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Palaeomagnetism of Proterozoic rocks from the Ukrainian Shield: new tectonic reconstructions of the Ukrainian and Fennoscandian shields

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Abstract

A palaeomagnetic study has been performed on Palaeo-Mesoproterozoic rocks from three crustal blocks of the Ukrainian Shield, southern Sarmatia. Primary remanent magnetizations have been isolated in 2.0 Ga monzonite, 2.0–1.8 Ga sandstone, 1.77–1.72 Ga anorthosite and from mafic dykes of probably Palaeo-Mesoproterozoic ages. On basis of these results a sequence of 2.0–1.72 Ga apparent polar wander has for the first time been defined for the Ukrainian Shield. Palaeomagnetic and geological data indicate that there has probably not been any large scale tectonic movements within Sarmatia since the Mesoproterozoic. This suggests that tectonic reconstructions for the Ukrainian Shield may also include Sarmatia. The calculated pole positions for the Ukrainian Shield are significantly different from poles of similar age from the Fennoscandian Shield. The tectonic reconstructions demonstrate that the relative position and orientation of the Ukrainian Shield as a part of Sarmatia in the time interval 2.0–1.78 Ga was different from its present position relative to Fennoscandia. One pole from the Ukrainian Shield falls on the ca. 1.6 or 1.3 Ga part of the Fennoscandian APWP. This pole may represent a time when Fennoscandia was already accreted to Ukrainia. Contemporaneous rifting of the two cratons at ca. 1.35 Ga indicates that they were already joined to each other at that time, which means that the final accretion should have taken place sometimes after ca. 1.8 Ga ago. © 2001 Elsevier Science B.V. All rights reserved.

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1. Introduction

The East European Craton (EEC) can be divided into three crustal segments with autonomous histories of development, Fennoscandia, Volgo–Uralia and Sarmatia (Fig. 1a; Bogdanova, 1993; Gorbatschev and Bogdanova, 1993). The autonomous histories of the three segments are indicated by lithological and age differences between rock units. The zones that subdivide the EEC represent different types of collisional and/or accretional interactions. For example, the 2.1 Ga Lipetsk–Losev volcanic Belt and the East Voronezh Province mark the boundary between Sarmatia and Volgo–Uralia, and the 2.0–1.95 Ga Osnitsk–Mikashevichi Belt marks the position of the Sarmatia–Fennoscandia junction.

Palaeomagnetism has successfully been used for the reconstruction of movements of plates or

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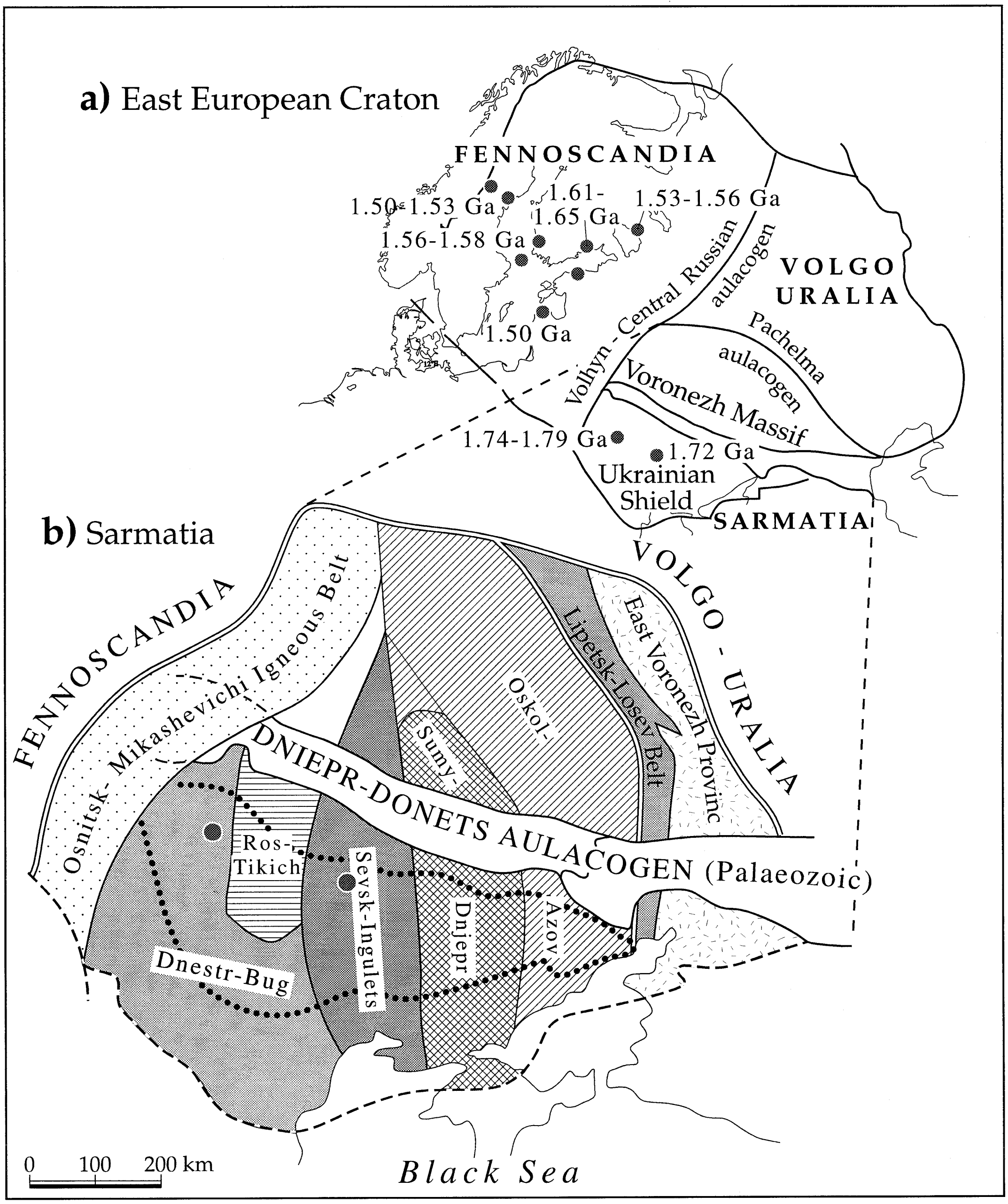

Fig. 1. (a) The East European Craton and the tectonic boundaries separating the Fennoscandian, the Volgo–Uralian and Sarmatian crustal segments (after Gorbatschev and Bogdanova, 1993). (b) The crustal domains in Sarmatia (Shchipansky and Bogdanova, 1996) and the Ukrainian Shield marked by the dotted line. The location of rapakivi–anorthosite complexes in the East European Craton is marked by filled circles together with the approximate age of the rocks.

continental blocks before they form the present day shields (e.g. Irving et al., 1984). In this study we want to test various hypotheses for the time of formation of the EEC. This we do by a palaeomagnetic study of rocks from the Ukrainian Shield as part of Sarmatia and by determining and comparing the positions of Sarmatia and Fennoscandia during the Proterozoic.

A first attempt to compare the palaeopositions (palaeolatitude and orientation) of the Ukrainian and Fennoscandian shields was made by Mikhailova and Kravchenko (1987). Their study indicated that the drift pattern of the Ukrainian Shield was significantly different from that of the Fennoscandian Shield. The time of consolidation of the EEC was suggested to be at 1.07–0.57 Ga. Subsequently, Elming et al. (1993) presented a more comprehensive attempt of Precambrian reconstructions of Fennoscandia and Ukraine based on databases that were compiled for the two shields. Unfortunately, the Ukrainian palaeomagnetic data were too sparse and the ages of magnetizations were often not well constrained, due to the absence of isotope age data, to yield reliable reconstructions. The reconstructions of Elming et al. (1993) indicated, however, similarities in positions, latitudinal drift and rotations of the two shields during the Palaeo-Mesoproterozoic, suggesting a close relationship between the shields in this period. Significantly different palaeopositions of Fennoscandia and Ukraine at ca. 1.2 Ga indicated that the time of final accretion of Fennoscandia to Sarmatia may have been in late Precambrian time, post 1.2 Ga, as was suggested by Mikhailova and Kravchenko (1987). With this study we want to improve the palaeomagnetic database for the Ukrainian Shield and test the reconstructions of Mikhailova and Kravchenko (1987) and Elming et al. (1993).

2. Geology

Sarmatia consists of five different Archaean domains (Fig. 1b), which are suggested to have been welded together into a coherent unit in the latest Archaean and earliest Palaeoproterozoic (ended 2.3–2.1 Ga ago; Bogdanova et al., 1996; Shchipansky and Bogdanova, 1996). The Oskol-Azov and Sevsk-Ingulets domains have been reworked substantially in the Palaeoproterozoic, while the Sumy-Dniepr Domain acted as a stable unit, however some anti-clockwise rotations of this domain may be indicated from dextral displacement of the Krivoy Rog–Kremanchug iron belt some 2.3–2.1 Ga ago (Shchipansky and Bogdanova, 1996). The Ros-Tikich Domain has mostly been metamorphosed in amphibolite facies in the early Palaeoproterozoic, while the relation to the ca. 3.4 Ga high grade Dnestr-Bug Domain is not clear.

