Palaeomagnetism and $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations of the Ediacaran traps from the southwestern margin of the East European Craton, Ukraine: relevance to the Rodinia break-up

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Abstract: A palaeomagnetic study and age determinations have been performed on Ediacaran basalts from the northwestern Ukraine. Whole-rock $^{40}\text{Ar}/^{39}\text{Ar}$ age determination revealed plateau ages at 590–560 Ma and 393 Ma, the latter probably reflecting a resetting of the radiometric system. Palaeomagnetic poles have been calculated from five basalt flows, two of which (A poles) are considered reliable with ages that range from 580 to 560 Ma. Tentative poles (B poles), calculated from most probably primary magnetizations, have ages estimated at 580–545 Ma. Secondary magnetizations, possibly of late Ediacaran or Devonian age, have also been isolated (C poles). Based on the new poles, Baltica drifted together with Laurentia from an equatorial position at c. 750 Ma to occupy high southern latitude positions at c. 580 Ma. Baltica during that time period was joined to Laurentia in a similar relative position to that at 750 Ma. The two shields then split up from each other and from c. 550 Ma Baltica drifted at moderately high latitudes and rotated some 180° during the final opening of the Iapetus ocean. This reconstruction suggests that during the Ediacaran glaciation Baltica occupied high-latitude positions, which contradicts the high-obliquity model to explain low-latitude Neoproterozoic glaciations.

The Ediacaran (542–600 Ma)–Cambrian transition is one of the most puzzling periods in Earth history. The final break-up of the Rodinia supercontinent and the immediate formation of another supercontinent, Gondwana, is coeval with the sudden appearance of the animal phyla and abrupt shifts in oceanic and atmospheric geochemistry (e.g. Meert & Lieberman 2004). A general understanding of the origin and evolution of Rodinia has evolved during the past few years from new geodynamic models, which are based on only the most reliable palaeomagnetic data from Baltica and Laurentia (e.g. Torsvik *et al.* 1996; Meert & Torsvik 2003; Pesonen *et al.* 2003). However, the subsequent history of Rodinia break-up and the independent drift of Baltica and Laurentia still remain poorly substantiated.

Attempts to define the relationships between these two continents have not resulted in any conclusive tectonic scenario, mainly because of the lack of high-quality palaeomagnetic data for the Late Precambrian. When the palaeomagnetic reliability criteria of Van der Voo (1990) are applied to the Neoproterozoic datasets of Baltica, the number of acceptable poles ($Q \geq 3$) for the Ediacaran is reduced to only 3–5 (Torsvik *et al.* 1996; Meert & Torsvik 2003). New Ediacaran poles and tectonic scenarios for Baltica have been presented by Popov *et al.* (2002), Nawrocki *et al.* (2004a, b) and Llano *et al.* (2005). The poles of Popov *et al.* (2002) passed field tests and are considered as primary magnetizations. Popov *et al.* (2002) and Llanos *et al.* (2005) also questioned the earlier presented Late Proterozoic–Early Ordovician apparent polar wander paths (APWPs) (Torsvik *et al.* 1996; Smethurst *et al.* 1998), thus the number of ‘reliable’ poles is unclear.

To test the geodynamic interpretations for Baltica additional reliable Ediacaran poles are thus needed. In this palaeomagnetic study samples have been collected from basalts of the Ediacaran traps, Ukraine, which range in age between 530 and 605 Ma (Velikanov & Korenchuk 1997). Earlier palaeomagnetic investigations have identified a stable remanence in these basalts (Glevasskaya *et al.* 1997, 2001; Nawrocki *et al.* 2004a). Based on these promising results, the present study is aimed at expanding the Ediacaran dataset for Baltica.

General geology

In Ukraine the Ediacaran rocks are subdivided into the Volyn, Mogilev–Podolia and Kanilov series (Sokolov & Fedonkin 1990) of sedimentary and volcanogenic rocks, which are exposed in the Peridnister region and near the town of Rivne. However, the sedimentary and volcanogenic rocks are not exposed in the same outcrops and the stratigraphic position of the basalts can therefore only be obtained from borehole information.

The basalts of the Volyn series, and their equivalents, occur on the western margin of the East European Craton, and occupy some 200 000 km² of the subsurface from Moldavia to Lithuania (Fig. 1a; Bialowolska *et al.* 2002). The traps are located on the southwestern erosional boundary of the Ukrainian Shield. The maximum thickness of the basalts is 225 m (Velikanov & Korenchuk 1997) and their distribution is controlled by NW–SE-trending faults. The isotopic and geochemical characteristics of these basalts are very similar to those of basalts from other continental trap provinces (Volovnik 1971; Bialowolska *et al.*

Fig. 1. (a) Map of the East European Platform, showing the boundaries separating different crustal segments, and location of the basalts of the Volyn series (dotted boundary). (b) Map showing the geology and location of the sampling sites in the northwestern part of Ukraine.

2002). The Ediacaran volcanic activity was intermittent, and the basaltic sheets are interbedded with tuffs and sedimentary rocks of different types. The effusive rocks of the Volyn series are derivatives of weakly differentiated tholeiitic magma.

Geological description of the sampling sites

The location of sampling sites is shown in Figure 1b. The sampling was restricted to rock quarries and the basalts and tuffs were collected from two areas located some 50 km apart. The northern area was sampled in two quarries situated on the western and eastern sides of the village Ivanchi (51.23°N, 26.06°E; sites 1 and 2, respectively). In the southern area the samples were collected near the villages Berestovets (50.86°N, 26.30°E; site 3) and Basaltovoe (50.86°N, 26.18°E; site 4).

The outcrops in the southern area are mainly characterized by massive weakly differentiated tholeiitic basalts with subvertical pillar jointing. Their thickness exceeds 20 m. The basaltic rocks are well-crystallized anamesite and dolerite and there are small numbers of 2–3 mm pyroxene (augite) inclusions. In the basalts there are often microxenocrysts (0.1–1.0 mm) of titanomagnetite and ilmenite, which originate from the early stage of magmatism. Titanomagnetite of late magmatic origin (1–3 µm) is sometimes also observed. The ore mineral content generally reaches 5–6%.

In the Berestovets (site 3) outcrop, a c. 12 m thick basalt flow is exposed in a quarry and samples were collected from several parts of it. A thick lava flow (over 20 m) is also exposed in the quarry at site 4.

In the northern area two basaltic flows interbedded with tuffs are exposed in a small quarry near the village Ivanchi (site 1). A prominent subhorizontal zone of baked tuffs is observed in contact with the basal boundary of the upper flow. The tuffs are brick-red and dense at the contact, but become looser and greenish grey with increasing distance from the basalt. The apparent thickness of the upper and lower flow is 6–8 m and up to 5 m, respectively. Quartz, chalcedony, red zeolite and black palagonite amygdaloids are common within the flows.

In the village of Ivanchi there is another quarry, the Great Rafalovka

quarry (site 2), with a relatively thick (over 20 m) basaltic flow. The basalt is massive with subhorizontal banding and is slightly amygdaloidal. The flow is underlain by psammitic tuffs, which are baked by the basalts.

Radiometric dating of the basalt

A $^{40}\text{K}/^{40}\text{Ar}$ whole-rock age of 560 Ma was derived from the basalt in the quarry at Basaltovoe (Savchenko *et al.* 1984). Using new decay constants and isotopic relationships (Steiger & Jäger 1977) this age of the basalts has been recalculated to be 545.5 Ma.

New $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations are here presented for the basalts from Ivanchi (site 1), Rafalovka (site 2) Berestovets (site 3) and Basaltovoe (site 4). Palaeomagnetic core samples were subjected to $^{40}\text{Ar}/^{39}\text{Ar}$ analysis at the geochronology laboratory of the University of Alaska Fairbanks. The cores were crushed, washed and sieved to obtain small (c. 0.5 mm) whole-rock chips for dating. The monitor mineral MMhb-1 (Samson & Alexander 1987) with an age of 513.9 Ma (Lanphere & Dalrymple 2000) was used to monitor neutron flux (and calculate the irradiation parameter, J). The samples and standards were wrapped in aluminium foil and loaded into aluminium cans of 2.5 cm diameter and 6 cm height. The samples were irradiated in position 5c of the uranium-enriched research reactor of McMaster University in Hamilton, Ontario, Canada, for 20 MW h.

Upon their return from the reactor, the sample and monitors were loaded into 2 mm diameter holes in a copper tray that was then loaded in a ultrahigh-vacuum extraction line. The monitors were fused, and samples heated in a stepwise manner, using a 6 W argon-ion laser following the technique described by York *et al.* (1981), Layer *et al.* (1987) and Layer (2000). Argon purification was achieved using a liquid nitrogen cold trap and a SAES Zr–Al getter at 400 °C. The samples were analysed in a VG-3600 mass spectrometer at the Geophysical Institute, University of Alaska Fairbanks. The argon isotopes measured were corrected for system blank and mass discrimination, as well as calcium, potassium and chlorine interference reactions following the procedures outlined by

McDougall & Harrison (1999). Plateau ages were based on the criteria of three or more contiguous fractions (with a roughly constant Ca/K ratio) accounting for more than 50% gas release, having a mean square weighted deviation (MSDV) less than 2.5. A summary of all the $^{40}\text{Ar}/^{39}\text{Ar}$ results is given in Table 1, with all ages quoted to the $\pm 1\sigma$ level and calculated using the constants of Steiger & Jäger (1977). Integrated and plateau ages include error in the irradiation parameter (0.5%).

