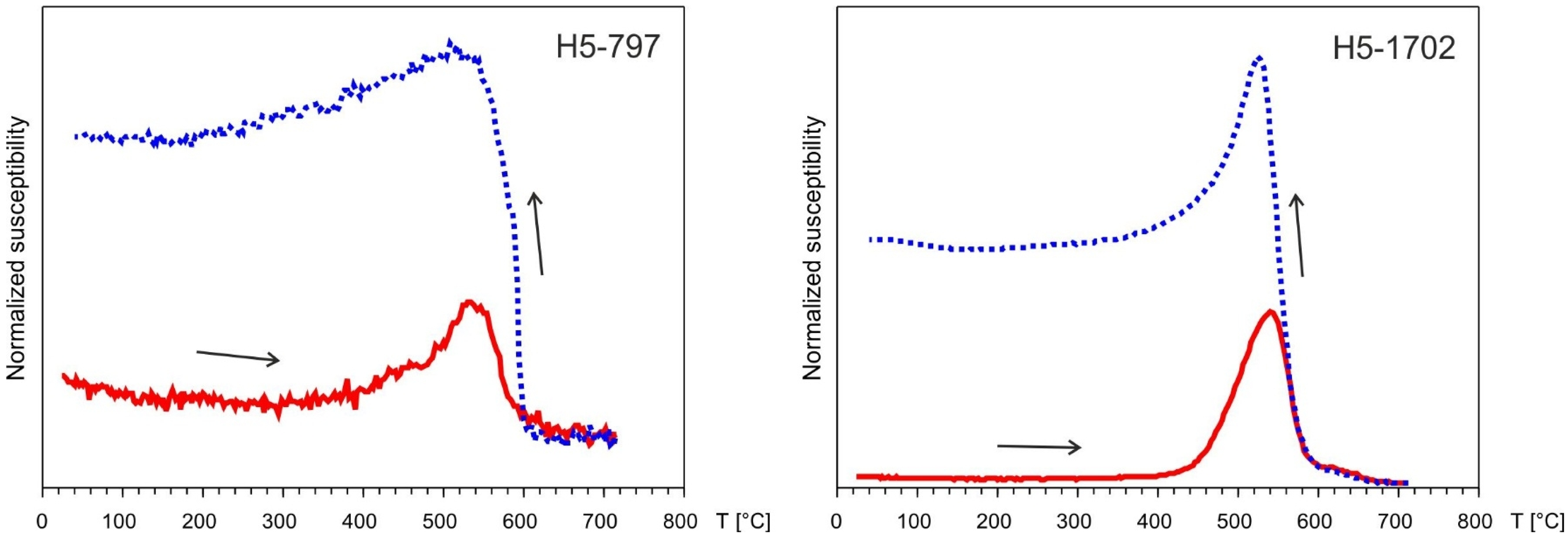

H5-797
H5-1702
Normalized susceptibility
Normalized susceptibility
0
100
200
300
400
500
600
700
800
T [°C]