Sarmatia is cut by the Palaeozoic Dniepr-Donets Aulacogen (DDA), separating the Voronezh Massif in the north from the Ukrainain Shield in the south. Lithological and structural similarities suggest that three of the Archaean domains have their equivalents on both sides of DDA, indicating that there has been no or only minor movements between these Archaean blocks since the Palaeozoic (Shchipansky and Bogdanova, 1996).

The Ukrainian Shield is located in the central part of southern Sarmatia (Fig. 1b) and consists of 3.5–0.6 Ga old metamorphic and magmatic rocks. The shield has also been intruded by numerous dyke swarms of various age, composition and direction of strike (e.g. Ahmetshina, 1975; Krutihovskaja et al., 1976).

The Fennoscandian Shield can be divided into a number of tectonomagmatic blocks, with the oldest part in the northeast (e.g. Gaál and Gorbatschev, 1987). The Archaean craton in the northeast was rifted in the earliest Palaeoproterozoic and partly reassembled during collisional orogeny at 2.0–1.8 Ga ago. This was followed by a gradual growth of the shield towards the southwest during several successive accretional events.

Sarmatia is different from Fennoscandia with respect to the chronology of crustal formation and by the extensive presence of Meso- and also Palaeoarchaean crust of 3.65–3.0 Ga (Bogdanova et al., 1996).

In the Mesoproterozoic anorthosite–rapakivi granite plutons intruded into the Fennoscandian and Ukrainian shields, now forming a discontinous belt running from the central and northwestern parts of the Ukrainian Shield to the central parts of the Fennoscandian Shield (Fig. 1a). The anorthosites and rapakivi granites in the Ukraine have been dated at 1.79–1.72 Ga (Amelin et al., 1994; Scherbak et al., 1995), while the plutons in Estonia, Finland and Sweden are younger and dated at 1.65–1.50 Ga

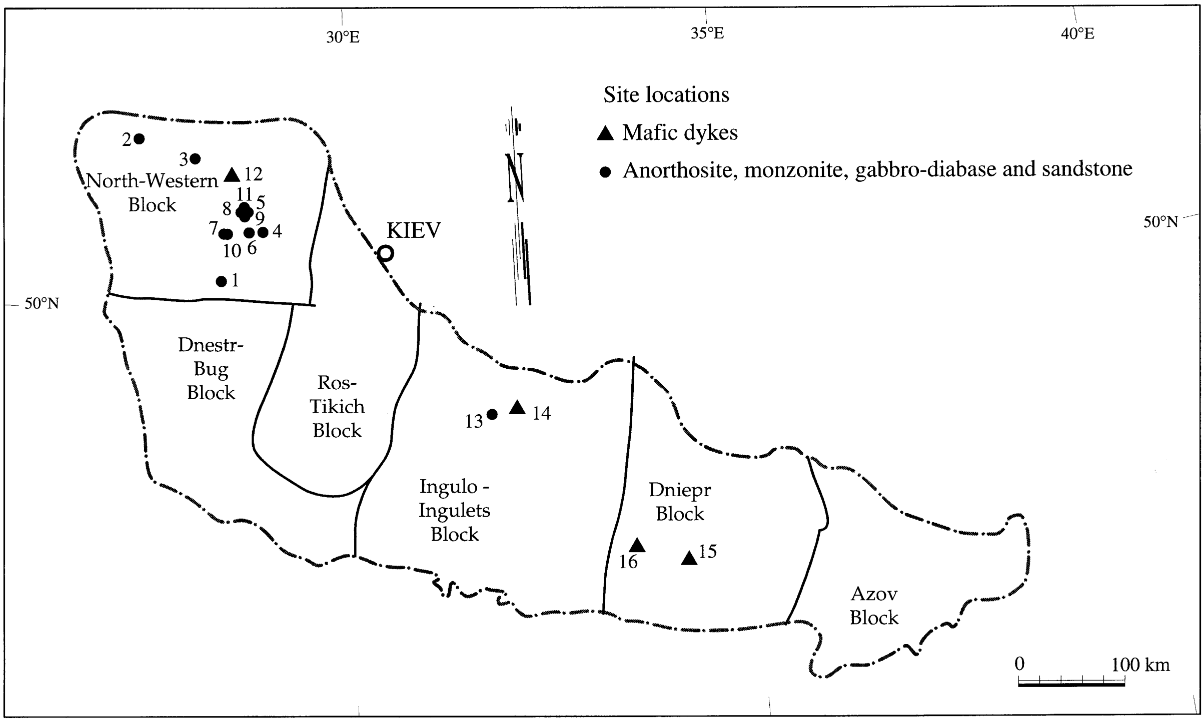

Fig. 2. The Ukrainian Shield and the different crustal blocks. The palaeomagnetic sites are marked by filled circles (anorthosite, monzonite, gabbro diabase and sandstone) and triangles (mafic dykes).

(Rämö, 1990; Vaasjoki et al., 1991; Kirs et al., 1997; Andersson, 1997). The Korosten anorthosite–rapakivi pluton in the North-Western Block of the Ukrainian Shield (Fig. 2) intrudes 2.43 Ga old metasedimentary rocks of the Teteriv Belt and volcanic-sedimentary rocks of ca 2.06 Ga (Scherbak et al., 1989). The age of the volcanic-sedimentary sequence was obtained from synorogenic granites of the Kirovograd–Zhitomir Complex.

3. Sampling

Before making a tectonic reconstruction of the Ukrainian Shield we need to know the time for welding of the various Archaean blocks of the shield. Previous geological studies indicate that the Ukrainian Shield may be regarded as one single craton at least since ca 1.7 Ga ago (Scherbak et al., 1981). However, to look for eventual tectonical differences between the blocks samples were collected from rocks of similar ages in the various blocks. The sampling was generally concentrated to rocks that have been radiometrically dated (U–Pb). This led us to focus on 1.7–2.0 Ga old rocks since rocks of these ages could be found in at least two blocks. Palaeomagnetic data of ages in this time sequence are also available from the Fennoscandian Shield (Pesonen et al., 1991), which would allow tectonic reconstructions of the Ukrainian Shield vs. the Fennoscandian Shield. Palaeomagnetic data from the Kola and the Karelian Archaean subprovinces in the Fennoscandian Shield (e.g. Khramov et al., 1997; Damm et al., 1997) suggest that no large scale movements have taken place within Fennoscandia since ca 2.12 Ga.

The major part of the samples was collected with a portable drill and the remaining ones as block samples. Orientations were done with magnetic and sun compasses. 682 samples were collected from 42 quarries and from a few outcrops in the North-Western Block, the Ingulo-Ingulets Block, the Dniepr Block and the Azov Block (Fig. 2). Here we report palaeomagnetic results from 16 quarries located in three of these blocks, although most of the data come from North-Western Block (Fig. 2).

The North-Western Block is composed of Proterozoic rock formations and includes the Osnitsa–Mikashevichi Igneous Belt, the Teteriv gneiss complex and the Korosten anorthosite–rapakivi pluton (Scherbak et al., 1998; Stepanyuk et al., 1998). The Palaeoproterozoic Teteriv metasediment-gneiss complex comprises the basement of the region and it is intruded by granitoids of the 2.08–2.06 Ga Zhitomir complex and rocks of the 2.02–1.98 Ga Bucky, Prutovka, Gorodnitsa and Kishin complexes (Scherbak et al., 1998). Rocks of the lower Teteriv strata are intruded by 2.43 Ga old igneous rocks of the Novograd–Volynsk Complex, which defines a minimum age for Teteriv metasediments. During a ca 170 Ma of tectonic stability sediments were deposited in depressions. These deposits, the Topilnya Series, of quartz sands and alevrite-clay sediments, are known in the Belokorovichy Structure. This was followed by anorogenic magmatism and the formation of rocks of the Korosten Complex 1.80–1.73 Ga ago (Amelin et al., 1994). The Korosten Pluton is one of the largest anorthosite–rapakivi plutons in the EEC, covering an area of ca 12 000 km^2. Anorthosite and gabbro-norite form ca 25% of the total area (Velikoslavinsky et al., 1978) and they have been intruded between 1.789 and 1.758 Ga ago (U–Pb of zircon; Amelin et al., 1994).

Rocks of predominantly anorthositic composition were collected in eight localities (sites 4–11) from the Korosten Complex. Basic dykes are related to the Korosten Complex (e.g. Amelin et al., 1994) and one dyke and its baked contacts of a 2.0 Ga Osnitsk granite (site 12) were sampled in an attempt to define an original magnetization of the rock. Samples were also collected from red and grey undeformed sediments of the Belokorovichy Structure (site 3). These sediments are deposited on ca 2.0 Ga old granodiorites and granites of the Osnitsk Complex and cut by ca 1.8 Ga old granitic dykes (Skobelev, personal communication, 1998), which give them an age in the interval 2.0–1.8 Ga. A ca 2.0 Ga monzonite (U–Pb; Skobelev et al., 1991) of the Bukinsky Massif (site 1) and a gabbroic dyke (site 2) of probably similar age were also included in the sampling.

In the Ingulo-Ingulets Block there is another anorthosite–rapakivi pluton, the Korson–Novomirgorod Pluton, which intruded into a heavely reworked Archaean terrain. This pluton is somewhat smaller than the Korosten Pluton and an anorthosite and a granite has been dated at 1.72 and 1.73 Ga, respectively (U–Pb; Scherbak et al., 1995). Samples were collected from the anorthosite (site13) and from a basic dyke (site 14), probably related to the anorthosite–rapakivi complex.

