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Notes

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Abstract: We present new geochronological and palaeomagnetic results from the late Palaeoproterozoic Akitkan Group in South Siberia. The zircon U–Pb conventional age of the rhyodacite from the upper part of the group is 1863 $\pm$ 9 Ma and the age of the dacite from the lower part of the group is 1878 $\pm$ 4 Ma. Palaeomagnetic study of sedimentary and some igneous rocks from the upper part of the group isolated a high-temperature characteristic component ($D = 193°$, $I = 19°$, $k = 51$, $\alpha_{95} = 7°$) which is supported by two of three applied conglomerate tests. However, the third intra-formational conglomerate test demonstrates a contaminating overprint of uncertain nature for a part of our collection. The analysis of data suggests that this overprint occurred at time when the geomagnetic field's direction was similar to that at the time of the deposition. Therefore the corresponding palaeomagnetic pole (22.5 °S, 97.4 °E, $dp = 1.5°$, $dm = 2.8°$) may be considered as representative for the deposition time. Palaeomagnetic study of the sediments in the lower part of the Akitkan Group isolated a stable primary remanence ($D = 189°$, $I = 8°$, $k = 111$, $\alpha_{95} = 5°$) supported by positive intra-formational conglomerate and fold tests. The palaeomagnetic pole (30.8 °S, 98.7 °E, $dp = 2.5°$, $dm = 5.0°$) is nearly coeval with the 1879 Ma Molson B pole from the Superior craton. We used these two poles to compare the relative position of Siberia and the Superior craton in the late Palaeoproterozoic. It is different from their reconstruction around 1000 Ma. This demonstrates their relative movements in the Mesoproterozoic.

Pre-Neoproterozoic history of the Earth has been a matter of increased interest in the last decade. Conceptions vary from denying the very existence of plate tectonics before 1 Ga (Stern 2005), to considering a single supercontinent during the Precambrian (Piper 2000), to suggesting assemblages and breakups of one or more supercontinents (Borukaev 1985; Khain & Bozhko 1988; Hoffman

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1991; Windley 1995; Rogers 1996; Dalziel 1997; Condie 2002; Rogers & Santosh 2002). The variety of opinions results (at least in part) from a paucity of pre-1000 Ma palaeomagnetic data. The latest version of the IAGA Global Palaeomagnetic Database (GPMDB, Pisarevsky 2005) contains about 1100 pre-Neoproterozoic results compared to about 8000 Phanerozoic poles. Moreover, a great majority of these data fail to pass reasonable reliability criteria, because many palaeomagnetic poles are very poorly dated and only a few are supported by field tests. In addition, many results were obtained without proper demagnetizing. As a result, there are only 45 very reliable pre-Neoproterozoic poles (Evans & Pisarevsky 2008).

Until recently, Precambrian drift of Siberia was poorly constrained due to lack of precise geochronology and few reliable palaeomagnetic data. For example, at least six different Siberia–Laurentia reconstructions in the youngest Precambrian supercontinent Rodinia (equally poorly constrained) have been published (Pisarevsky *et al.* 2003, 2007; Pisarevsky & Natapov 2003). This situation improved greatly due to recent palaeomagnetic (Gallet *et al.* 2000; Pavlov *et al.* 2000, 2002) and geochronological (Rainbird *et al.* 1998; Sklyarov *et al.* 2003; Gladkochub *et al.* 2006*a*) studies. However, the Siberian role in pre-Rodinian palaeogeography is still practically unknown. There are only three relatively reliable (i.e. well-dated and properly demagnetized) Siberian palaeopoles of a pre-Rodinian age, namely 1503 Ma Kuonamka dykes, 1384 Ma Chieress dykes (Ernst *et al.* 2000) and 1850 Ma Shumikhin granites (Didenko *et al.* 2003, 2005). Moreover, Kuonamka and Chieress poles, as admitted by Ernst *et al.* (2000), are based on a small number of dykes, so the palaeosecular variation is not adequately averaged. Hence there are no reliable (*s.s.*) Siberian palaeopoles for the 800 Ma time interval between the *c.* 1045 Ma Malgina pole of Gallet *et al.* (2000) and the 1850 Ma Shumikhin pole of Didenko *et al.* (2003, 2005). As a result, most of pre-Rodinian global palaeogeographic reconstructions (Rogers 1996; Rogers & Santosh 2002; Zhao *et al.* 2002) place Siberia in the position next to north Laurentia by analogy with early Rodinian reconstructions (Hoffman 1991), or based on vague geological suggestions, some of which were made many years ago and have since proved to be incorrect. On the other hand, Siberia is very important for the pre-Rodinian palaeogeography, because it was almost surrounded by Mesoproterozoic passive margins (Pisarevsky & Natapov 2003; Gladkochub *et al.* 2006*b*), which means that Siberia could be a core of some late Palaeoproterozoic–Mesoproterozoic supercontinent in the same way as Laurentia apparently was the core of Rodinia (Dalziel 1997). Consequently, the need of precisely dated and highly reliable Siberian Meso- and Palaeoproterozoic palaeomagnetic poles is obvious. Here we present results of palaeomagnetic and geochronological studies of two sections of the late Palaeoproterozoic Akitkan Group in South Siberia.