Of the seven samples dated, four (U99.97 and U106.97, site 2; U87.97, site 3; U145.96, site 4) with ages at 580 ± 9 Ma and 684 ± 123 Ma, 393 ± 8 Ma and 582 ± 7 Ma, respectively, meet the criteria for a plateau (Fig. 2), although the 684 ± 123 Ma age is not very precise because of low K content. Sample U118.00 (site 1; Fig. 2) does not meet the criteria for a plateau, with only 35% release as a result of excess argon at low temperatures and young ages associated with a high C/K phase (>25) at high temperatures. However, its age of 561 ± 14 Ma is consistent with those for other samples from the area. Two samples (U88.97, site 3; and U 144.96, site 4) have higher than acceptable MSWD values (3.7 and 11.0, respectively) as a result of either argon loss or excess argon.

Most samples contain less than 1% non-radiogenic argon, and this lack of isotopic variations makes isochron analysis impossible. Only the sample from site 2 has a sufficient isotopic variation to allow for an isochron. The isochron age of 579 ± 12 Ma is identical to the plateau age for the same sample (580 ± 9 Ma). The sample does not show evidence of excess argon with a $^{40}\text{Ar}/^{36}\text{Ar}$ initial value of 292 ± 27 (Fig. 2).

In general, the samples fall into two groups. In the first are basalts with an age of 580–590 Ma as seen in four samples. There is evidence of some excess Ar in these samples (e.g. site 1), again indicative of some post-cooling alteration; however, none have well-constrained isochrons showing the excess argon and we do not feel that the resetting was sufficient to affect the

interpreted age. In the second group the basalt samples U87.97 and U88.97 have plateau ages of 393 and 369 Ma, respectively. Both show evidence of severe argon loss, and diffusion loss modelling indicates the true age to be significantly higher. Thus the plateau age of 393 Ma for sample U87.97 may reflect a cooling or blocking age; however, the true closure age is older, having been partially reset at a later time.

Shumlyanskyy & Andréasson (2004) isolated zircon crystals from the basalt sampled in a drill-hole not far from the Rafalovka quarry in the northern area. U–Pb dating on four crystals yielded a concordant age of 576 ± 14 Ma, similar to our $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 561 ± 13 Ma and 580 ± 9 Ma determined from the Ivanchi and Rafalovka quarries (sites 1 and 2). A second set of four crystals yielded a 'nearly' concordant age of 514 ± 8 Ma. This age is significantly younger and may indicate a complicated isotopic system that may have been reset, as indicated by the fact that the zircons are not completely concordant.

Palaeomagnetic sampling and laboratory technique

The samples were collected using a portable drilling machine as block samples and were oriented with magnetic and sun compasses. Independent teams from the Academy of Sciences in Kiev and Luleå University of Technology collected the samples. A total number of 148 samples were collected for this study, of which 93 were field drilled samples.

Standard thermal and alternating field (AF) demagnetization procedures were applied and the laboratory work was carried out in the palaeomagnetic laboratories of Luleå University of Technology, Sweden and Ludwig-Maximilians-Universität, Munich, Germany. Schoensted TSD-1 furnaces were used for the stepwise thermal demagnetization and the remanence measurements, and the AF demagnetizations were performed using 2G cryogenic magnetometers. At Luleå University of Technology, the anisotropy of magnetic susceptibility (AMS) and its

Table 1. $^{40}\text{Ar}/^{39}\text{Ar}$ whole-rock ages of basalts from the Ukrainian part of the Ediacaran volcanic trap formation in the western part of the East European Platform

Site	Integrated age (Ma)	Plateau age (Ma)	Plateau information	Other comments
Site 1 Upper basalt U118.00	587 ± 9	561 ± 14	5 fractions 53% ^{39}Ar release MSWD = 2.9	Excess Ar, does not include high-Ca fractions
Site 2 U99.97	537 ± 10	580 ± 9	5 fractions 63% ^{39}Ar release MSWD = 0.8	Good plateau Isochron: 579 ± 12 Ma
Site 2 U106.97	373 ± 88	684 ± 123	5 fractions 43% ^{39}Ar release MSWD = 0.6	Low K content, not precise
Site 3 U87.97	340 ± 7	393 ± 8	11 fractions 74% ^{39}Ar release MSWD = 1.2	Ar loss, but good plateau
Site 3 U88.97	275 ± 6	369 ± 12	9 fractions 40% ^{39}Ar release MSWD = 3.7	Severe Ar loss, maximum age at c. 500 Ma
Site 4 U144.96	569 ± 5	593 ± 5	4 fractions 59% ^{39}Ar release MSWD = 11	Isochron 597 ± 23 Ma Initial = 2093 ± 1633
Site 4 U145.96	602 ± 4	582 ± 7	8 fractions 41% ^{39}Ar release MSWD = 1.4	Excess Ar, plateau is last 41% release

Fig. 2. (a, b, d and e) $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra from whole-rock basalt samples of sites 1–4; (c) an inverse isochron plot for the sample from site 2. The samples were obtained from palaeomagnetic drill cores. (See text for analytical procedures.)

variation with temperature were measured with a Kappa bridge KLY-3 and CS-3 instrument, respectively (Geofyzika, Brno).

The demagnetization results were analysed using orthogonal plots and stereographic projections, and the remanence directions and site means were defined using principal component analysis (Kirschvink 1980) and Fisher statistics (Fisher 1953), respectively. Generally, no directions defined with a mean angular deviation (MAD) higher than 10° were accepted for calculation of site means.

Palaeomagnetic results

The results from the analyses of the two collections of samples in this study are generally consistent (see Table 2 and Fig. 3). The calculated site mean directions, presented below, are therefore based on the combined results from the sample directions of the two teams.

Samples from the same outcrops at three of the sites have also been collected by colleagues from the All-Russia Petroleum Research and Geological Prospecting Institute in St. Petersburg, Russia. The results from that work are so far presented only in

abstracts (Iosifidi *et al.* 2001a, b), but are consistent with the data from the present study (Fig. 3).

The northern area

The natural remanent magnetization (NRM) of the upper level basalt at site 1 (Ivanchi) is of high coercivity (H_c) with a median destructive field (MDF; Dunlop & Özdemir 1997) of 15–30 mT (Fig. 4a). Thermal demagnetization reveals a low unblocking temperature (T_{ub}) component (0–400 °C) with a direction close to that of the PEF and a high H_c component unblocking at 510 °C (Fig. 4b), indicating fine-grained titanomagnetite to be the carrier (declination (D) = 31.8° , inclination (I) = -58° ; Table 2; Fig. 3). Visual inspection of the Upper Ivanchi basalts shows a brownish red pigment on their surface. Optical examination and microprobe analysis using a Camebax system indicate that this pigment consists mainly of hematite and maghemite. This may explain the low-temperature component, which forms only a small part of the total NRM. The virtual geomagnetic pole (VGP) position calculated from high- H_c and high- T_{ub} magnetization plots at 4°N , 1.6°E (Table 2).

The remanent magnetization of the lower basalt layer at site 1 is also of high coercivity with a MDF of 15–50 mT (Fig. 4e). A high-coercivity component ($D = 36.7^\circ$, $I = -82.0^\circ$; Table 2) is identified after demagnetizations in fields of c. 40 mT and in temperatures above 350 °C. During thermal demagnetizations c. 75% of the intensity drops at temperatures between 530 and 590 °C (Fig. 4f), suggesting that titanomagnetite is the main carrier of this remanence. The VGP calculated from the sample mean direction for the three datasets of the high- T_{ub} magnetization of this Lower Ivanchi basalt is located at 38.4°N , 14.5°E (Table 2; Fig. 5).

The volcanic tuff that separates the upper and lower basalt layers also carries stable remanent magnetizations of different coercivities and unblocking temperatures. In the green tuff deposited on top of the lower basalt, thermal demagnetization reveals a high unblocking component at $D = 186.6^\circ$, $I = 83.2^\circ$, probably carried by hematite (isolated at 550–650 °C; Fig. 4d; Table 2). In the baked red tuff in contact with the basal part of the upper basalt a high- H_c magnetization (>60 mT; Fig. 4c) is isolated, with a direction ($D = 23.7^\circ$, $I = -47^\circ$; Table 2) that is similar to that of the upper basalt and most probably a result of remagnetization.

The NRM of the basalt at site 2 (Rafalovka) is of rather high coercivity with an MDF at c. 15 mT (Fig. 4g). Characteristic magnetizations with well-defined directions are isolated in alternating fields higher than 25 mT ($D = 270.6^\circ$, $I = 74.2^\circ$; Table 2; Fig. 3). The unblocking temperature spectrum is fairly wide, however, with a c. 35% drop in intensity between 500 and 570 °C, suggesting that titanomagnetite is the carrier of remanence (Fig. 4h). The VGP calculated from the mean directions plots at 43°N , 343.7°E (Table 2; Fig. 5).