In the Dniepr Block the sampling was restricted to basic dykes of possibly Mesoproterozoic age (sites 15 and 16). These dykes intruded gneisses and granites of 3.20–2.95 Ga (Scherbak et al., 1989).

4. Palaeomagnetic results

The specimens were demagnetised by conventional alternating field and thermal demagnetization techniques and the remanent magnetizations were measured with a spinner (Czech-JR5) and a SQUID (2G-DC) magnetometer. The components of remanence were defined by least square fit of vectors in three-dimensional space (Torsvik, 1986; Torsvik et al., 1996) and the maximum angular deviation (MAD) is generally smaller than 6°.

4.1. North-Western Block

The oldest rocks in this study are the monzonite (site 1; Fig. 2) of the Bukinsky Massif and the gabbroic dyke (site 2) that intruded in a granite of the Osnitsk Complex in the North-Western Block. There are ten age determinations of rocks from the Bukinsky Massif and they all fall within a time span of 1.93–2.01 Ga. In the quarry of site 1, there are two age determinations of the monzonite, one is at 1.99 Ga and another is at 2.01 Ga (U–Pb, zircon; Skobelev et al., 1991). The natural remanent magnetization (NRM) of this monzonite is partly of low coercivity and the blocking temperature interval is wide (Fig. 3a). Apart from a soft component with a direction similar to that of the present earth magnetic field (PEF), a characteristic magnetization of high coercivity (20–65 mT) and unblocking temperature (500–570°C) has been defined in fifteen samples (65% of the samples), which form a fairly well defined site mean (decl. = 32.2°, incl. = 36.9°; Table 1) with an α_{95} of 8.1°. Other, but poorly defined directions are

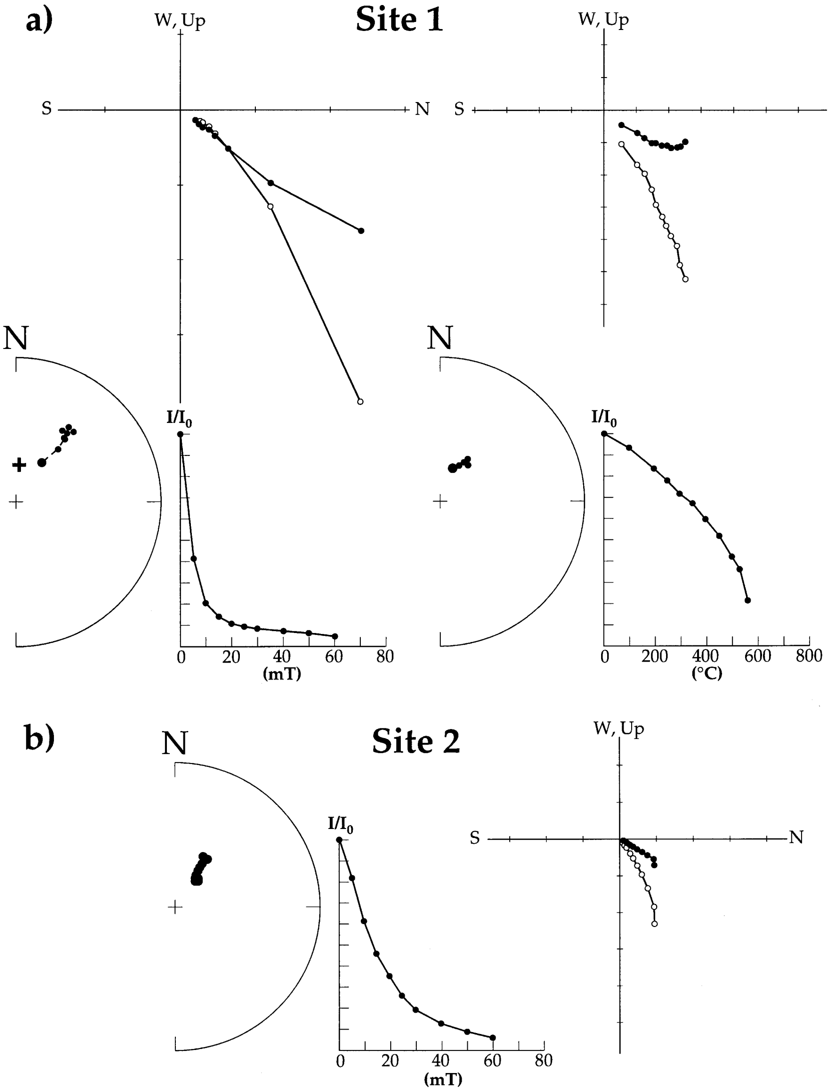

Fig. 3. Stereographic and vector plots showing examples of behavior of remanent magnetization during alternating field and thermal demagnetizations for (a); the 2.0 Ga monzonite at site 1 and (b); the gabbro–diabase at site 2. Both sites are located in the North-Western Block. In the stereographic plot the solid (open) symbols denote downward (upward) projections and the cross denotes the direction of the present Earth's field. In the vector plot the solid (open) symbols denote the end-point of the vector plotted on the horizontal (vertical) plane.

sometimes indicated at very high blocking temperatures (570–600°C). However, these directions of high inclinations are traced only in a few samples and are not consistent.

The gabbroic dyke does not have any clear chilled margins and it carries fragments of the country rock, which suggest that the dyke intruded when the host rock was still hot. This means that the age of the dyke is probably similar to that of the host rock, the ca 2.0 Ga Osnitsk granite. The coercivity of remanence is usually fairly hard with a median destructive field (MDF) of ca 14 mT. In the low coercivity range directions of magnetization similar to the PEF is usually observed (Fig. 3b), while characteristic magnetizations, forming a rather well defined mean (decl. = 45.5°, incl. = 53.7°, α_{95} = 7.8°; Table 1),

Table 1
Palaeomagnetic results from the Ukrainian Shield (note: Lat/Long = position of the sites; site = code for the locality; n = number of samples; P = polarity; Decl., Incl. = mean declination and inclination of characteristic magnetization; α_{95} = radius of 95% confidence circle; k = Fisher precision parameter (Fisher, 1953); P_{lat}, P_{lon} = latitude and longitude of the calculated pole positions; D_p, D_m = semiaxes of 95% confidence oval about the pole; A_{95} = the radius of the 95% confidence circle of the mean pole; Age = U–Pb, Zircon and baddeleyite and geologically estimated ages of the rocks (for references see text))

Lat/long site	n	p	Decl.	Incl.	α_{95}	k	P_{lat}	P_{lon}	D_p/D_m	A_{95}	Age (Ga)
North-Western block											
Monzonite and gabbro-diabase: ca 2.0 Ga											
50.15/28.37(1)	15	n	32.2	36.9	8.1	24	51.0	155.9	5.6/9.5	11.0	2.0
51.30/27.23(2)	10	n	45.5	53.7	7.8	39	58.3	126.9	7.6/10.9	13.3	
Sandstone-argillite: 2.0–1.8 Ga											
51.12/28.00(3)	18	n	26.9	11.0	7.9	20	39.2	172.5	4.1/8.0	9.0	2.0–1.8
Anorthosite: 1.77–1.74 Ga											
50.52/28.90(4)	19	r	217.3	19.1	7.3	22	21.5	168.9	5.2/10.3	11.5	1.76
50.70/28.72(5)	6	r	205.1	11.5	7.6	79	29.5	179.7	3.9/7.7	8.6	1.76
50.52/28.66(6)	17	n	42.5	−22.2	6.8	28	16.9	162.8	3.8/7.2	8.1	
50.52/28.41(7)	9	n	37.8	−21.6	8.3	39	20.1	168.6	4.6/8.8	9.8	
50.70/28.60(8)	10	r	216.0	15.8	4.8	103	23.5	169.2	2.5/4.9	5.5	1.74
50.67/28.66(9)	15	c	210.3	10.8	7.1	30	28.1	173.9	3.6/7.2	8.0	
50.52/28.47(10)	9	n	33.2	−3.8	14.3	14	30.4	169.1	7.2/14.3	16.0	1.77
50.72/28.65(11)	4	r	208.8	9.3	9.2	102	29.3	175.2	4.7/9.3	10.4	
Basic dyke											
50.97/28.48(12)	17	n	33.2	1.1	3.8	88	32.3	168.1	1.9/3.8	4.2	
Baked contact	7	n	39.6	0.2	10.0	38					
Host rock	4	n	20.6	38.7	33.3	9					
Ingulo-Ingulets Block											
Anorthosite: 1.72 Ga											
49.81/31.53(13)	18	c	51.7	−21.9	4.6	57	14.0	159.1	2.6/4.9	5.5	1.72
Basic dyke											
49.19/31.86(14)	18	n	45.6	−11.8	4.5	59	22.1	161.8	2.3/4.6	5.1	
Dniepr Block											
Basic dyke											
47.80/34.10(15)	16	r	205.9	44.1	3.8	95	12.8	190.3	3.0/4.8	5.7	
47.96/33.24(16)	15	n	43.1	−0.4	3.9	96	29.1	161.8	2.0/3.9	4.4	

are isolated after demagnetizations in field higher than 20 mT.