Geology and sampling

The Siberian craton (Fig. 1a) is a Palaeoproterozoic collage of mostly Archaean superterranes, assembled at 2.1–1.8 Ga (Rosen *et al.* 2005). Collision of two major superterranes, Aldan and Anabar, resulted in the formation of the Akitkan orogenic belt (Fig. 1a; Rosen *et al.* 1994; Condie & Rosen 1994; Rosen *et al.* 2005) at *c.* 1.88 Ga (Poller *et al.* 2005). Post-collisional extension caused emplacement of voluminous 1.87–1.84 Ga granites (Neymark *et al.* 1991; Donskaya *et al.* 2002; Larin *et al.* 2003; Poller *et al.* 2005) and a development of the 550 km-long and 60 km-wide North-Baikal volcano-plutonic belt (Fig. 1b; Larin *et al.* 2003; Donskaya *et al.* 2005) composed of *c.* 4500 m of volcanic and volcano-sedimentary strata of the Akitkan Group and their co-magmatic granitoids (Bukharov 1987). Red sediments of the Akitkan Group were the primary target for our palaeomagnetic study.

Mazukabzov *et al.* (2006) proposed a new stratigraphic scheme of the Akitkan Group, based on the synthesis of previous schemes (Mats 1965; Salop 1967; Mats *et al.* 1968; Bukharov 1987) and on the new geochronological data (Larin *et al.* 2003; Poller *et al.* 2005). According to this scheme, the Akitkan Group unconformably overlies the Archaean foliated granites and schists and metavolcanicas of the Palaeoproterozoic Sarma Group in the southern part of the belt (Donskaya *et al.* 2005, 2007). The Sarma Group is cut by the 1910 Ma Kocherikovo granite (Bibikova *et al.* 1987). Direct contacts of the Akitkan Group with the Archaean basement and with the Sarma Group have been observed near the Malaya Kosa cape on the shores of Lake Baikal (Fig. 1d).

The lower part of the Akitkan Group is composed mostly of the clastic Malaya Kosa suite overlain by felsic volcanics and minor sediments of the Khibelen suite (Fig. 1d). In the northern part of the North-Baikal belt, the Akitkan Group is underlain by Palaeoproterozoic metamorphic rocks and granitoids. The Khibelen suite is also exposed here (Fig. 1c) and is conformably overlain by the volcano-sedimentary upper part of the Akitkan Group (Chaya suite). In the study area the Akitkan Group is conformably overlain by the sandstones of the Okun Group, which is, in turn, unconformably overlain by the Neoproterozoic Baikal Group. The Baikal Group directly overlies the Akitkan Group