Measurements of AMS of the upper basalt at site 1 show that the basalt is magnetically foliated, with a foliation plane that trends 153°E and dips 34° . For multi-domain magnetite the axis of k_{max} is parallel to the long axis of an ellipsoidal grain. In flowing magma elongated particles line up parallel to the flow direction (Knight & Walker 1988) and when the magma crystallizes the oxides will mimic the fabric of the pre-existing silicate phases (Tauxe *et al.* 1998). Here the mean direction of k_{max} indicates a flow direction at c. $D = 272^\circ$ and $I = 30^\circ$, which is close to the dip direction of the foliation plane. This may suggest that this site is close to the volcanic centre of the upper basalt. Also, the lower basalt is magnetically foliated with a

Table 2. Palaeomagnetic data from the basalt sampled in four quarries in NW Ukraine and calculated mean of sites

Site	<i>N, n</i>	<i>D</i>	<i>I</i>	α_{95}	<i>K</i>	<i>P</i> _{lat}	<i>P</i> _{long}	<i>A</i> ₉₅	Group or reference
Site 1, Ivanchi (51.23°N, 26.03°E)									
<i>Luleå</i>									
Upper basalt	20	32.7	−59	2.7	152	5.1	1.3	3.5	
Lower basalt	7	23.1	−81	5.4	128	35.1	17.9	10.2	
Tuff, red upper high <i>H</i> _c	3	6.7	−41.3	12.4	100	−14.8	19.7	11.8	A3*
Tuff, green–red, high <i>T</i> _{ub}	3	186.6	83.2	5.3	536	37.9	24.1	10.3	A4
<i>Kiev</i>									
Upper basalt	6	29	−55	6.9	95	0.6	2.9	8.2	
Lower basalt	12	46.2	−83	7.2	38	40.2	12.3	8.3	
Tuff, red upper high <i>H</i> _c	3	44.8	−47	37.1	12	1	348.7	40.5	
Mean upper basalt	2*, 26	30.8	−57			2.8	2.0		A3
	2, 26*	31.8	−58	2.5	134	4	1.6	3.2	A3
Mean lower basalt	2*, 19	33.6	−82			37.7	15.2		A2
	2, 19*	36.7	−82	4.7	52	38.4	14.5	9.0	A2
Mean red tuff	2*, 6	24.8	−45.8			−8.6	3.8		A3*
	2, 6*	23.7	−47	18.5	14	−7.8	5.1	19.2	A3*
<i>St. Petersburg</i>									
Upper basalt	12	23	−57	5.3	68	1.1	8	6.6	Iosifidi <i>et al.</i> 2001a
Lower basalt	5	114	−83	4.8	251	44.1	43.7	9.3	Iosifidi <i>et al.</i> 2001a
Tuff	9	6.5	−63.3	3.5	213	6.2	21.4	4.9	Iosifidi <i>et al.</i> 2001a
Site 2, Rafalovka (51.23°N, 26.06°E)									
<i>Luleå</i>									
High <i>H</i> _c	16	264.9	72.4	2.6	202	38.9	342.7	4.3	
<i>Kiev</i>	7	276.1	76.6	5.3	131	47.1	347.1	9.4	
Mean	2*, 23	269.8	74.6			43	344.7		A1
	2, 23*	270.6	74.2	2.4	155	43	343.7	4.1	A1
<i>St. Petersburg</i>	17	283	75	4.2	75	48.9	341.5	7.3	Iosifidi <i>et al.</i> 2001a
Site 3, Berestovets (50.86°N, 26.30°E)									
<i>Luleå</i>									
High <i>H</i> _c	15	111.5	−19	6.3	37	21.1	285.6	4.7	
Low <i>H</i> _c	13	83.2	53.7	6.8	38	29.9	97.5	7.9	
Low <i>T</i> _{ub} (200–450 °C)	5	85.5	42.4	7.5	103	21.5	103.5	7.2	
<i>Kiev</i>									
High <i>H</i> _c	8	124	−21	8.2	46	29.5	275	6.2	
Low <i>H</i> _c	7	77	51	10.1	37	31.9	101.8	11.2	
Mean high <i>H</i> _c	2*, 23	117.7	−20.1			25.4	280.8		C1
	2, 23*	115.8	−19.8	5	34	24.1	282.3	3.7	C1
Mean low <i>H</i> _c – <i>T</i> _{ub}	2*, 12	81.6	46.8			26.5	103.6		B1
	2, 12*	80.9	47.5	6.1	44	27.3	103.6	6.3	B1
Site 4, Basaltovoe (50.86°N, 26.18°E)									
<i>Luleå</i>									
High <i>H</i> _c	12	120.7	−23.2	5.2	70	28.5	279.3	4	
Low <i>H</i> _c	9	85.5	46.2	9	34	23.7	101.1	9.2	
Low <i>T</i> _{ub} (200–450 °C)	11	95.3	44.8	11.8	16	17	95	11.8	
<i>Kiev</i>									
High <i>H</i> _c	10	109	−16	7.3	45	18.3	286.5	5.4	
Low <i>H</i> _c	10	64.6	48.8	10.3	23	38.3	114.4	11	
Mean high <i>H</i> _c	2*, 22	114.7	−19.7			23.3	283		C2
	2, 22*	115.6	−20	4.9	41	24	282.4	3.7	C2
Mean low <i>H</i> _c – <i>T</i> _{ub}	2*, 21	80.5	47.8			27.8	103.5		B2
	2, 21*	81.1	47.7	8.6	20	27.3	103.2	9.0	B2
<i>St. Petersburg</i>									
High <i>H</i> _c	6	110.9	−9.1	11.3	36	16.6	282.6	7.9	Iosifidi <i>et al.</i> 2001a
Low <i>H</i> _c	15	67.8	55.7	4.3	79	40.6	105.8	5.2	Iosifidi <i>et al.</i> 2001a
Rafalovka									
(A1)	29	269	66	2.4	117	36	333	3.5	Nawrocki <i>et al.</i> 2004a
Basaltovoe									
(B2)	21	67	55	3.9	67	40	102	4.9	Nawrocki <i>et al.</i> 2004a
(C2)	21	114	−26	3.9	68	26	282	2.8	Nawrocki <i>et al.</i> 2004a
Mean A1	3*/52	269.4	71.7	8.5	211	40.3	339.7	8.7	A1
Mean B2	3*/42	76.4	50.4	19	43	31.9	104.5	20.9	B2
Mean C2	3*/43	114.5	−21.8	11.5	116	24.1	284.0	8.8	C2
Mean of sites									
5: A1, A2, A4	3*/74	240.5	80.9	15.1	68	40.4	5.6	28.6	c. 580
6: A1, A2, A3, A3*, A4	5*/106	216.4	70.2	17.7	20	19.8	4.4	28.3	580–561
7: B1, B2	2*/54	78.7	49			29.6	104.1		582–545
C1, C2	2*/66	114.5	−21.8			24.1	283.1		Late Ediacaran? Devonian?

N, number of data collections; *n*, number of samples included in the calculated mean; *D*, *I*, mean declination and inclination, respectively; α_{95} , cone of confidence at a 95% probability level; *P*_{lat}, *P*_{long}, latitude and longitude of the palaeomagnetic pole; *A*₉₅, confidence circle of the pole at a 95% probability level. Site means used for calculation of palaeomagnetic mean poles are shown in bold type. For the calculation of site mean directions the results from sites 1, 2 and 4 of Iosifidi *et al.* (2001a, b) are not included, as those results so far have been reported only in abstracts. Data of Nawrocki *et al.* (2004a) are included in the calculation of site means A1, B2 and C2 and thus also included in the mean of sites and mean palaeomagnetic poles 5–7.

*Level of the mean calculation.

Fig. 3. Palaeomagnetic site mean directions with 95% confidence limits for basalts from the four quarries. (a) Mean directions for sites 1–3 as calculated from the results of the teams from Kiev and Luleå in this study (Table 2) and from Iosifidi *et al.* (2001a, b; sites 1 and 2; Table 2). (b) Mean directions (two components) for site 4 calculated by the two teams and from Iosifidi *et al.* (2001a, b). (c) Mean directions calculated from samples, combining the results from the two collections of samples in this study (Table 2). Open (closed) symbols denote negative (positive) inclinations.

foliation plane that trends 349°E and dips 18° . No clear magnetic lineation could be defined from this basalt.

AMS measurements of the Rafalovka basalt show that the rock is magnetically foliated with a foliation plane that trends 257°E and dips 19° . Here the mean direction of k_{max} indicates a flow direction at $c. D = 167^\circ$ and $I = 3^\circ$, which is different from that of the Ivanchi upper basalt. The fact that the basalts are magnetically foliated argues for at least the coarse-grained magnetic minerals to be primary.

The southern area

Three significantly different magnetization directions are identified in samples from the two quarries in the southern area (sites 3 and 4). The coercivity of remanence (H_c) is generally very low with an MDF < 5 mT. However, a stable high-coercivity remanence can be isolated in AF fields of 40 mT and higher (Fig. 4i and k). The intensity of this component of magnetization is of the order of 10^{-2} – 10^{-1} A m $^{-1}$, which is often not more than a few per cent of the NRM. The mean directions of this magnetization are well defined (the basalt at site 3 has not been studied by other teams) at $D = 115.8^\circ$, $I = -19.8^\circ$ and $D = 115.6^\circ$, $I = -20^\circ$; for site 3 and site 4, respectively (Table 2; Fig. 3).