The sedimentary rocks of the Belokorovichy Structure were sampled at different levels. Grey sandstone and argillite are found in the lower levels, which are overlain by red and pink sandstone. The intensities of the natural remanant magnetizations are low ($0.1–1.0 \times 10^{-3}$ A m^{-1}) and the characteristic magnetization is generally carried by high coercivity magnetite and hematite (Fig. 4a). The directions of characteristic magnetization reveal dual polarity and fall in the first and third quadrants in the stereographic net. In the grey sandstone, it is sometimes not possible to separate the two components and the directions are then distributed along a great circle that connects the components of opposite polarity. The magnetization of the pink sandstone is sometimes soft, with a direction of magnetization similar to that of PEF. However, high coercivity and blocking temperature components of the pink sandstone have directions of two polarities and the majority of them are of low inclinations and fall in the first quadrant in the stereographic net. A

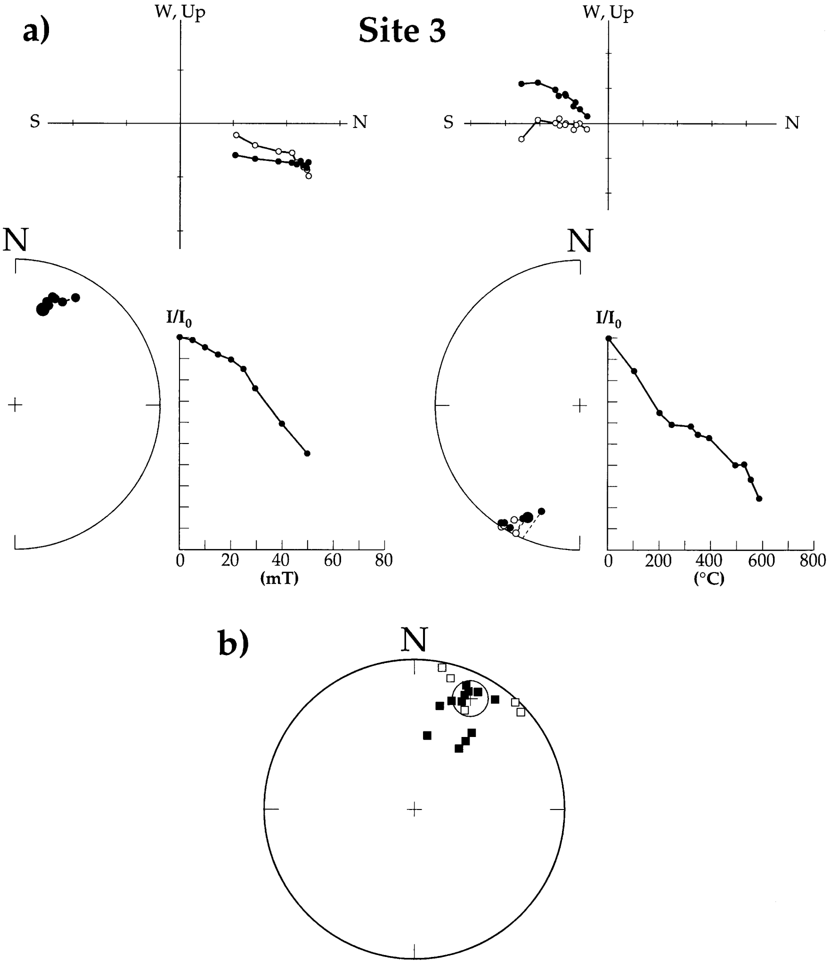

Fig. 4. (a) Example of remanence behaviour during alternating field and thermal demagnetizations of 2.0–1.8 Ga grey and red sandstone of the Belokorovichy Structure (site 3) in the North-Western Block. Note the dual polarity. Symbols as in Fig. 3. (b) The distribution of directions of the characteristic magnetization after inverting the reversed directions and the calculated site mean with circle of confidence.

clustering of directions in the first quadrant is also recognized for the red sandstone and the magnetization is often of very high coercivity.

The directions of characteristic magnetization of the grey, pink and red sandstone form, after tilt correction and inverting the reversed directions (Fig. 4b), a fairly well defined mean (decl. = 26.9°, incl. = 11.0°; Fig. 5; Table 1) with an $\alpha_{95} = 7.9°$.

The anorthosites (age ca 1.79–1.74 Ga) in the North-Western Block usually carry a stable remanence of high coercivity, which usually unblocks in a narrow temperature range at ca 580° (Fig. 5). Petrographical studies and the thermomagnetic analyses indicate that the carrier of remanent magnetization is thin isolated needle-like and lamellar ferromagnetics in plagioclase that is the result of disintegration of high temperature plagioclase and pyreoxenites at a late magmatic stage of rock formation (Mikhailova et al., 1994). The thickness of the lamella does not exceed 10 μm and the unblocking temperatures suggest that the carrier of the characteristic magnetization is magnetite. The site means are fairly well defined with α_{95} varying between 4.8 and 14.3° (Table 1) and with directions in the first and third quadrant of the stereographic net, which means that there is a difference in polarity of

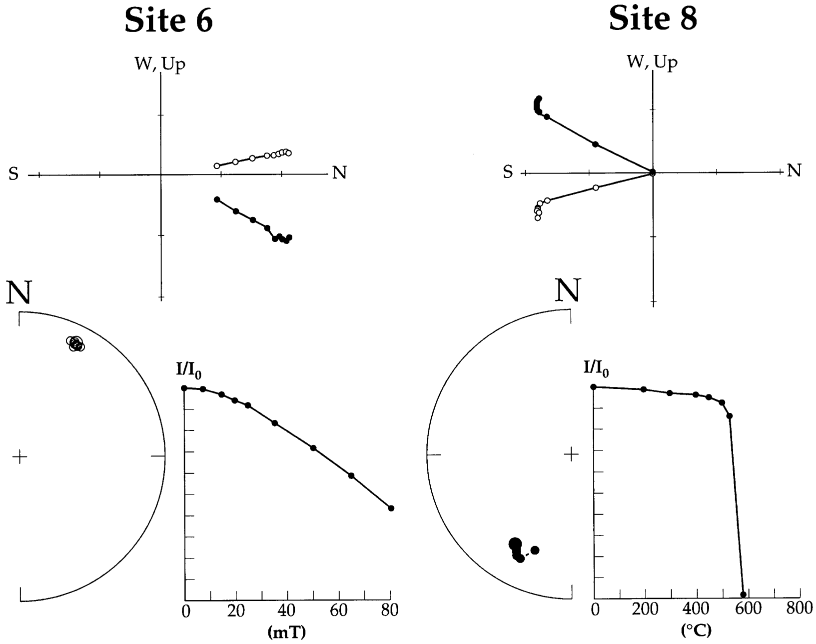

Fig. 5. Examples of alternating field and thermal demagnetizations of anorthosite of the Korosten Pluton in the North-Western Block. Note the difference in polarity between the sites. Petrological studies, the narrow unblocking range and temperature indicate that the carrier of remanence is thin lamellar magnetite. For conventions see Fig. 3.

the magnetization between sites (Fig. 6a). If the reversed magnetizations in the third quadrant are turned into normal (Fig. 6b), notice that the site means form a trend in the directions, with a gradual increase in inclinations related to an increase in declinations.

Some rocks that carry directions of magnetization of reversed polarity (sites: 4, 5 and 8; Fig. 6a; Table 1) have been radiometrically dated (U–Pb, zircon, baddeleyite) at 1.758 ± 0.001, 1.759 ± 0.0009 Ga (Amelin et al., 1994) and 1.744 Ga (Scherbak et al., 1995). A granite, closely related to the anorthosite at site 10, has also been dated and the poorly defined age falls into the time span of the anorthosites (U–Pb, 1.774 ± 0.023 Ga; Amelin et al., 1994). The anorthosite at site 10 carries a magnetization of normal polarity. However, there is no clear tendency of normal polarity magnetizations being older than magnetizations of reversed polarity.

U–Pb data from the Korosten Complex suggest that the different magmatic phases were emplaced during a period of at least 0.030 Ga. They were emplaced as a series of distinct igneous episodes, with an interval of

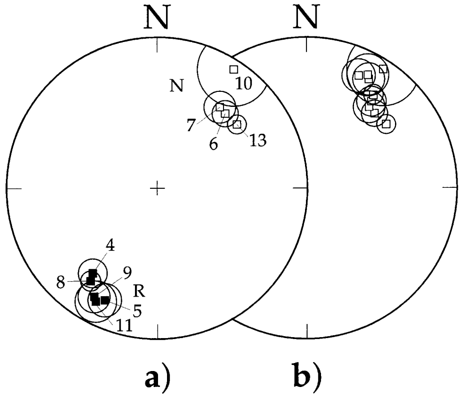

Fig. 6. (a) Site mean directions with confidence circles of the characteristic magnetization for anorthosite samples from the Korosten (sites 4–11) and Korson Novomirgorod (site 13) plutons of the North-Western and Ingulo-Ingulets blocks, respectively. (b)The distribution of site mean directions after inverting the reversed magnetizations. Note the trend of mean directions varying from ca decl. = 33°, incl. = − 4° to decl. = 52°, incl. = − 22°.

0.006–0.015 Ga between the pulses (Amelin et al., 1994). The gradual change in site mean directions may reflect this trend in ages.