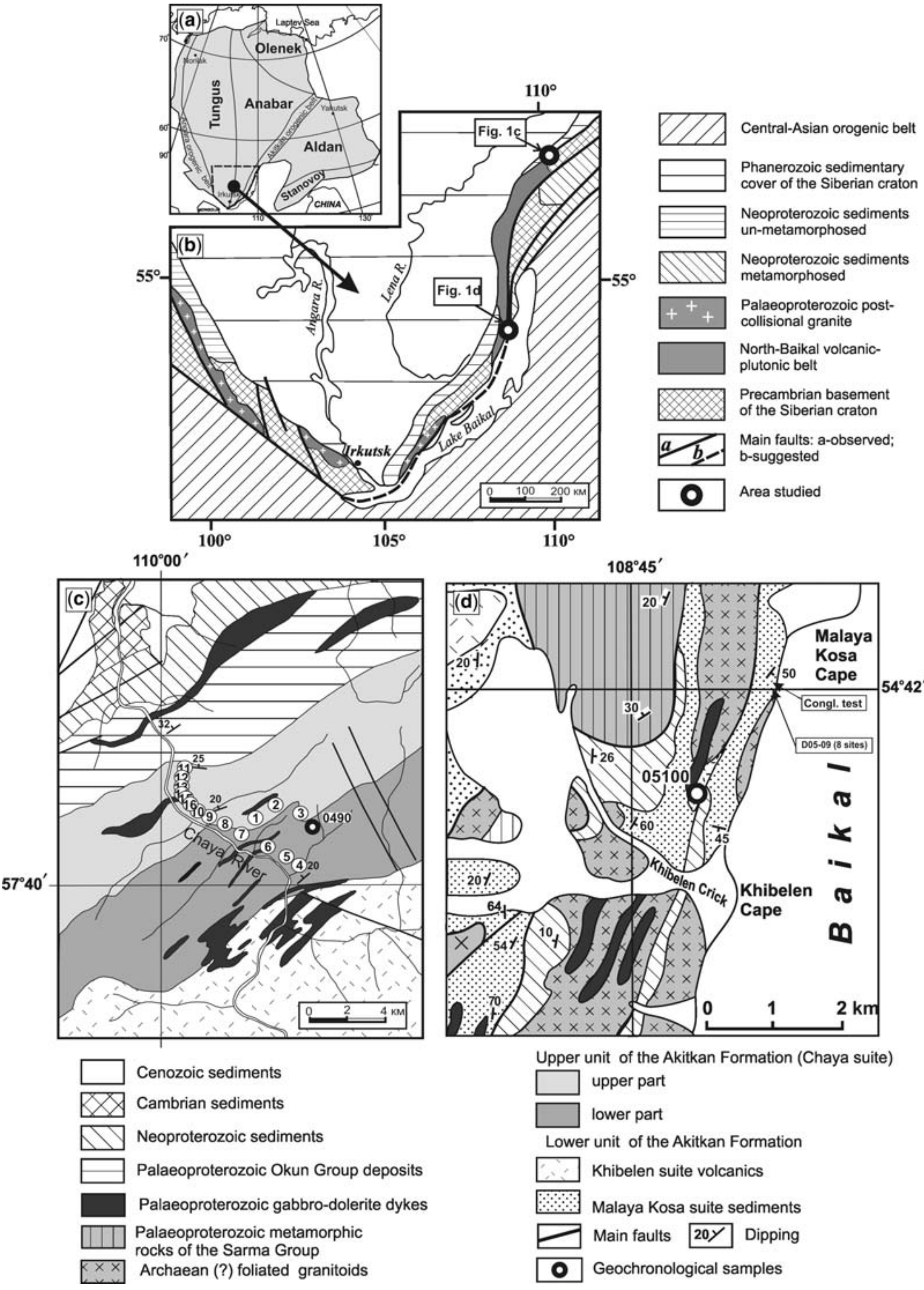

Fig. 1. Geographic setting and geological sketch map of the study area: (**a**) major tectonic elements of the Siberian craton (after Pisarevsky *et al.* 2007); (**b**) geology of the southern part of the Siberian craton (modified after Donskaya *et al.* 2007); (**c**) geological map of Chaya area (northern part of the Akitkan orogenic belt), the prefix D04 is omitted in the numbers of sample sites; (**d**) geological map of the Khibelen area (southern part of the Akitkan orogenic belt).

in most other areas. All of these rocks are weakly metamorphosed up to a low-temperature greenschist grade. The whole stratigraphy is folded; the age of this folding has been considered as Riphean (Meso- to Neoproterozoic) until recently. However, Zorin *et al.* (2008) reported 420 Ma syn-kinematic granites in the Chaya river section, so it is more likely that the pulses of folding continued at least until the Silurian.

The age of the Akitkan Group is constrained by the U–Pb zircon dates of 1866 $\pm$ 6 Ma (Neymark *et al.* 1991) and 1869 $\pm$ 6 Ma (Larin *et al.* 2003) from two distinct horizons of felsic volcanics in the lower to middle part of the Akitkan Group. The 1854 $\pm$ 5 Ma U–Pb zircon age on rhyolite of the upper part of the Akitkan Group (Larin *et al.* 2003) is probably close to the end of its deposition. The whole Akitkan Group (and the Okun Group in one of our sampling localities) is intruded by thick gabbro-diabase intrusions of the Chaya complex with the Sm–Nd age of 1674 $\pm$ 29 Ma (Fig. 1c; Gladkochub *et al.* 2007). The correlatives of these intrusions with similar mineral composition, geochemistry, isotope data and field relationship (Gladkochub *et al.* 2007) exist near the Khibelen Cape in the southern part of the belt (Fig. 1d).