In the low- H_c range, significantly different directions of magnetization are identified at both sites 3 and 4 ($D = 80.9^\circ$, $I = 47.5^\circ$; ≤ 10 mT; $D = 81.1^\circ$, $I = 47.7^\circ$, respectively). At site 3 the low- H_c component is also isolated in a low unblocking temperature (T_{ub}) range at temperatures between 200 and 400 $^\circ\text{C}$ ($D = 85.5^\circ$, $I = 42.4^\circ$; Table 2). The T_{ub} spectrum is wide (Fig. 4j); however, the highest gradient in the intensity curve is between 250 and 350 $^\circ\text{C}$ and $c. 80\%$ of the intensity is lost at temperatures below 400 $^\circ\text{C}$. During the thermal demagnetization there is a tendency for a gradual shallowing of inclination in the remanence directions with increasing temperature. This may suggest that this remanence direction is a result of imperfect cleaning and a mixture between the low- T_{ub} (and low- H_c) component and the high- H_c magnetization. However, great circle analyses of the thermal data reveal intersecting great circles at $D = 85.8^\circ$ and $I = 35.1^\circ$, not far from the calculated mean direction. At site 4 a magnetization similar to the low- H_c component is also isolated during thermal demagnetization in the temperature range 200–450 $^\circ\text{C}$ ($D = 95.3^\circ$, $I = 44.8^\circ$; Table 2; Fig. 4l). Generally the high- H_c component of magnetization is not isolated during thermal cleaning; however, in a few samples

the high- H_c component is traced at temperatures above 480 $^\circ\text{C}$ (Fig. 4l). At these temperatures the intensity of remanence is only a few per cent of that of the NRM. Great circle analysis of the thermal data shows an intersection point ($D = 100.7^\circ$, $I = 37.3^\circ$), which is similar to the calculated mean direction. The VGP's calculated from the low- H_c and low- T_{ub} components at sites 3 and 4 are located at 27.3°N , 103.6°E and 27.3°N , 103.2°E , respectively (Table 2), whereas the poles calculated from the high- H_c magnetizations plot at 24.1°N , 282.3°E and 24°N , 282.4°E , respectively.

AMS data show that the Berestovets basalt is magnetically lineated, with a direction of K_{max} at $D = 337^\circ$ and $I = 1^\circ$, which suggests an almost horizontal lava flow at this site.

Origin of the carriers of remanence

In samples from the northern area, sites 1 and 2, the remanent magnetization unblocks in a fairly wide blocking temperature range, but with a main gradient in the curve of demagnetization at temperatures above 530 $^\circ\text{C}$. Susceptibility measurements during continuous heating of the lower basalt at site 1 and the basalt at site 2 indicate that the magnetic minerals are predominantly titanomagnetite with Curie temperatures (T_c) of 510–550 $^\circ\text{C}$ at site 1 and 550–580 $^\circ\text{C}$ at site 2. The titanomagnetite is oxidized during heating and inverts to titanohematite, which is seen from the irreversible curve during cooling.

Corresponding measurements of the basalt at site 3, Berestovets, and site 4, Basaltovye, show that a large part (40–60%) of the magnetization is carried by titanomagnetite and/or titanomaghemite with T_c of $c. 320^\circ\text{C}$ and there are no indications of minerals with T_c higher than 580 $^\circ\text{C}$. The curves are irreversible and titanomagnetite–titanomaghemite is probably oxidized to titanohematite. Thermal demagnetizations of the NRM demonstrate unblocking temperatures where the main part is unblocked at temperatures below 400 $^\circ\text{C}$ and with the main decrease in intensity between 250 and 350 $^\circ\text{C}$. Only in some samples are there portions of the remanent magnetizations that unblock at temperature higher than 400–450 $^\circ\text{C}$.

In polished thin sections of samples from Basaltovoe (site 4) the titanomagnetite that is found is mainly of a single phase and without any decomposition or oxidation structures. Titanomaghemite of two different types is also identified, one related to late magmatic oxidation in the final stage of cooling of the rock and the other related to later alteration.

The composition of the remanence carriers at sites 3 and 4 has

been studied by micro X-ray analysis. This shows that the bulk content of ulvöspinel (x) is 0.5–0.6 and that the single-phase oxidation parameter (Z) does not exceed 0.2. This means that the remanence-carrying mineral is predominantly titanomagnetite, with subordinate titanomaghemite. From the Fe/Ti ratios, determined for 20 grains of titanomagnetite using a Camebax microprobe, the Curie temperature (T_c) for the titanomagnetite is calculated at c. 300°C (Dunlop & Özdemir 1997).

Thermal demagnetization of the samples from the basalts shows that c. 90% of the NRM unblocks in a wide unblocking temperature (T_{ub}) spectrum, but with the sharpest drop in intensities at c. 300°C, which is the similar to the Curie temperature estimated from the composition of the titanomagnetite and from the susceptibility–temperature measurements.

The occurrence of single-phase titanomagnetites with such low Curie points as seen in the Basaltovoe basalts may be explained by a reduction environment in magma formed at a depth of 40–60 km. This is supported by the existence of native iron in the basalts of Berestovets (Lazarenko *et al.* 1966).

Curie temperatures ranging from 320 to 450°C are characteristic of titanomaghemite (Readman & O'Reilly 1972). The existence of titanomaghemite is also confirmed by the Camebax microprobe study and by examination of polished sections. The increased Mn content and the sculptured surface (brain-like structure) of the grains indicate that some of the opaque minerals are titanomaghemite (Ramdor 1960; Prévot 1971). The titanomaghemite content varies in individual grains depending on the extent of oxidation. In Basaltovoe unmixing has occurred during lava cooling in an oxygen-poor setting, as shown by the presence of native iron.

Maghemitization occurred mainly at the grain surface or in fissures of the titanomagnetite. The interiors of large grains remained unoxidized.

The basalts at sites 3 and 4 also contain interstitial chlorophaeite and palagonite, which are the result of alteration of volcanic glass. Chemical analyses show high contents of Fe_2O_3 (Bialowolska *et al.* 2002), which are related to the alteration products and a higher degree of alteration in the site 3 basalt than in the basalt at site 4. Secondary minerals such as Fe-oxides are known to be related to such palagonitization (see, e.g. Fisher & Schmincke 1984) and here fine-grained magnetite is found in relation to the alteration products.

Discussion

Palaeomagnetic data

The magnetizations of the basalts in the northern area, sites 1 and 2, are most probably primary. This is demonstrated by the significantly different remanence directions isolated in the Lower and Upper Ivanchi basalt (site 1), the antipodal magnetizations of the green tuff separating the upper and lower basalts, the remagnetization of the tuff in contact with the upper basalt and the different polarity of the basalts at sites 1 and 2, which are only a few kilometres apart.

The two isolated directions of remanence for site 3 and 4 fall more or less on a great circle that includes the direction of the PEF. This makes us suspect that the low H_c and T_{ub} is a result of imperfect cleaning and separation of the high- H_c component from a PEF magnetization. However, the low- T_{ub} (and low- H_c) components are well defined, and great circle analyses confirm that the isolated directions are not the result of mixtures between a PEF magnetization and the high- H_c magnetization. In our interpretation, the low- T_{ub} (200–400°C) magnetization is carried

by titanomagnetite and titanomaghemite, where the titanomaghemite is the result of oxidation of titanomagnetite during the final cooling of the lava, and with directions of magnetization that reflect a primary magnetization of the basalt. The high- H_c component, which makes up less than 10% of the NRM, is sometimes also traced during thermal treatment at temperatures above 500°C and it is then stable up to c. 550°C. With reference to the coercivity, T_{ub} and the mineralogical studies, this magnetization is interpreted to be carried by the fine-grained titanium-poor magnetite dispersed in the alteration products of volcanic glass (chlorophaeite and palagonite). A low-temperature post-cooling alteration, responsible for the high- H_c secondary magnetization, is also reflected by the severe argon loss in samples from sites 3 and 4 and by the partial resetting of the radiometric system as demonstrated in the $^{40}\text{Ar}/^{39}\text{Ar}$ whole-rock age determinations. The partial resetting is reflected in a plateau age of 393 Ma at site 3, where the alteration is most developed.

The VGPs calculated from the high H_c and T_{ub} of the lower basalt at site 1 (A2 pole; Table 2, Fig. 5) and the basalt at site 2 (A1 pole) are located at a similar latitude, but separated by c. 35° longitude, and they do not fall on the APWP suggested by Smethurst *et al.* (1998; Vendian–Permian) or Torsvik *et al.* (1996; Upper Riphean–Vendian). The age of these poles is estimated at 580 Ma on basis of the $^{40}\text{Ar}/^{39}\text{Ar}$ plateau (580 ± 9 Ma) and isochron (579 ± 12 Ma) ages of our study. Stratigraphically the basalt at site 2 is considered older than that at site 1. The poles of the upper basalt and the baked tuff at site 1 (A3 and A3* poles; Table 2, Fig. 5) represent younger magnetizations and plot on the Silurian part of APWP. Our data, however, indicate a magnetization age for these poles of 561 ± 13 Ma. A palaeomagnetic study of the basalts at sites 2 and 4 has also been performed by Nawrocki *et al.* (2004a) and the calculated VGP for site 2 (Table 2) is very close to the corresponding pole (A1 pole) of our study and the results from Iosifidi *et al.* (2001a, b).