4.2. Ingulo-Ingulets Block

In the Ingulo-Ingulets Block we also have palaeomagnetic results from a radiometrically dated anorthosite (site13; Fig. 2). This anorthosite of the Korson-Novomirgorod Pluton has an age of 1.720 ± 0.010 Ga (U–Pb; Scherbak et al., 1995). After erasing a soft component with a direction similar to PEF a hard component is isolated, which is stable in demagnetizing fields up to 90 mT. The direction of this high coercivity magnetization (decl. = 51.7°, incl. = − 21.9°; Table 1; Fig. 7) is similar to those of the anorthosites of the Korosten Pluton.

4.3. Basic dykes

Dykes are important for palaeomagnetic studies not only because they often carry a stable magnetization, but they also give an opportunity for testing of eventual remagnetizations.

Basic dykes were collected from all three blocks (sites 12, 14, 15 and 16; Fig. 2). In the North-Western Block, these dykes cut Korosten rapakivi granite and

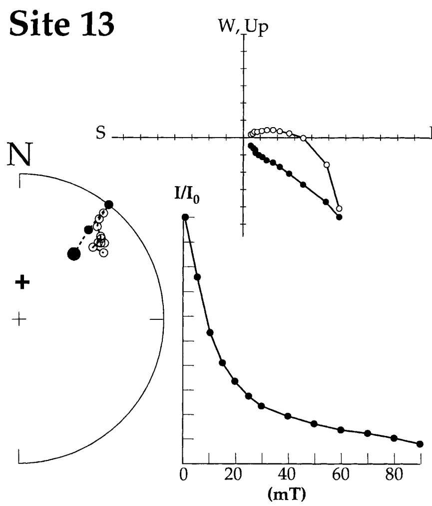

Fig. 7. A stable component of remanence is isolated in the anorthosite of the Korson Novomirgorod Pluton after demagnetization in an alternating field of ca 25 mT. For conventions see Fig. 3.

older granites. For one dyke, the Behi dyke (site 12) it is possible to demonstrate a positive baked contact test (Fig. 8a). This is because the direction of the characteristic magnetization of the dyke is not different from that of the baked ca 2.0 Ga old Osnitsk granite, but significantly different from that of the unbaked granite. The mean direction of magnetization of the unbaked granite (decl. = 20.6°, incl. = 38.7°; Table 1) is poorly defined, but significantly different from the magnetization of the dyke and the baked granite. The magnetization of the basic dyke (decl. = 33.2°, incl. = 1.1°, Table 1) is therefore proven to be original. The mean direction of the host rock is similar to the mean direction determined for the 2.0 Ga old gabbro-monzonite of the Bukinsky massif (decl. = 32.2°, incl. = 36.9°; site 1) and may therefore also be original. Basic dykes are suggested to have intruded in relation with the emplacement of anorthosite at ca 1.76 Ga ago (Amelin et al., 1994), which explains why the direction of magnetization of the dyke is similar to those of the anorthosites (e.g. sites 5 and 10; Table 1; Fig. 6a).

Also in the Ingulo-Ingulets Block a basic dyke (site 14) carries a well defined characteristic magnetization (decl. = 45.6°, incl. = − 11.8°; Table 1), however no baked contact test results are available for this site. We note, however, that the direction of magnetization of this dyke is similar to that of the anorthosite in this block (site 13; Fig. 7) and therefore resemble and relate to the situation in the Korosten Pluton.

Palaeomagnetic results from two basic dykes in the Dniepr Block (sites 15 and 16; Fig. 2) indicate that there may be two different generations of dykes. The characteristic magnetizations are of high to relatively high coercivities and have unblocking temperatures that suggest the carriers to be fine-grained magnetite. The site mean directions (site 15: decl. = 205.9°, incl. = 44.1°; site 16: decl. = 43.1°, incl. = −0.4°; Table 1; Fig. 8b) are well defined ($\alpha < 4°$) and significantly different from each other. The remanence direction of site 16 is similar to the directions identified in the anorthosites and dykes that we find in the North-Western Block and in the Ingulo-Ingulets Block, while that of site 15 has not been recognized elsewhere.

The positive baked contact test of the dyke at site 12, carrying a direction very similar to those of the anorthosites, indicates that the magnetization is

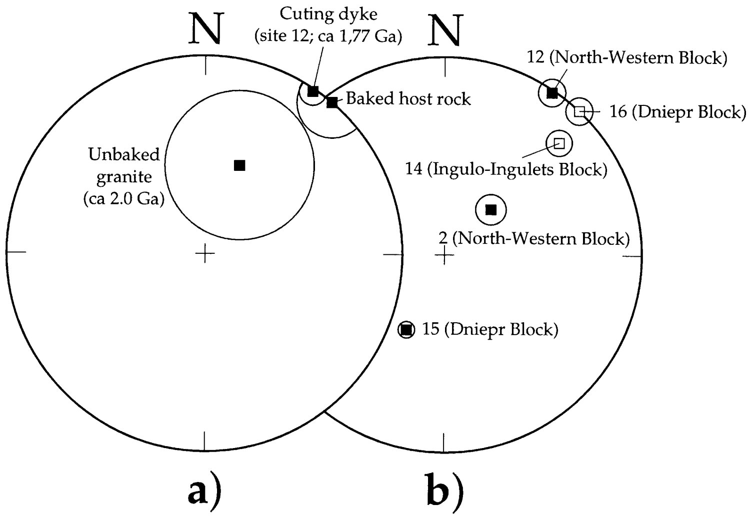

Fig. 8. (a) The mean direction and confidence circles of a mafic dyke, the baked and the unbaked granite host rock in the North-Western Block (site 12). The significantly different directions of remanence identified in the unbaked ca 2.0 Ga granite demonstrate an example of a positive baked contact test. (b) Mean remanence directions identified in mafic dykes from the North-Western Block (site 12), the Ingulo-Ingulets Block (site14) and the Dniepr Block (sites 15 and 16).

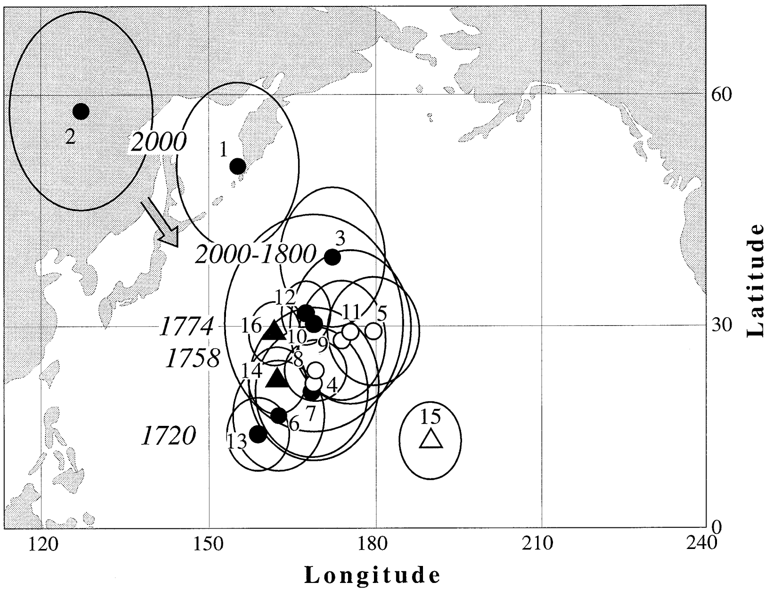

Fig. 9. Palaeomagnetic poles and the 95% confidence circles calculated from the mean directions of magnetization of Palaeo-Mesoproterozoic monzonite, sandstone, anorthosite and mafic dykes in the Ukrainian Shield. The codes used for the poles are as in Fig. 2. Closed (open) symbols denote normal (reversed) polarity.

primary. The absence of signs of metamorphism and the presence of dual polarities of magnetization in the anorthosites also suggest a primary origin of the remanent magnetizations. Since the directions of magnetizations of the anorthosites and the basic dykes from the various blocks are similar, it also suggests that there has been no, or only little, relative movements between the blocks since ca 1.77 Ga ago. This means that the Ukrainian Shield can be treated as one coherent unit since that time.

4.4. Virtual geomagnetic poles

As there are no signs of remagnetizations and the tectonical differences between the various blocks within the Ukrainian Shield and between the Ukrainian Shield and the Voronez Massif seem insignificant, the data justify calculations of virtual geomagnetic poles (VGPs) representing the whole Sarmatia for ages of ca 2.0 Ga and younger. The VGPs calculated for the anorthosites of the