The North-Baikal volcano-plutonic belt is a typical post-collisional belt formed under extension. The collision stage within the 'Akitkan orogenic belt' is marked by the Kaltygey Cape granulites (Poller *et al.* 2005) and the 1910 Ma Kocherikovo granite (Bibikova *et al.* 1987). All of these collision-related complexes are uncomfortably overlain by unmetamorphosed volcanics and sediments of the post-collisional North-Baikal volcano-plutonic belt (the Akitkan Group).

We chose two areas for our study: outcrops along the Chaya river in the northern part of the North-Baikal belt (Fig. 1c) and coastal outcrops near the Malaya Kosa Cape of Baikal Lake in the southern part of the belt (Fig. 1d).

In the first area, felsic volcanics of the lower part of the Akitkan Group (Khibelen suite) are conformably overlain by volcano-sedimentary rocks of the upper part of the Akitkan Group (Chaya suite). The latter was our primary target. It is gently (up to 30°) dipping to the NW and has a combined thickness of *c.* 3100 m (Fig. 1c). The Chaya suite in its lower part consists of grey-green and reddish sandstones with thin layers of tuffs, tuffites, siltstones, conglomerates and lenses of felsic volcanics. Typical thickness of this part is about 1600 m. The *c.* 1500 m upper part of the suite is composed of violet, reddish and green arkoses, polymictic sandstones, and conglomerates. Minor rhyolite flows and small bodies of porphyric andesites also exist. We collected 128 oriented block samples in both parts of the suite from 9 sedimentary sites (cherry-red siltstones and silty sandstones) and from two igneous units of rhyolites (site D04-12) and porphyric andesites (site D04-13). Two sets of pebbles from conglomerate layers (sites D04-3 and D04-11, Fig. 1c) were collected for the conglomerate test. These pebbles were derived from felsic volcanics of the underlying Khibelen suite and/or from the sedimentary rocks of the Chaya suite itself (intra-formational conglomerates). The geochronological sample 0490 has been collected from the rhyodacite flow in the middle part of the section (Fig. 1c).

In the second locality (Fig. 1d), we collected 41 oriented block samples from 8 sites of the *c.* 160 m-thick cherry-red and grey siltstones and fine-grained sandstones of the lower part of the Akitkan Group (Malaya Kosa suite) near its direct contact with Archaean tonalite, 3.3 km south of the Malaya Cosa Cape. These sediments are underlain by a 7 m-thick basal conglomerate and are interbedded with layers of tuffs, tuffites and intra-formational conglomerates. Twenty-nine oriented pebbles from one of these intra-formational conglomerates were collected for the conglomerate test. This is the southeastern limb of an anticline; the strata are dipping to the SE with angles of dip varying from 40°–50°. The geochronological sample 05100 has been collected from felsic volcanics of the overlying Khibelen suite (Fig. 1d).

A magnetic compass was used for orientation; declinations measured directly against the outcrop and at *c.* 1 m away gave consistent results, suggesting that no local magnetic anomalies are present to bias our results (e.g. lightning or strong outcrop magnetizations). Two to four cubic specimens with sides of 2 cm were trimmed from each oriented block sample.

Methods and techniques

Samples were analysed in the palaeomagnetic laboratories at the Geological Institute (GIN RAS, Moscow), at the Institute of Physics of Solid Earth (IFZ RAS, Moscow) and at the Geophysical Observatory (GO, Borok, Russia). Remanence composition was determined by detailed stepwise thermal demagnetization (12–18 steps, to 680 °C), using a TD-48 furnace (ASC Scientific) and home-made furnaces (residual field of *c.* 10–20 nT), and the JR-4 (AGICO) spinner-magnetometer. To monitor possible mineralogical changes during heating and for the magnetic fabric study, magnetic susceptibility was measured in selected samples after each heating step using a KLY-2 (AGICO) kappa-bridge. Magnetic mineralogy was investigated from detailed spectra of the unblocking temperatures and Curie temperatures obtained using the

Curie balance and the 2-component thermomagnetometer of GO Borok (IFZ RAS). Magnetization vectors were isolated using Principal Component Analysis (Kirschvink 1980).