Although the basalt flows are thick (c. 10–20 m) the time of cooling is probably not long enough for the rock to acquire a magnetization that averages out secular variations. This means that the calculated poles at least to some extent may reflect secular variations in the geomagnetic field. As secular variations are probably not averaged out, a palaeomagnetic pole has to be calculated from the mean of the VGPs and for this we also include the results of Nawrocki *et al.* (2004a). A mean pole (pole 5, Table 2) can be calculated from the A1 (site 2, Rafalovka) and the A2 (site 1, Ivanchi lower basalt) poles and the pole calculated from the mean direction of the green tuff at site 1 (A4). These VGPs are of different polarities, which means that there are age differences significant enough to average out secular variation. However, as the pole from the green tuff (A4) is based on only three samples, this poorly defined palaeomagnetic pole (40.4°N, 5.6°E, $A_{95} = 28.6^\circ$; Table 3), representing an age of c. 580 Ma, should be used with caution. A mean pole (pole 6, Table 2) can also be calculated from the mean poles A1 (site 2), A2 (site 2 lower basalt), A4 (site 2 green tuff) adding the probably younger poles A3 (site 2 upper basalt) and A3* (site 2 red tuff), as the $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the basalts are similar and with overlapping error bars. Also, this palaeomagnetic pole is poorly defined (19.8°N, 4.4°E, $A_{95} = 28.3^\circ$) and represents an age of c. 580–561 Ma.

Neither the poles of the basalts of sites 3 and 4, as calculated from the high- H_c components (C poles, Table 2, Fig. 5), nor the poles of the low- T_{ub} components (B poles) fall on the Vendian–Permian APWP. For the B and C poles it is difficult to assign an age as there are two generations of magnetization in these

Fig. 4. Examples of remanence behaviour during AF and thermal treatment of samples from the various sites. In the vector plots the declination is marked by crosses and the inclination by filled diamonds.

basalts. The $^{40}\text{Ar}/^{39}\text{Ar}$ data, chemical analyses and mineralogical studies suggest that the alteration is selective. We interpret the B magnetizations to have survived alteration, whereas the C magnetization is considered to be a result of alteration.

Based on the stratigraphy, the basalts at sites 3 and 4 are considered to be younger than those at sites 1 and 2. The site 4 flows are lithologically correlated with basalts of the upper Sawatycze series in Poland. Tuffs overlying the Sawatycze basalts

have been radiometrically dated at 551 ± 4 Ma (U–Pb, zircon; Compston *et al.* 1995), compared with the $^{40}\text{Ar}/^{39}\text{Ar}$ age from this study (plateau age 582 ± 7 Ma) and a K/Ar age of 545.5 Ma (Savchenko *et al.* 1984) for the site 4 basalt. At present, we do not know why the $^{40}\text{Ar}/^{39}\text{Ar}$ age is significantly older than these other ages. However, in our interpretation the B poles represent primary magnetizations and the age of these poles is therefore suggested to be 580–545 Ma, whereas the age of the C pole is

Fig. 4. (continued)

lower and is related to the alteration of the volcanic glass, probably by diagenesis burial metamorphism. There is a $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 393 Ma for the basalt; however, the C poles are still different from the well-defined Devonian part of the APWP.

From the VGPs of sites 3 and 4, palaeomagnetic poles are calculated from the mean of B1 and B2, and C1 and C2, respectively (Table 3). Also for these poles, the data from

Nawrocki *et al.* (2004a; B2 and C2; Table 2) are used for calculation of a mean pole position. As the mean of the B1 and B2 poles is calculated from only two site means with magnetizations of the same polarity it is still uncertain if secular variation has been fully averaged out and the mean pole is therefore tentative. Although the C pole is also calculated from two site means (Table 3), the magnetization is of chemical origin and secular variation is most probably averaged out.

Pre-Ordovician palaeomagnetic data are limited and even if we select only those that fulfil at least three of the criteria in the Van der Voo grading system (Van der Voo 1990), such as those of Torsvik *et al.* (1996) and Meert & Torsvik (2003), some have still been questioned (e.g. Popov *et al.* 2002). A crucial problem is to assign a correct age of the magnetization. This is expressed by the large dispersion of poles of coeval age. The assigned magnetization age of the Fen Complex (583 ± 15 Ma; Meert *et al.* 1998) is very close to that of the basalts at sites 1 and 2 (580 ± 9 Ma and 561 ± 13 Ma) and not very different from the magnetization age (555 Ma) assigned to the White Sea sedimentary rocks of Popov *et al.* (2002). However, the palaeomagnetic pole positions are very different (Fig. 5), and are separated by more than 160° longitude and 70° latitude.

The pole calculated from the White Sea sediments (Popov *et al.* 2002) passes reversal and stratigraphic field tests and is interpreted to reflect a magnetization age of *c.* 555 Ma (Table 3). The magnetization is considered to have been acquired during diagenetic processes, at a time not very different from that of the radiometric age (U–Pb, of a single zircon grain). However, the time for acquisition of the magnetization carried by hematite may not necessarily be related to the U–Pb age. Diagenetic processes in sedimentary rocks can be slow and the mixing of red and grey and greenish grey sediments indicates a reduction environment, which retards the oxidation of the primary magnetic minerals. In stable cratonic environments diagenetic processes within sedimentary rocks may continue for at least 50–100 Ma after deposition (Piper 1987). Iosifidi & Khramov (2005) reported palaeomagnetic results from Vendian, *c.* 545–570 Ma old, sedimentary rocks in Podolia, Ukraine. The magnetization of these rocks is complex and seven components were identified. From the calculated pole positions, five of them were suggested to be

Fig. 5. Virtual geomagnetic poles and mean pole positions shown with 95% confidence circle. A, B and C denote poles of this study (Table 2), and Z denotes the pole of component Z of Popov *et al.* (2002). Poles 5 and 6 are mean poles calculated from poles A1, A2 and A4, and poles A1, A2, A3, A3* and A4, respectively (Table 2). The other poles are selected poles from other studies: 1, Andarum limestone; 2, Dividal Group; 3, Fen complex (see Table 3 for references). The bold continuous line marks the Neoproterozoic–Palaeozoic APWP of Smethurst *et al.* (1998). The position of the reversed pole of the Fen complex (pole 3') close to the Permian part of the path should be noted.

Table 3. Baltic and Laurentian reference poles for the Late Proterozoic–Ediacaran

Rock	P_{lat}	P_{long}	α_{95}	P_{lat}^*	P_{long}^*	A_{95}	Age (Ma)	Pole	Reference
Baltica									
Andarum Limestone	52	111				8.3	500	1	Torsvik & Rehnström 2001
Dividal Group	58.4	122.5				10.2	535		Rehnström & Torsvik 2003
Winter Coast Z component	25.3	312.2				2.9	555	Z	Popov <i>et al.</i> 2002
Fen tinguates	50.1	143.9				6.2	583		Piper 1981
Fen Central Complex	56	150				8.3	583	3	Meert <i>et al.</i> 1998
Egersund Dyke	22	51				14	616, 650	4	Poorter 1972
Late Proterozoic mean	−28.1	17.3				7.6	750	8	Torsvik <i>et al.</i> 1996
Laurentia									
Late Cambrian mean	−3	344	−4	22.1	12	495	9	Meert 1999	
Tapeats sandstone	−5	338	−5.7	16.1	3	508	10	Elston & Bressler 1977	
Callander Complex	46	301	44.2	338.2	6	575		Symons & Chiasson 1991	
Catoctin A	42	297	40.8	334.6	9	564		Meert <i>et al.</i> 1994a, b	
Sept Iles Intrusion B	44	315	43	352.2	5	546		Tanczyk <i>et al.</i> 1987	
Mean 575–564 Ma			43.4	341.9	10.3		11		
Brock Inlier Sills	2	345	1.5	22.9	16	723		Park 1981a, b	
Franklin Dykes	−9	333	−9.8	11.2	5	723		Christi & Fahrig 1983	
Top Little Dal	−24	339	−25	17.7	11	778		Morris & Aitken 1982	
Tsezotene sills	2	338	1	15.9	5	778		Park 1981a, b	
Mean 778–723 Ma			−8	16.9	15.2		12		

The Laurentian poles (for references see Meert & Torsvik 2003) follow an earlier evaluation of Meert & Torsvik (2003). The mean poles are calculated here. All poles are referred to as south poles; P_{lat}^* , P_{long}^* , poles that are rotated into the reference frame of Baltica using the fit of Bullard *et al.* (1965).

the result of Palaeozoic remagnetization, whereas a high- T_{ub} component (P2) recorded in all rock units was considered to reflect a Vendian magnetization. However, because of the complex nature of the remanent magnetization, it is difficult to assign a correct age for this magnetization. The corresponding pole position of this component (40°N, 276.5°E) falls close to the Z pole (24°N, 312°E) of Popov *et al.* (2002) and to the C pole of our study pole (23°N, 286°E; Table 3), which we consider to be secondary of late Ediacaran age, or possibly as young as Devonian age. Piper (1981) made a palaeomagnetic study of the Alnö Complex, for which he assigned an age of 553–537 Ma. The calculated pole in that study (8°N, 272°E) does not fall far from the C poles either. However, no vector analyses were made for the identification of the characteristic magnetizations and the site mean directions are scattered. New $^{40}\text{Ar}/^{39}\text{Ar}$ data also suggest a different age (589 Ma) for these rocks (T. H. Torsvik, pers. comm.).