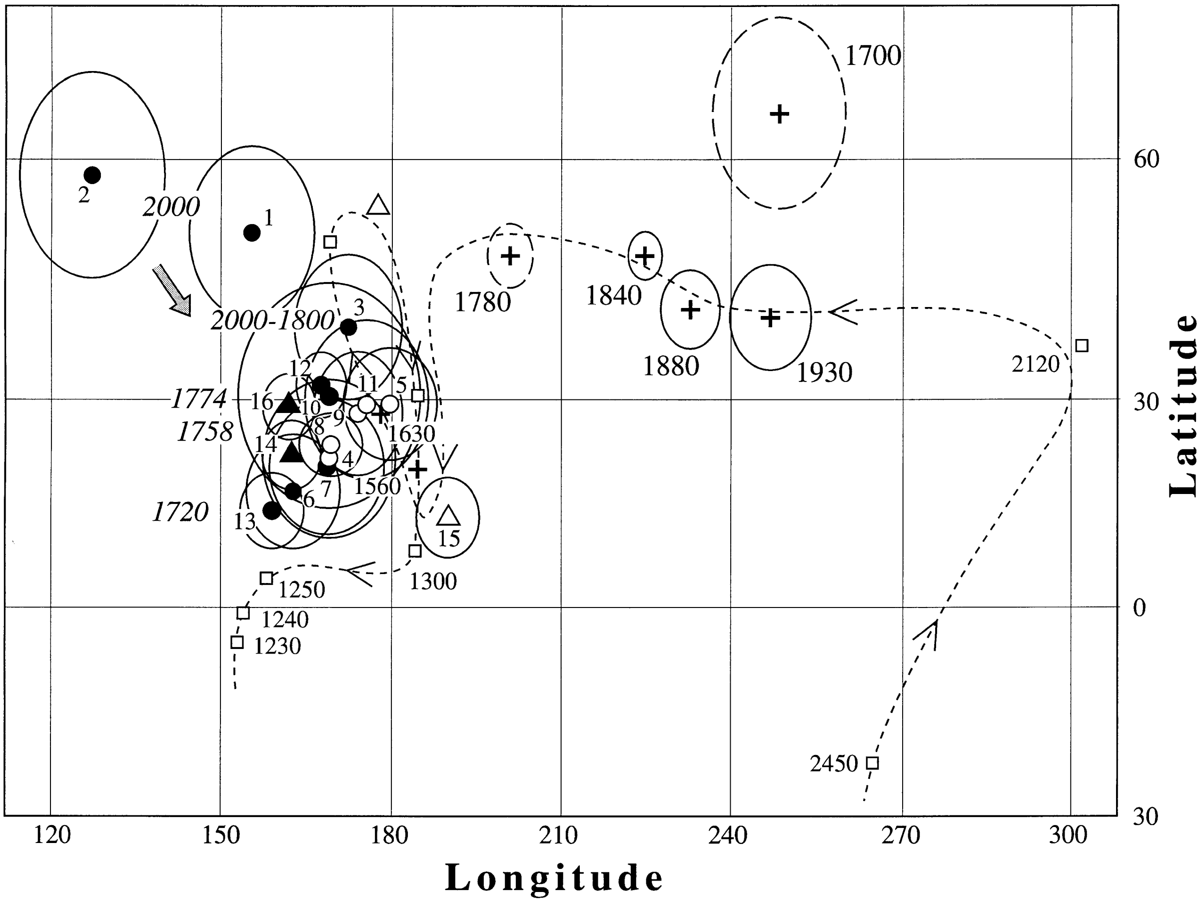

Fig. 10. Palaeomagnetic poles and U–Pb and geologically estimated ages of Palaeo-Mesoproterozoic rocks from three crustal blocks of the Ukrainian Shield. The poles 13 and 14 are from rocks in the Ingulo-Ingulets Block, the poles 15 and 16 are from rocks of the Dniepr Block, while the rest of the poles orignate from the North-Western Block. The gradual change in pole positions is related with a change in age from ca 2.0 to 1.72 Ga, indicating that the distribution of poles is the result of apparent polar wander. The codes for the Ukrainian poles are as in Fig. 9. Fennoscandian reference poles (Mertanen and Pesonen, 1997), and two other not as well defined poles (hatched confidence circles; Elming, 1994; Mattsson and Elming, 1999) within the age interval 1.93–1.56 Ga, are denoted by crosses. A segment of the APWP for the Fennoscandian Shield (Elming et al., 1993) is marked by the hatched line and grand mean poles are marked by open squares. Closed (open) symbols for the Ukrainian poles denote normal (reversed) polarity.

North-Western Block form a pattern of poles with positions varying from ca 30°N/180°E to 17°N/163°E (Fig. 9; Table 1). The poles of the 2.0 Ga Bukinsky Massif (Pole 1; $P_{lat} = 51.0$, $P_{lon} = 155.9$) and the gabbro-diabase of similar age (Pole 2; $P_{lat} = 58.3$, $P_{lon} = 126.9$) are located at higher latitudes, significantly different from those of the younger anorthosites and dykes (poles 4–16). The pole of the Bukinsky massif is similar to one of the poles calculated from ca 1.97 Ga old granite, granodiorite and diorite collected in one locality in the Mikachevichi Belt ($P_{lat} = 56.7$, $P_{lon} = 168.5$; Iosifidi et al., 1998). Also the pole calculated for the ca 2.0–1.8 Ga sedimentary rocks ($P_{lat} = 39.2$, $P_{lon} = 172.5$) is significantly different from the poles of the 1.77–1.74 Ga anorthosites and it is located between the poles of the Bukinsky massif and those of the anorthosites. The VGP for the anorthosite of the Ingulo-Ingulets Block ($P_{lat} = 14.0$, $P_{lon} = 159.1$) falls into the pattern of poles of the North-Western Block and also the VGPs of the basic dykes (Fig. 9; Table 1) of the different blocks follow the same pattern, except for one (site 15) that is significantly different.

Plotting the U–Pb ages and geologically estimated ages for the corresponding poles (Fig. 10), the changes in pole positions seem to be reflected by a gradual change in age from higher ages of poles at high latitudes to lower ages for poles at low latitudes. This means that the distribution of poles may be explained by apparent polar wander. The 2.0 Ga pole (pole 1) and the 2.0–1.8 Ga pole (pole 3) fit into this pattern and it is likely that we have managed to define a 2.0–1.72 Ga sequence of the APWP for the Ukrainian Shield and Sarmatia.

4.5. Comparison with VGPs of Fennoscandia

Comparing the VGPs for the Ukrainian Shield with the apparent polar wander path (APWP) of Fennoscandia (Elming et al., 1993), the sequence of the APWP for the Ukraine is significantly different from that of corresponding age interval of Fennoscandia (Fig. 10). However, the APWP for Fennoscandia does not represent a continuous time series of poles, but rather the successive orogeneses and a direct comparison is thus not necessarily possible. The Ukrainian poles (Table 1) and the reference poles for Fennoscandia in the time interval 2.0–1.6 Ga (Table 2) are generally not coeval and this part of the APWP for Fennoscandia is partly poorly constrained by highly graded and well dated poles. This means that we may consider the possibility that the Fennoscandian and Ukrainian shields were joined at ca 2.0 Ga and in such a case the Ukrainian poles could fill the gap of missing poles for the Fennoscandian APWP.

Table 2
Palaeomagnetic poles for Fennoscandia (Note: Rock names given in italics refer to key poles, while other rock names refer to palaeomagnetic poles which are not as well defined. *B/N/n* denote number of formations/sites/samples. * denote the statistical level used in the mean calculation. D_{ref}, I_{ref} are palaeomagnetic directions calculated with respect to the reference location for Fennoscandia (Kajaani, 64.1°N, 27.7°E). P_{lat}, P_{lon}, Latitude and longitude of the poles. A_{95} denote the half-angle of the 95% confidence circle of the poles. Age, radiometrical and palaeomagnetically estimated ages of the rocks. References: (1) Mertanen and Pesonen, 1994; (2) Mertanen and Pesonen, 1997; (3) Neuvonen et al., 1981; (4) Elming, 1994; (5) Damm et al., 1997; (6) Khramov et al., 1997; (7) Mattsson and Elming, 1999; (8) Bylund and Elming, 1992)

Rock (nb)	*B/N/n*	D_{ref}	I_{ref}	P_{lat}	P_{lon}	A_{95}	Age (Ma)	Ref.
Tsuomasvarri gabbro-diorite	1/8*/27	329.0	34.4	40.2	247.3	6.8	1931±2	1
Mean syn-late SF gabbros	4*/21/130	340.3	31.7	41.2	233.0	4.9	1880	2
Haukivesi lamprophyres	12*/12/25	347.5	40.2	48.0	225.0	2.9	1837–1840	3
Post SF intr.	5*/17/82	5.0	39.0	48.0	201.0	4.0	1780	4
Rybeka sill	2/7*	357.6	22.9	37.8	210.6	5.5	1770±12	5
Shoksha-Sheltozero diabase	3/8*	354.5	15.9	33.9	214.3	10.0	1770±12	6
Rätan granite	1/3/29*	337.9	61.2	64.8	248.5	11.5	1702±6	7
Dala porphyry, Blyberget	1/5*	347.8	66.0	73.0	236.0	12.8	1700	8
Dala porphyry, Hållan	1/6*	15.0	78.5	83.0	79.0	10.6	1700	8
Mean Subjotnian quartz-porphyry dykes	2*/12/14/55	26.3	10.4	28.2	177.7		1630	1
Mean Subjotnian diabase dykes	2*/13/13/50	21.7	−7.8	20.1	184.6		1560	1