Fig. 5

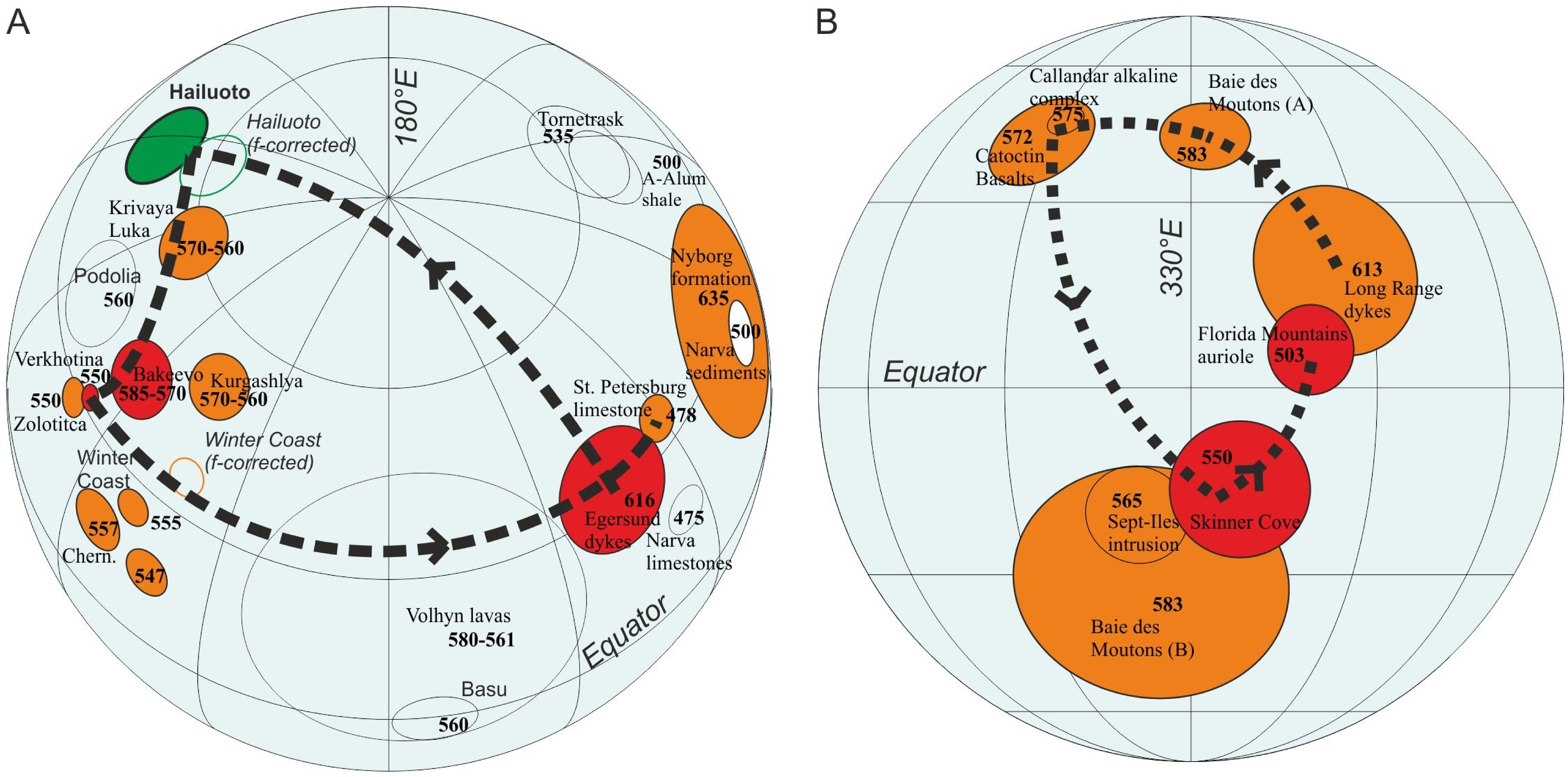

A
Hailuoto
Hailuoto
(f-corrected)
180°E
Tornetrask
535
500
A-Alum
shale
Krivaya
Luka
570-560
Podolia
560
Nyborg
formation
635
500
Narva
sediments
Verkhotina
550
Bakeevo
585-570
Kurgashlya
570-560
550
Zolotitca
St. Petersburg
limestone
478
Winter Coast
(f-corrected)
Winter
Coast
557
555
Chern.
547
616
Egersund
dykes
475
Narva
limestones
Volhyn lavas
580-561
Equator
Basu
560
B
Callandar alkaline
complex
575
572
Catoctin
Basalts
Baie des
Moutons (A)
583
330°E
613
Long Range
dykes
Florida Mountains
auriole
503
Equator
550
565
Sept-Iles
intrusion
Skinner Cove
583
Baie des
Moutons (B)