In the selection of data made by Meert & Torsvik (2003) only five poles in the time span 750–500 Ma were considered reliable: a mean 750 Ma pole, Egersund dykes (Poorter 1972; T. H. Torsvik *et al.* unpubl. data), the Fen Complex (Meert *et al.* 1998), the Torneträsk Formation (or Dividal Group; Torsvik & Rehnström 2001; Rehnström & Torsvik 2003) and the Andarum Limestone (Torsvik & Rehnström 2001; Table 3, Fig. 6).

The 750 Ma pole was calculated from the mean of Kildinskaya Formation, the Sredny–Kildin Red Beds and the Vadsö Group (Torsvik *et al.* 1996, and references therein). The 616 Ma

Egersund pole of Poorter (1972; see also Storetvedt 1966; Bingen *et al.* 1998) plots on a younger part of the APWP for Baltica (Smethurst *et al.* 1998), indicating the possibility of an Early Ordovician remagnetization. However, from unpublished data (J. G. Meert & T. H. Torsvik) a new pole has been calculated for the Egersund dykes, with a position (48°N, 20°E; Fig. 6) that is significantly different from younger poles and is similar to the 580 Ma pole of our study (41°N, 6°E). The dykes carrying this magnetization intrude Late Proterozoic anorthosite from which significantly different directions of magnetization have been isolated (Hargraves & Fish 1972), which means that the dyke magnetization is probably original.

The rocks within the Fen Complex are generally altered; however, the palaeomagnetic data (Meert *et al.* 1998) were obtained from satellite dykes that are suggested to have escaped alteration. Biotite and phlogopite separates from two dykes were dated using the $^{40}\text{Ar}/^{39}\text{Ar}$ step heating technique at 583 ± 15 Ma. This age is supported by similar ages obtained from K–Ar, Rb–Sr, U–Pb and Pb–Pb ages in previous studies (see Meert *et al.* 1998). The minor alteration of the dykes and the consistent age determinations made Meert *et al.* (1998) argue for the magnetization to be original and of c. 583 Ma age. However, no palaeomagnetic field tests have been presented to demonstrate a primary magnetization and the calculated pole is close to the Permian pole of Baltica. Because of the presence of a large Permian plutonic massif, located only c. 12 km west of the study area, and a large number of Permian dykes in the Fen Province, the palaeomagnetic pole (56°N, 150°E) calculated from the Fen Complex should be used with caution, as was also pointed out by Meert *et al.* (1998).

The Fen pole is located not far from the pole of the Dividal Group (58°N, 123°E; Rehnström & Torsvik 2003; Fig. 5). That pole was calculated from geographically well-separated sites of siltstone in northern Sweden and Norway. The assigned age 535 Ma is a stratigraphic age based on trace fossils. Also, the late Cambrian (c. 500 Ma) pole of the Andarum Limestone (Torsvik & Rehnström 2001) plots close to the pole of the Dividal Group (Torneträsk formation). This pole is based on only one site and was reported by Torsvik & Rehnström (2001) as preliminary.

As the C poles of our study probably are the result of remagnetization (as the Z component of Popov *et al.* (2002) may also be) and the age of magnetization is unclear, this reduces the number of poles for the Late Proterozoic–Ediacaran to the Late Proterozoic mean pole (750 Ma; Torsvik *et al.* 1996), the Egersund pole, the A pole and tentatively the B pole of our study, and the poles of the Fen complex, the Dividal Group and the Andarum Limestone. The contradictory results for the A poles of this study and the Fen Complex are puzzling. The age for the Fen Complex based on mineral ages of various radiometric systems seems very reliable, whereas the origin of the magnetization is more difficult to define. The A poles of our study are poorly defined and the ages are whole-rock ages. However, the A poles are of different polarity and significantly different, which argues for original magnetizations. Another argument for the A poles being primary is the similarity in pole position to that of the Egersund dykes, which have an assigned age of magnetization (616 Ma) similar to that of the A poles. We propose that the A poles represent a magnetization of c. 580 Ma and 580–561 Ma, respectively, and the Fen pole is possibly a Permian overprint.

The mean B pole is, on stratigraphic and radiometric grounds, considered younger than the A poles and the assigned age (582–545 Ma) is not far from the age estimated for the Dividal Group

Fig. 6. Late Proterozoic palaeomagnetic poles for Baltica and Laurentia rotated into the reference frame of Baltica based on the Bullard *et al.* (1965) fit (rotation pole is at 88°N, 27°E, rotation angle is 38°, Table 3). Poles 1–4 denote reliable poles for Baltica as selected by Meert & Torsvik (2003): 1, Andarum limestone; 2, Dividal Group; 4, Egersund dykes. Poles 5, 6 and 7 are poles calculated in this study (Table 2); pole 8 is the Late Proterozoic mean pole (Table 3). The Fen pole (3; Table 3) is regarded here as of Permian age and therefore not included. The rotated Laurentian poles (9–12, Table 3) are shown with dashed 95% confidence circles and the magnetization ages are shown in bold type. The bold continuous line marks the Neoproterozoic–Palaeozoic path of Smethurst *et al.* (1998). The similar Ediacaran pole positions for the two continents should be noted.

(535 Ma). The pole positions are also similar, which may argue for a primary origin of this pole.

If these groups of poles are based on original magnetizations and correct ages it means there have been a rapid changes in pole positions from *c.* 616 to 535 Ma, indicating a very high drift rate for Baltica during the Ediacaran or that there may have been a true polar wander (TPW). However, a TPW origin for the high rate is not supported by Late Precambrian palaeomagnetic data from Australia, from which no significant TPW is indicated (e.g. Wingate & Giddings 2000; Pisarevsky *et al.* 2001).

Tectonic reconstructions

Calculating the positions and orientation of Baltica based on these selected Late Proterozoic–Ediacaran poles (south poles; Fig. 6), Baltica drifted from an equatorial position at 750 Ma (Fig. 7) to a high southern latitude (*c.* 75°S) at 616 Ma. Baltica kept a high latitudinal position until *c.* 580 Ma (*c.* 70°S) and 580–561 Ma (*c.* 55°S) and then moved to a lower latitude (*c.* 40°S) at 580–545 Ma. From 580–561 to 580–545 Ma Baltica rotated some 180° clockwise, a rotation that continued to 535 Ma, and then drifted at somewhat higher latitudes until *c.* 500 Ma. During the Ediacaran, Baltica in this scenario drifted with a high latitudinal drift rate of up to *c.* 13 cm a⁻¹ and rotated at least 5° Ma⁻¹, depending on the age assigned to the poles.

For the reconstruction of the Rodinia break-up, this scenario for Baltica can be compared with that of Laurentia. The reference poles for Laurentia follow an earlier evaluation presented by Meert & Torsvik (2003). All palaeopoles (Table 3) are referred to as south poles and they are graded $Q \geq 3$ in the Van der Voo (1990) grading system. For this comparison, the North American poles have been rotated into the reference frame of Baltica using the Bullard *et al.* (1965) fit with a rotation pole at 87°N, 27°E and a rotation angle of 38°. The 750 Ma and 580–560 Ma poles in this fit (Fig. 6) are similar for Baltica and Laurentia, whereas the *c.* 500 Ma poles are very different. The similarity in Ediacaran pole positions suggests that Baltica and Laurentia drifted together from *c.* 750 Ma to 580–560 Ma (Fig. 7). Both continents occupied positions close to the equator at 750 Ma, and they moved to high southern latitudes at 580–560 Ma. In our reconstruction Baltica was still joined to Laurentia after the initial break-up of Rodinia and until 580–560 Ma in a similar relative position as for 750 Ma. This is contrary to what has been suggested in earlier reconstructions (e.g. Torsvik *et al.* 1996; Meert & Torsvik 2003; Cocks & Torsvik 2005), in which Baltica at 580 Ma was attached to Laurentia, but at an orientation *c.* 180° different from the

reconstruction presented here. A close link between Baltica and eastern Laurentia during the late Precambrian is supported by a close relationship between Baltica and eastern Laurentia trilobites (Meert & Lieberman 2004). The volcanic trap formation in Ukraine marks the initial rifting between southwestern Baltica and an unknown continent. The rifting and opening of the Iapetus ocean is suggested to have started in the south, as indicated by Vendian dykes younging northward along the eastern margin of Laurentia, and resulted in a northward opening by 570 Ma (Cawood *et al.* 2001). In our interpretation the final opening is reflected by the large clockwise rotation of Baltica between 580–561 Ma and *c.* 550 Ma. This reconstruction should, however, be regarded with caution, as it is partly based on tentative and poorly defined poles.

Tillites have been recognized in drill cores from southern Volhyn (Makhnach & Veretennikov 1976) and they are overlain by sediments of early Ediacaran age. From studies of tillites in Belarus, western Ukraine and western and central Russia, Veretennikov (1998) suggested that a global glaciation occurred at *c.* 600 Ma. Neoproterozoic near-equatorial glaciogenic deposits have been found (e.g. Hoffman *et al.* 1998; Sohl *et al.* 1999; Evans 2003; Kilner *et al.* 2005) with contiguous evaporates, redbeds and warm-water carbonates. This palaeoclimatic paradox has been explained by a high obliquity of the ecliptic of the Earth (Williams 1975, 1993), which may result in low-latitude glaciation. Others have suggested a global glaciation to explain the cold climate at low latitudes (e.g. Tarling 1974), and a hypothesis of one or more generations of ‘Snowball Earth’ occurring between 800 and 600 Ma has been presented (Kirschvink 1992; Hoffman *et al.* 1998). The reconstructions here suggests that Baltica from *c.* 616 to 580–561 Ma occupied high-latitude positions, contrary to the model of a high-obliquity Earth, in which a low-latitude glaciation would have been expected. The high-latitude position of Baltica means that a ‘Snowball Earth’ is not required to explain the glaciation; however, given the indications of low-latitude Late Precambrian glaciations elsewhere, our results are consistent with a ‘Snowball Earth’.