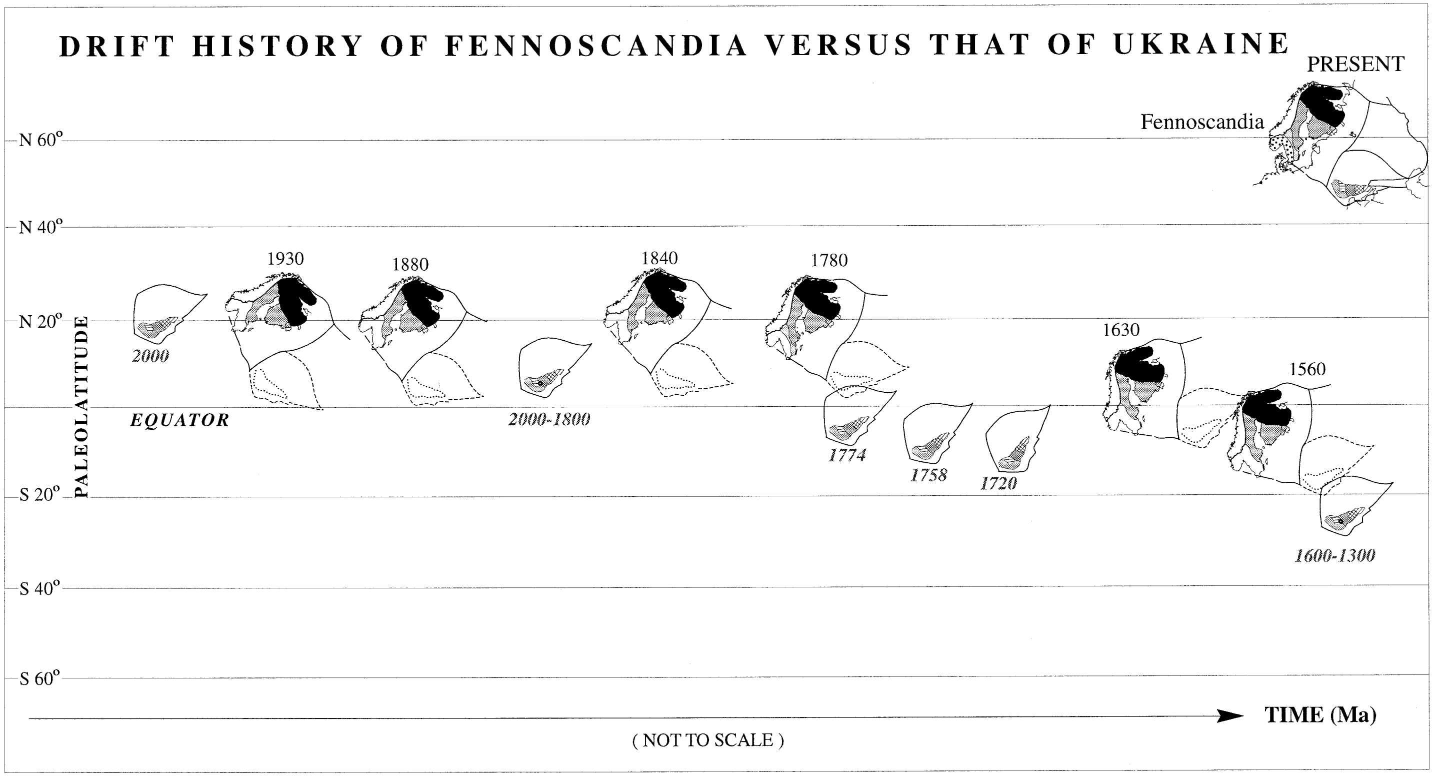

Fig. 11. The drift history of the Ukrainian Shield as a part of Sarmatia vs. that of Fennoscandia from ca 2.0 to 1.56 Ga. The shields are plotted in palaeomagnetically defined orientations and palaeolatitudes. They have been shifted arbitrary to the right since their longitudes cannot be determined by palaeomagnetic methods (see Pesonen et al., 1989). The tectonic reconstructions for Fennoscandia are based on key palaeomagnetic poles, except for the reconstruction at 1.78 Ga, which is based on the more poorly defined pole of PSF intrusions.

From Svecofennian rocks in the Fennoscandian Shield 1.84 and 1.88 Ga key poles ($P_{lat} = 48°$, $P_{lon} = 225°$ and $P_{lat} = 41°$, $P_{lon} = 233°$, respectively; Table 2) can be calculated from the mean direction of lamprophyres (Neuvonen et al., 1981) and of a number of well dated (U–Pb) gabbro massifs (e.g. Mertanen and Pesonen, 1997). Also the 1.93 Ga old pole of the Tsuomasvarri gabbro-diorite ($P_{lat} = 247°$, $P_{lon} = 40°$; Mertanen and Pesonen, 1994) may be considered as a key pole. This means that for the Fennoscandian Shield there are three key poles, 1.93, 1.88, and 1.84 Ga that may be used for a comparison with the 2.0, 2.0–1.8 Ga poles of Ukraine

Fennoscandian poles calculated from magnetizations ages of ca 1.8 Ga are not so frequent and no key poles have been determined for such an age. However, in northern Sweden (Post Svecofennian intrusions: decl. = 5°, incl. = 39°; Elming, 1994; Table 2) and in the western part of Russia (e.g Rybeka sill: decl. = 358°, incl. = 23°; Damm et al., 1997; Shoksha-Sheltozero diabase: decl. = 355°, incl. = 16°, Khramov et al., 1997) poles have been calculated for rocks of 1.78 and 1.77 Ga, respectively. These pole positions (e.g. the Post Svecofennian intrusions: $P_{lat} = 48°$, $P_{lon} = 201°$; Table 2; Fig. 10) fall into the trend of APW for Fennoscandia.

There is new data from a 1.70 Ga old granite (U–Pb; Delin, 1996) in central Sweden that is comparable in age with the youngest anorthosites (1.720 ± 0.010 Ga) of our study in the Ukraine. This granite, the Rätan granite, has been sampled in three sites and a characteristic magnetization has been isolated in high coercivity and blocking temperature ranges (decl. = 337.9°, incl. = 61.2°; Mattsson and Elming, 1999; Table 2). The mean direction is similar to what has been defined from some ca 1.70 Ga (U–Pb; Lundqvist and Persson, 1996) old Dala-porphyries (e.g Blyberget: decl. = 348°, incl. = 66°; Hållan: decl. = 15°, incl. = 79°; Bylund and Elming, 1992). However, the calculated pole position for the Rätan granite ($P_{lat} = 64.8°$, $P_{lon} = 248.5°$; Table 2; Fig. 10) is very different from an expected ca 1.70 Ga pole of the APWP of Fennoscandia (ca $P_{lat} = 40°$, $P_{lon} = 185°$; Elming et al., 1993) and as yet no test for the origin of magnetization has been performed, this pole must still be regarded as uncertain.

When comparing the Ukrainain 2.0 and 2.0–1.8 Ga poles with the 1.93, 1.88, and 1.84 Ga poles of Fennoscandia it is clear that the pole positions of the Ukraine are very different and do not fall into the pattern of APW for Fennoscandia. For the Ukrainian poles of 1.77–1.72 Ga we may compare the positions with the 1.78 Ga pole of Fennoscandia, even though that pole is not a key pole. Also these poles are significantly different from each other, which indicates that the Ukrainian Shield was not in its present position relative to Fennoscandia during the period 2.0–1.78 Ga.

The pole position of the basic dyke at site 15 ($P_{lat} = 12.8°$, $P_{lon} = 190.3°$; Table 1) is similar to the 1.56 Ga reference pole of Fennoscandia ($P_{lat} = 20.1°$, $P_{lon} = 184.6°$; Table 2, Fig. 10), but also to the 1.3 Ga grand mean pole ($P_{lat} = 8°$, $P_{lon} = 184°$) of the APWP (Elming et al., 1993). It is still too early to say if this pole represents a time when the Fennoscandian Shield was accreted to the Ukrainian Shield into a relative position similar as today. However, if it does, it means that the time of accretion of Fennoscandia to Ukraine should be sometimes between 1.78 and 1.56 or 1.3 Ga.

5. The drift of the Ukrainain Shield (and Sarmatia) vs. Fennoscandia

Even if the palaeomagnetic data for the Fennoscandian and the Ukrainian shields in the age interval 2.0–1.63 Ga are not perfectly coeval, the data may justify the calculation of positions and orientations of the Ukrainian Shield as a part of Sarmatia for a comparison with that of Fennoscandia during the Palaeo-Mesoproterozoic.

The orientations and/or the positions of the Ukrainian Shield at 2.0–1.8 Ga were different from the present day position of Ukraine relative Fennoscandia as calculated from the Fennoscandian reference poles at 1.93, 1.88, and 1.84 Ga (Fig. 11). At 2.0 Ga the Ukrainian Shield moved from a position at 20°N to occupy a position close to the equator in the time interval 2.0–1.8 Ga. The position was then similar to its present relative position vs. Fennoscandia, however, the orientation of the Ukraine was 50° different. In the time interval 1.78–1.72 Ga there are no key poles for Fennoscandia and tectonic reconstructions in this interval are therefore less reliable. At 1.78 Ga the Ukrainian Shield still occupied a position close to the equator and south of its present relative position vs. Fennoscandia. At this time the orientation was also different and from 1.78 to 1.72 Ga Ukraine rotated counterclockwise. A tectonic reconstruction for Fennoscandia on basis of the ca 1.70 Ga pole is still not justified by the data due to uncertainties of the origin of magnetization.

With reference to the earlier work by Elming et al. (1993), the orientation and estimated position of Ukraine was similar to that of Fennoscandia around 1.3 Ga, possibly indicating that Fennoscandia at that time was already accreted to Ukraine. On basis of pole 15 both the orientation and the position of the Ukrainian Shield is close to its present day position relative Fennoscandia (Fig. 11). If this pole represents a magnetization at ca 1.56 Ga, it means that the Fennoscandian Shield may have accreted to the Ukrainian Shield and Sarmatia into its present relative positions sometimes during 1.72–1.56 Ga.