Fig. 6

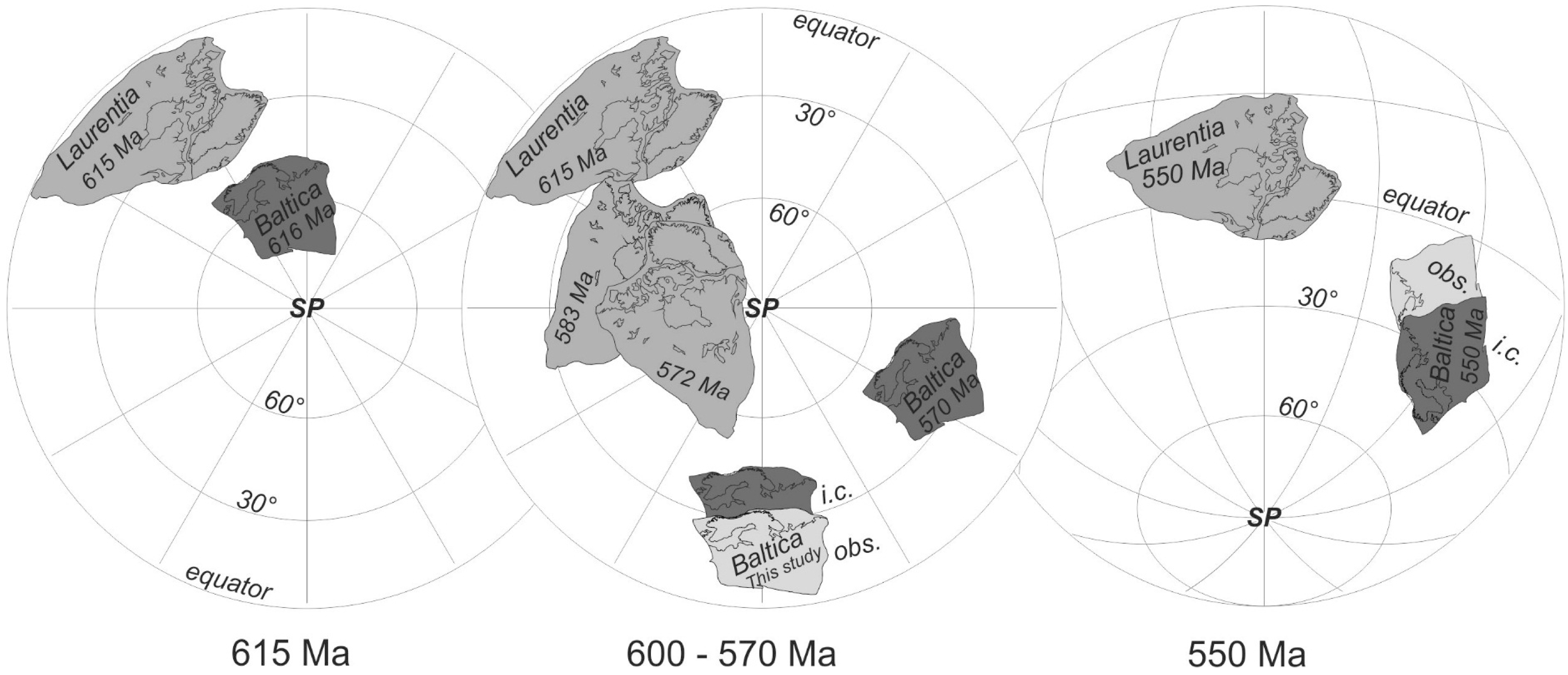

Laurentia
615 Ma
Baltica
616 Ma
SP
60°
30°
equator
615 Ma
equator
30°
60°
Laurentia
615 Ma
583 Ma
572 Ma
SP
Baltica
570 Ma
i.c.
Baltica
This study
obs.
600 - 570 Ma
Laurentia
550 Ma
equator
obs.
30°
Baltica
550 Ma
i.c.
60°
SP
550 Ma

Fig. 7