The U–Pb isotope study has been performed in the geochronological laboratory of the Vernadsky Institute of Geochemistry and Analytical Chemistry (RAS, Moscow) by conventional methods. Decomposition of zircon microsamples (0.5–0.1 mg) and chemical separation of U and Pb for isotopic analyses has been done using the method of Krogh (1973). U and Pb concentrations were determined by the isotopic delution method using a mixed spike $^{208}Pb + ^{235}U$, blanks being 0.1 ng of Pb and 0.005 ng of U. Isotopic compositions were measured on a multi-collector mass spectrometer TRITON. The ISOPLOT program (Ludwig 1999) was used for processing of experimental data. Common-lead correction was introduced for the age 1850 Ma by the model of Stacey & Kramers (1975). The precision in U–Pb isotopic ratios was 0.5%. All errors are given at the 2σ level. Preliminary selective decomposition (SD) was undertaken for some zircon fractions (Mattinson 1994) to improve the concordance of U–Pb ratios in zircons. According to this method, the preliminary treatment of zircons was done in concentrate HF for 8 hours at 150 °C. After that, the crystalline residue was treated twice with 3.1 N HCl at 180 and at 200 °C during 6 and 10 hours. The crystalline residue was washed twice with 1N HNO_3 and decomposition has been done conventionally.

Upper part of the Akitkan Group

Geochronology. The geochronological sample 0490 has been collected from the rhyodacite lens within coarse sediments of the lower part of the Chaya suite in the River Suslinka (right Chaya's tributary, Fig. 1c). The rock composition corresponds to that of quartz-porphyry. Elements of fluidal texture are locally visible. Geochronological U–Pb analyses were done for two grain size fractions of the accessory zircon and for the residue after selective decomposition (Mattinson 1994). One of the points (+100 μm, Table 1) was not used for discordia calculations. The lower intercept of discordia reflecting the time of lead loss is 154 ± 170 Ma. The upper intercept age is 1863.2 ± 8.7 Ma (Fig. 2a). The morphology of zircon is magmatic and permits us to regard the age of 1863 ± 9 Ma as the time of crystallization of the melt parent for rhyodacites. This age is similar to the earlier reported result of 1854 ± 5 Ma by Larin *et al.* (2003) from another outcrop of felsic volcanics of the upper part of the Akitkan Group. We conclude that the age of the upper part of the Akitkan Group is close to 1860 Ma.

Magnetic minerals and rock magnetism. The natural remanent magnetization (NRM) of red sediments ranges from $0.8–5.0 \times 10^{-5}$ A m^{-1}, and their magnetic susceptibility from about $1–3 \times 10^{-4}$ SI units. The ratio of remanent magnetization to induced magnetization (Koenigsberger's Q parameter) usually reflects the grain size distribution of the magnetic material and, therefore, how magnetically 'hard' or 'soft' the particles are. Q varies from 2–5 (Fig. 3a) for the studied red beds, indicating their high palaeomagnetic stability, which is typical for most red beds (Khramov 1987). NRM of igneous rocks vary from $0.9–1.0 \times 10^{-5}$ A m^{-1}, and their magnetic susceptibility from about $1–7 \times 10^{-4}$ SI units. Q varies mostly from 2–5 (Fig. 3a), but two rhyolite samples have Q close to 1.

Magnetic saturation v. temperature (J_s-T) curves are mainly of Q- and R-types (see fig. 1–24 of Nagata 1961). The J_s-T curves (both the first and the second heating) of the porphyritic andesite (Fig. 3b) indicate two bends corresponding to Curie temperatures of 560 °C and 670 °C. Saturation remanence (J_{rs}-T) curves for the same rock indicate two blocking temperatures of 560 °C and 660 °C (Fig. 3b). These data suggest that the studied porphyritic andesite crystallized in subaerial conditions producing low-titanium titanomagnetite. Its deuteric oxidation during initial cooling caused generation of magnetite and hematite (Ade-Hall *et al.* 1971; Pechersky *et al.* 1975; McElhinny & McFadden 2000).

Both J_s-T and J_{rs}-T curves for red beds indicate that hematite carry the bulk of their remanence, but some minor magnetite is also indicated by small bends near 560–570 °C on the J_s-T curves (Fig. 3c).

Palaeomagnetism. Some typical examples of thermal demagnetization are shown in Figure 4. The majority of samples exhibit an interpretable demagnetization plot with the exception of the weak and unstable rhyolites (site D04-12). The present-day field (PDF) has been destroyed in most samples at 200–300 °C. In the rest of the samples this component is practically absent (Fig. 4b, d, e, f).