6. Discussion

There are evidences of several stages of accretion and westward growth of Palaeoproterozoic crustal domains in the western EEC marked by a number of rock belts of different lithologic and tectonic patterns (Bogdanova et al., 1996). According to seismic reflection data (Stephenson et al., 1996) the complexes of stacked Fennoscandian terranes plunge southeastwards beneath the edge of Sarmatia. Alkaline magmatism and high-T metamorphism affected the Sarmatian crust at ca 2.1 Ga ago, presumably in connection with the beginning of southeast directed subduction of oceanic crust beneath the Sarmatian edge (Bogdanova and Gorbatschev, 1998). The crustal terranes adjacent to Sarmatia are bounded by fault zones with ages of blastomylonites between ca 1.80 and 1.65 Ga (Bogdanova and Gorbatschev, 1998), which may indicate a time period of accretion of Fennoscandia to Sarmatia. In the Middle Riphean (1.35–1.05 Ga) rifting took place along the present day margins of Baltica (Nikishin et al., 1996). This coincided in time with the intra-plate Jotnian rifting, which may suggest that Fennoscandia at this stage was joined with the Ukrainian Shield. The time for final accretion of Fennoscandia to Ukraine sometimes after ca 1.8 Ga, as suggested by palaeomagnetic data seems therefore likely.

For further constraints on the tectonic reconstructions and timing of the accretion of the Fennoscandian Shield to the Ukrainian Shield, the drift velocities of the shields may be of help. Changes in plate motions seem to be reflected by intra plate stresses that are preferentially controlled by processes affecting the plate boundaries (Zoback, 1992; Zoback et al., 1993) due to interactions with other plates. For Fennoscandia, the Jotnian rifting (1.35–1.25 Ga) and the Sveconorwegian–Grenvillian orogeny (1.25–0.8 Ga) are characterized by anomalously high drift rates (Elming et al., 1993). The drift rates are generally higher in the Neoproterozoic when compared with that of Mesoproterozoic, and from 1.7 to 1.35 Ga the rate of APW and latitudinal drift is fairly low. Present-day motions show higher drift rates for smaller plates than for larger plates (Minster and Jordan, 1978). Therefore, it is possible that Fennoscandia in this time interval may have been a part of a larger plate. This is supported by tectonic reconstructions of Buchan et al. (2000), who proposed that Fennoscandia and Laurentia were united at ca 1.25 Ga. Hoffman (1989a) suggested, on basis of geological and isotope data, that a supercontinent was assembled in the Mesoproterozoic. This supercontinent may then have included Laurentia, Fennoscandia and probably also the Ukrainian Shield as a part of Sarmatia. The lithological and structural similarities that exist when comparing the Voronezh Massif and the Ukrainian Shield, and the lack of large scale tectonical differences within the Ukrainian Shield since at least 1.7 Ga ago, suggest that Sarmatia acted as one coherent unit in the Mesoproterozoic. Sometimes after ca 1.25 Ga Fennoscandia and Sarmatia split up from Laurentia (Park, 1992), which may be reflected by the increasing drift rate of Fennoscandia.

The tectonic environment where the rocks of the anorthosite–rapakivi complexes intruded is one of the major problems in understanding the Precambrian geology. Different models of the origin have been presented and the action of mantle plume or mantle diapirism in an extensional environment is one (Morse et al., 1988; Anderson and Bender, 1989). Other models, like a convecting upwelling mantle plume (Hoffman, 1989b), thermal response to crustal thickening by earlier convergent plate tectonics (Van Schmus and Zietz, 1987), and underthrusting (Duchesne et al., 1998), have been discussed.

In the Fennoscandian Shield there are seven major rapakivi complexes and ca fifteen minor ones, extending from the Salmi batholite in Russia in the east over southern Finland, Estonia and Latvia to central Sweden in the west (Fig. 1a). There seem to be a systematic distribution of intrusion ages for the rapakivi rocks. Apart from the Salmi batholite, there is a westward trend of decreasing age of the complexes, from the Wiborg batholite (1.65–1.62 Ga, Vaasjoki et al., 1991) in the east to a number of small complexes (1.53–1.50 Ga, Andersson, 1997) in the west. The rapakivi magmatism occurred in areas of different crustal thickness, however, some of the largest complexes are found in an east–west trending zone where the crust is thinner (Korja et al., 1993; Korja and Heikinen, 1995) and from seismic data mafic underplating is indicated by highly reflective lamellar structure in relation with the Åland rapakivi complex (Korja and Heikinen, 1995).

The Korosten Pluton is a ca 6 km thick layered intrusion and seismic reflection and gravity data indicate that also here the crust below the intrusion is extensively intruded by mafic intrusions (EUROBRIDGE Seismic Working Group, 1998).

In the Sveconorwegian (Grenvillian) Province of Fennoscandia many granitoids that intruded subcontemporaneously with anorthosites are found in relation with major boundaries between terranes (Duchesne et al., 1998). Deep seismic data indicate that these boundaries become surfaces along which underthrusting has taken place. It has been suggested that the anorthosites were produced from melting of tongues of lower mafic crust, tongues that were produced in relation with large relative movements between terranes (Duchesnes et al., 1998). However, for Fennoscandia in general there are no clear relations between the occurrence of rapakivi intrusions and terrane boundaries.

In a model presented by Windley (1993) the Palaeo-Mesoproterozoic anorogenic igneous intrusions formed as a result of crustal thickening related to an earlier compressional event. When the crust was thick enough the deeper parts began to melt and rapakivi granites intruded into an extended crust. A similar process of crustal thickening was also suggested to be the cause of rapakivi granites in an Archaean terrain in northern Laurentia. The compressional regime was here the result of continental collision and the granites formed ca 30 Ma after the compressional event. For Fennoscandian this model seems less likely as no crustal thickening is related to the rapakivi complexes.

An alternative model for the formation of the rapakivi–anorthosite is the action of a mantle plume. Palaeomagnetic data from rapakivi granites and related dykes in Fennoscandia show that the rocks were intruded in palaeolatitudes between 0 and 27° north (Moakhar and Elming, 1998). The data also indicate a tendency of increasing palaeolatitudes with decreasing ages of the rocks (ca 1.65–1.50 Ga). A similar pattern of changing palaeolatitudes is also noticed from the anorthosites in the Ukrainian Shield, with an increase in palaeolatitudes from ca 2° south to 11° south during 1.77–1.72 Ga. This means that if a mantle plume is the origin of the anorthosites it has not been stationary. Both the Fennoscandian Shield and the Ukrainian Shield occupied low latitudinal positions during the intrusions of the anorthosite–rapakivi complexes and they may have passed over a moving hot spot located close to the equator. However, with such a model it is still hard to explain the wide geographical distribution of rapakivi intrusions of similar ages in Fennoscandia.

More geological and geophysical constraints are needed for a final reconstruction of the tectonic environment of these anorogenic intrusions.

7. Conclusions

Primary remanent magnetizations have been isolated in 1.77–1.72 Ga anorthosites and basic dykes and probably also in 2.0 Ga gabbro-monzonite, a gabbro-diabase of similar age, and in 2.0–1.8 Ga sandstone. By comparison of palaeomagnetic results from rocks of similar age (i.e. 1.77–1.72 Ga) from different blocks it is suggested that the Ukrainain Shield can be regarded as a coherent unit at least since ca 1.77 Ga ago. Geological data also indicate that there has probably not been any large scale tectonic movements within Sarmatia since the Mesoproterozoic. This means that the tectonic reconstructions for the Ukrainian Shield may also include Sarmatia. On basis of the new palaeomagnetic results VGPs were calculated for the Ukrainain Shield and an APWP from 2.0 to 1.72 Ga was defined. This sequence of poles do not coincide with poles of similar ages from the Fennoscandian Shield. Although the poles are not perfectly coeval, the tectonic reconstructions demonstrates that the orientation and position of the Ukrainian Shield in that time interval was different from the present position relative to Fennoscandia. One, not dated, pole from the Ukrainian Shield falls on the ca 1.6 or 1.3 Ga part of the Fennoscandian APWP, which may represent a time when Fennoscandia was joined with the Ukrainian Shield. Rifting in the Fennoscandian Shield and in the boundary between Sarmatia and Fennoscandia occurred contemporaneously at 1.35 Ga, indicating that Fennoscandia at this time already had been accreted to Ukraine. The palaeomagnetic data therefore suggest that the final accretion of Fennoscandia to the Ukrainian Shield as a part of Sarmatia took place sometimes after ca 1.8 Ga, an interpretation which is supported by the 1.8–1.65 Ga ages of blastomylonites in fault zones in the crustal terranes adjacent to Sarmatia. Fennoscandia and Sarmatia may then have formed a part of a supercontinent that was assembled in the Mesoproterozoic.

Different models for the tectonic environment of the anorthosite–rapakivi intrusions have been discussed. Signs of underplating and to some extent a thinning of the crust in relation with rapakivi complexes in Fennoscandia are indicated from seismic data. Palaeomagnetic data show that both Fennoscandia and the Ukrainian shields were located at low latitudes when the anorthosite–rapakivi rocks intruded at 1.65–1.5 and 1.77–1.72 Ga, respectively, and that the latitudinal positions changed during the time of intrusion. If a hot spot is the origin of the intrusions, it means that it has not been stationary. Underplating in relation with intra-plate tectonic is another possible source of the anorogenic intrusions. However, non of these models can yet satisfactory explain the formation of these anorthosite–rapakivi intrusions.

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