Conclusion

Calculating the position and orientation of Baltica based on the new poles of this study, Baltica occupied high southern latitude positions at 580–561 Ma. In our tentative reconstruction, Baltica, after the initial break-up of Rodinia, may still have been joined with Laurentia in a relative position similar to that at 750 Ma. This contradicts earlier palaeogeographical reconstructions in

Fig. 7. Proposed reconstructions of Baltica and Laurentia from 750 to 500 Ma as calculated from selected palaeomagnetic poles and poles of this study (Tables 2 and 3) and for the Gondwana megacontinent at *c.* 550 Ma (Torsvik *et al.* 1996). The rapid latitudinal drift and rotation rates for the period of this study should be noted. (See text for discussion.)

which Baltica at c. 580 Ma had an orientation that was 180° different. In our scenario the volcanic trap formation in Ukraine marks the initial rifting between southwestern Baltica and an adjacent continent. The northward opening of the Iapetus ocean was completed with a large clockwise rotation of Baltica relative to Laurentia at c. 550 Ma. The reconstruction also suggests that Baltica during the Ediacaran glaciation occupied latitudinal positions as high as 60–70°S, which contradicts the high-obliquity Earth model, but is consistent with a 'Snowball Earth' model to explain low-latitude Neoproterozoic glaciations.

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References

- BIALOWOLSKA, A., BAKUN-CZUBAROW, N. & FEDORYSHYN, Y. 2002. Neoproterozoic flood basalts of the beds of the Volhynian Series (East European Craton). *Geological Quarterly*, **46**, 37–53–57.
- BINGEN, B., DEMAÏFFE, D. & VAN BREMEN, O. 1998. The 616 Ma old Egersund Dyke Swarm, SW Norway, and the Late Neoproterozoic opening of the Iapetus Ocean. *Journal of Geology*, **106**, 565–574.
- BULLARD, E.C., EVERETT, J.E. & SMITH, A.G. 1965. The fit of the continents around the Atlantic. *Philosophical Transactions of the Royal Society of London, Series A*, **258**, 41–51.
- CAWOOD, P.A., MCCAUSLAND, P.J.A. & DUNNING, G.R. 2001. Opening Iapetus: constraints from the Laurentian margin in Newfoundland. *Geological Society of America Bulletin*, **113**, 443–453.
- CHRISTI, K.W. & FAHRIG, W.F. 1983. Paleomagnetism of the Borden dykes of the Baffin island and its bearing on the Grenville Loop. *Canadian Journal of Earth Science*, **20**, 275–289.
- COCKS, L.R. & TORSVIK, T.H. 2005. Baltica from the late Precambrian to mid-Palaeozoic times: the gain and loss of a terrane's identity. *Earth-Science Reviews*, **72**, 39–66.
- COMPSTON, W., SAMBRIDGE, M.S., REINEFRANK, R.F., MOCZYDLOWSKA, M., VIDAL, G. & CLAESSON, S. 1995. Numerical ages, volcanic rocks and the earliest faunal zone within the Late Precambrian of east Poland. *Journal of the Geological Society, London*, **152**, 599–611.
- DUNLOP, D.J. & ÖZDEMİR, Ö. 1997. *Rock Magnetism*. Cambridge University Press, Cambridge.
- ELSTON, D.P. & BRESSLER, S.L. 1977. Paleomagnetic poles and the polarity zonation from Cambrian and Devonian strata of Arizona. *Earth and Planetary Science Letters*, **36**, 423–433.
- EVANS, D.A.D. 2003. A fundamental Precambrian–Phanerozoic shift in the Earth's glacial style? *Tectonophysics*, **375**, 353–385.
- FISHER, R.A. 1953. Dispersion on a sphere. *Proceedings of the Royal Society of London, Series A*, **217**, 295–305.
- FISHER, R.V. & SCHMINCKE, H.-U. 1984. *Pyroclastic Rocks*. Springer, Berlin.
- GLEVASSKAYA, A.M., MIKHAILOVA, N.P. & KRAVCHENKO, S.N. 1997. Vendian geomagnetic poles of East-European Platform (EEP) based on paleomagnetic studies of west Ukrainian sections. In: BOSTRÖM, R., CAROZZI, I., ARLEFJÄRD, I., WAHLBERG, A.-S. & DOVNER, P.O.J. (eds) *IGA Abstract. Books, Section 1, 11*, Uppsala. 59.
- GLEVASSKAYA, A.M., MIKHAILOVA, N.P. & KRAVCHENKO, S.N. 2001. Paleomagnetism of the Volyn and Mogilev–Podolian Vendian Series from the southwestern part of the East-European Platform. *Geophysical Journal*, **20**, 149–175.
- HARGRAVES, R.B. & FISH, J.R. 1972. *Paleomagnetism of anorthosite in southern Norway and comparison with an equivalent in Quebec*. Geological Society of America, Memoirs, **132**, 141–148.
- HOFFMAN, P.F., KAUFMAN, A.J., HALVERSON, G.P. & SCHRAG, D.P. 1998. A Neoproterozoic snowball Earth. *Science*, **281**, 1342–1346.
- IOSIFIDI, A.G., BACHTADSE, V., KHRAMOV, A. & KUZNETSOVA, A. 2001a. Paleomagnetic data for Vendian basalts from Ukraine. In: TROYAN, V.N. ET AL. (eds) *3rd International Conference on Problems of Geocosmos, Abstracts Volume*, St. Petersburg. 74–75.
- IOSIFIDI, A.G., BACHTADSE, V., TAIAT, J., KHRAMOV, A.N. & KUZNETSOVA, A.V. 2001b. Paleomagnetic data from Vendian rocks from the Ukraine. XXVI EGS General Assembly, Nice 2000. *Geophysical Research Abstracts*, **3**, 1206.
- IOSIFIDI, A.G. & KHRAMOV, A.N. 2005. Multicomponent magnetization of Vendian sedimentary rocks in Podolia, Ukraine. *Russian Journal of Earth Sciences*, **7**, 1–14.
- KILNER, B., MAC NIOCAILL, C. & BRAISE, M. 2005. Low-latitude glaciation in the Neoproterozoic of Oman. *Geology*, **33**, 413–416.
- KIRSCHVINK, J.L. 1980. The least-square line and plane and the analysis of paleomagnetic data. *Geophysical Journal of the Royal Astronomical Society*, **62**, 699–718.
- KIRSCHVINK, J.L. 1992. Late Proterozoic low latitude glaciation: the snowball Earth. In: SCHOFF, J.W. & KLEIN, C. (eds) *The Proterozoic Biosphere: a Multidisciplinary Study*. Cambridge University Press, Cambridge, 511–512.
- KNIGHT, M.D. & WALKER, G.P.L. 1988. Magma flow directions in dikes of the Koolau Complex, Oahu, determined from magnetic fabric studies. *Journal of Geophysical Research*, **93**, 4301–4319.
- LANPHERE, M.A. & DALRYMPLE, G.B. 2000. *First-principles calibration of ³⁸Ar tracers: implications for the ages of ⁴⁰Ar/³⁹Ar fluence monitors*. US Geological Survey, Professional Papers, **1621**.
- LAYER, P.W. 2000. ⁴⁰Ar/³⁹Ar age of the El'gygytyn impact event, Chukotka, Russia. *Meteoritics and Planetary Science*, **35**, 591–600.
- LAYER, P.W., HALL, C.M. & YORK, D. 1987. The derivation of ⁴⁰Ar/³⁹Ar age spectra of single grains of hornblende and biotite by laser step heating. *Geophysical Research Letters*, **14**, 757–760.
- LAZARENKO, E.K., MATKIVSKY, O.I., VINAR, O.M., SHASKINA, V.P. & GNATIV, O.M. 1966. *Mineralogy of igneous complexes of Western Volynian*. Lviv University Press, Lviv (in Ukrainian).
- LLANOS, M.P.I., TAIT, J.A., POPOV, V. & ABALMASSOVA, A. 2005. Palaeomagnetic data from Ediacaran (Vendian) sediments of the Arkhangelsk region, NW Russia: an alternative apparent polar wander path of Baltica for the Late Proterozoic–Early Palaeozoic. *Earth and Planetary Science Letters*, **240**, 732–747.
- MAKNACH, A.S. & VERETENNIKOV, P.E. 1976. *The Riphean and Vendian of Byelorussia*. Nauka i Technika, Minsk.
- MCDUGALL, I. & HARRISON, T.M. 1999. *Geochronology and Thermochronology by the ⁴⁰Ar/³⁹Ar Method*, 2nd. Oxford University Press, New York.
- MEERT, J.G. 1999. A paleomagnetic analysis of Cambrian true polar wander. *Earth and Planetary Science Letters*, **168**, 131–144.
- MEERT, J.G. & LIEBERMAN, B.S. 2004. A paleomagnetic and palaeobiographical perspective on latest Neoproterozoic and early Cambrian tectonic events. *Journal of the Geological Society, London*, **161**, 477–487.
- MEERT, J.G. & TORSVIK, T.H. 2003. The making and unmaking of a supercontinent: Rodinia revisited. *Tectonophysics*, **375**, 261–288.
- MEERT, J.G., TORSVIK, T.H., EIDE, E.A. & DAHLGREN, S. 1998. Tectonic significance of the Fen Province, S. Norway: constraints from geochronology and paleomagnetism. *Journal of Geology*, **106**, 553–564.
- MEERT, J.G., VAN DER VOO, R. & PAYNE, T. 1994a. Paleomagnetism of the Catocin volcanic province: a new Vendian–Cambrian apparent polar wander path for North America. *Journal of Geophysical Research*, **99**, B3, 4625–4621.
- MEERT, J.G., HARGRAVES, R.B., VAN DER VOO, R., HALL, C.M. & HALLIDAY, A.N. 1994b. Paleomagnetic and ⁴⁰Ar/³⁹Ar studies of late Kibaran intrusives in Burundi, East Africa: implications for late Proterozoic supercontinents. *Journal of Geology*, **102**, 621–637.
- MORRIS, W.A. & AITKEN, J.D. 1982. Paleomagnetism of the Little Dal Lavas, McKenzie Mountains, Northwest Territories, Canada. *Canadian Journal of Earth Science*, **19**, 2020–2027.
- NAWROCKI, J., BOGUCKI, A. & KATINAS, V. 2004a. New late Vendian palaeogeography of Baltica and the TESZ. *Geological Quarterly*, **48**, 309–316.
- NAWROCKI, J., ZYLINSKA, A., BULA, Z., GRABOWSKI, J., KRZYWIEC, P. & POPRAWA, P. 2004b. Early Cambrian location and affinities of the Brunovistulian terrane (Central Europe) in the light of palaeomagnetic data. *Journal of the Geological Society, London*, **161**, 513–522.
- PARK, K.K. 1981a. Paleomagnetism of basic intrusions from the Brock inlier, Northwest Territories, Canada. *Canadian Journal of Earth Science*, **18**, 1637–1641.
- PARK, J.K. 1981b. Analysis of multicomponent magnetization of the little Dal Group, McKenzie Mountains, Northwest Territories, Canada. *Journal of Geophysical Research*, **86**, 5134–5146.
- PESONEN, L.J., ELMING, S.-Å. & MERTANEN, S. ET AL. 2003. Palaeomagnetic configuration of continents during the Proterozoic. *Tectonophysics*, **375**, 289–324.
- PIPER, J.D.A. 1981. Magnetic properties of the Alnö complex. *Geologiska Föreningens i Stockholm Förhandlingar*, **103**, 9–15.
- PIPER, J.D.A. 1987. *Palaeomagnetism and the Continental Crust*. Open University Press, Milton Keynes.
- PISAREVSKY, S.A., LI, Z.X., GREY, K. & STEVENS, M.K. 2001. A palaeomagnetic study of Empress 1A, a stratigraphic drillhole in the Officer basin: evidence for a low-latitude position of Australia in the Neoproterozoic. *Precambrian Research*, **110**, 93–108.
- POORTER, R.P.E. 1972. Preliminary palaeomagnetic results from the Fen Carbonate complex, S. Norway. *Earth and Planetary Science Letters*, **17**, 194–198.

- POPOV, V., IOSIFIDI, A. & KHRAMOV, A. 2002. Paleomagnetism of upper Vendian sediments from the Winter Coast, White Sea Region, Russia: implications for the paleogeography of Baltica during Neoproterozoic times. *Journal of Geophysical Research*, **107**, 2315–2322.
- PRÉVOT, M. 1971. A method for identifying naturally occurring titanomagnetites. *Journal of Geophysics*, **37**, 339–348.
- RAMDOR, P. 1960. *Die Erzminerale und Verwachsungen*. Akademie, Berlin.
- READMAN, P.V. & O'REILLY, W. 1972. Magnetic property of oxidized titanomagnetite. *Journal of Geomagnetism and Geoelectricity*, **24**, 69–90.
- REHNSTRÖM, E.F. & TORSVIK, T.H. 2003. Cambrian sediments and Proterozoic granites in the Dividalen–Torneträsk area, northern Scandinavia: palaeomagnetism and U–Pb geochronology. *Geologiska Föreningens i Stockholm Förhandlingar*, **125**, 131–138.
- SAMSON, S.D. & ALEXANDER, E.C. 1987. Calibration of the interlaboratory ^{40}Ar – ^{39}Ar dating standard, MMhb1. *Chemical Geology*, **66**, 2734.
- SAVCHENKO, N.A., BERNADSKAYA, L.G. & DOLGOVA, V.I. ET AL. 1984. *Palaeo-volcanism of the Ukraine*. Naukova Dumka, Kiev (in Russian).
- SHUMLYANSKY, L. & ANDRÉASSON, P.G. 2004. New geochemical and geochronological data from the Volyn Flood Basalt in Ukraine and correlation with large igneous events in Baltoscandia. *Geologiska Föreningens i Stockholm Förhandlingar*, **126**, 85–86.
- SMETHURST, M.A., KHRAMOV, A.N. & TORSVIK, T.H. 1998. The Neoproterozoic and Palaeozoic palaeomagnetic data for the Siberian Platform: from Rodinia to Pangea. *Earth-Science Reviews*, **43**, 1–24.
- SOHL, L.E., CHRISTIE-BLICK, N. & KENT, D.V. 1999. Paleomagnetic polarity reversals in Marinoan (ca. 600 Ma) glacial deposits of Australia: implications for the duration of low-latitude glaciation in Neoproterozoic time. *Geological Society of America Bulletin*, **111**, 1120–1139.
- SOKOLOV, V.B. & FEDONKIN, M.A. (EDS) 1990. *The Vendian System*. Springer, Berlin, 2.
- STEIGER, R.H. & JÄGER, E. 1977. Subcommittee on geochronology: convention on the use of decay constants in geo and cosmochemistry. *Earth and Planetary Science Letters*, **36**, 359–362.
- STORETVEDT, K.M. 1966. Remanent magnetization of some dolerite intrusions in the Egersund area, southern Norway. *Geofysiske Publikasjoner*, **26**, 1–17.
- SYMONS, D.T.A. & CHIASSON, A.D. 1991. Paleomagnetism of the Callander Complex and the Cambrian apparent polar wander path for North America. *Canadian Journal of Earth Science*, **24**, 355–363.
- TANCZYK, E.I., LAPOINTE, P., MORRIS, W.A. & SCHMIDT, P.W. 1987. A paleomagnetic study of the layered mafic intrusions at Sept-Îles, Quebec. *Canadian Journal of Earth Science*, **24**, 1431–1438.
- TARLING, D.H. 1974. A palaeomagnetic study of Eocambrian tillites in Scotland. *Journal of the Geological Society, London*, **130**, 163–177.
- TAUXE, L., GEE, J.S. & STAUDIGEL, H. 1998. Flow directions in dikes from anisotropy of magnetic susceptibility data: the bootstrap way. *Journal of Geophysical Research*, **103**, 17775–17790.
- TORSVIK, T.H. & REHNSTRÖM, E.F. 2001. Cambrian palaeomagnetic data from Baltica: implications for true polar wander and Cambrian palaeogeography. *Journal of the Geological Society, London*, **158**, 321–329.
- TORSVIK, T.H., SMETHURST, M.A. & MEERT, J.G. ET AL. 1996. Continental break-up and collision in the Neoproterozoic and Palaeozoic: a tale of Baltica and Laurentia. *Earth-Science Reviews*, **40**, 229–258.
- VAN DER VOO, R. 1990. The reliability of paleomagnetic data. *Tectonophysics*, **184**, 1–9.
- VELIKANOV, V.A. & KORENCHUK, L.V. 1997. Phases of magmatism and their relationship to sedimentation in the Late Precambrian (Riphean–Vendian) in the Volyn–Podolian region. *Geol. J.*, **1–2**, 124–130 [in Russian].
- VERETENNIKOV, N.V. 1998. Report at the International Symposium 'Paleoclimates and Evolution of Paleogeographical Environments in the Geological History of the Earth', 27–31 August 1998, Petrozavodsk.
- VOLOVNIK, B.YA., 1971. *Petrology and mineralogy of the Volyn–Podolian continental flood basalts*. PhD thesis, University of Lvov [in Russian].
- WILLIAMS, G.E. 1975. Late Precambrian glacial climate and the Earth's obliquity. *Geological Magazine*, **112**, 441–544.
- WILLIAMS, G.E. 1993. History of the Earth's obliquity. *Earth-Science Reviews*, **34**, 1–45.
- WINGATE, M.T.D. & GIDDINGS, J.W. 2000. Age and palaeomagnetism of the Mundine Well dyke swarm, western Australia: implications for an Australia–Laurentia connection at 755 Ma. *Precambrian Research*, **100**, 335–357.
- YORK, D., HALL, C.M., YANASE, Y., HANES, J.A. & KENYON, W.J. 1981. $^{40}\text{Ar}/^{39}\text{Ar}$ dating of terrestrial minerals with a continuous laser. *Geophysical Research Letters*, **8**, 1136–1138.

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