The shallow SSW high-temperature component was isolated in almost all samples and in some cases this is the only component carried by both magnetite and hematite (Fig. 4a, f, h). However, in some cases the hematite is its only carrier (Fig. 4b, c, d). Note that the direction of the high-temperature component and the thermal demagnetization behaviour is similar in porphyritic andesites (Fig. 4e, f) and in sedimentary rocks (other diagrams in Fig. 4). Stereoplots of the high-temperature component are shown in Figure 4i and j (sample

Table 1. *U–Pb isotopic data for zircons from sample No. 0490 (dacite-porphyry from Upper unit of the Akitkan Formation) and from sample No. 05100 (dacite from Lower unit of the Akitkan Formation)*

No.	Fraction (μm)	Weight (g)	Contents (ppm)		Isotopic composition of Pb			Isotopic ratio and age, Ma		
			U	Pb	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{207}Pb$	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
				Sample 0490 (rhyodacite from the upper part of the Akitkan Formation, Chaya River area)						
1	+100	0.00102	381.47	125.86	3170	8.6038	5.9930	0.2976	4.5979	1833.1 ± 1.0
2	−100 + 75	0.00100	258.13	81.20	7551	8.7110	6.1440	0.2860	4.4607	1849.9 ± 1.0
3	−75	0.00760	471.05	177.09	8170	8.7360	6.1590	0.3321	5.2200	1846.8 ± 1.0
4	+75, SD				2866	8.4660	6.6340	0.3191	4.9939	1856.4 ± 1.1
5	+75, SD				18710	8.7525	7.0203	0.3144	4.9262	1858.2 ± 1.0
				Sample 05100 (dacite from lower part of the Akitkan Formation, the Khibelen Cape area)						
1	+100	0.00114	218.59	82.43	1416	7.3264	2.8220	0.2911	5.1093	2060.8 ± 1.0
2	+75	0.00110	258.29	91.02	534	7.2304	2.2207	0.2504	3.9077	1850.9 ± 1.2
3	−75	0.00100	225.50	82.96	1050	7.9031	2.4694	0.2755	4.3238	1861.0 ± 2.0
4	−75, SD				4610	8.4841	3.1101	0.3209	5.0358	1861.2 ± 0.9

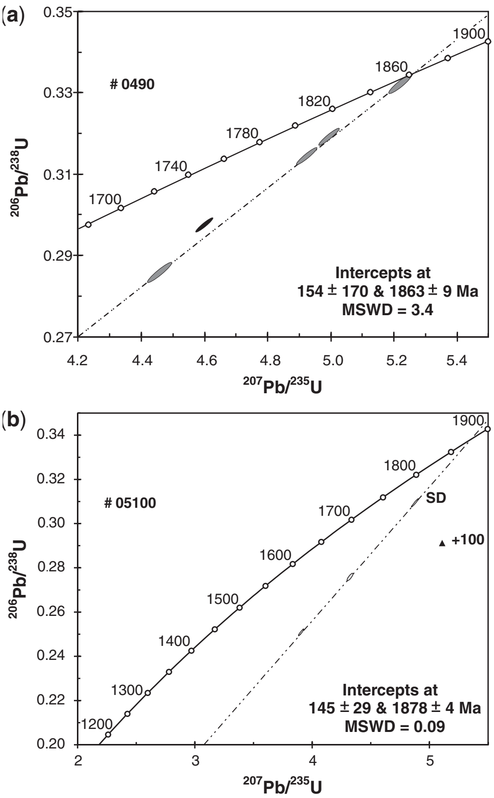

Fig. 2. U–Pb Concordia diagram and analysis of zircon crystal from (**a**) rhyodacite from Upper unit of the Akitkan Formation and (**b**) dacite from Lower unit of the Akitkan Formation.

and site means correspondingly). Table 2 shows the corresponding sample and site means and statistics, palaeomagnetic pole calculate from means after the tilt correction. The fold test of Enkin (2003) is indeterminate. Interestingly, our result (D = 192.5°, I = 18.6°, k = 50.5, α_{95} = 6.9°) is close to the old result of Gurevich from the same rocks (D = 191°, I = 14°, k = 30, α_{95} = 2°) based on a bulk demagnetization of samples without vector diagrams (Gurevich 1982, in Pisarevsky 2005, entry 5595). This proves once more that not all old results are bad results. Importantly, Gurevich reported the presence of both polarities.

In the medium-temperature interval (somewhere between 300 and 580 °C) many Zijderveld plots show some evidence of the possible presence of an additional remanence component (Fig. 4b, c, d, g). In some cases this 'medium-temperature component' may be artificial due to the overlapping of the blocking temperature spectra of the low- and high-temperature components. However, it is probably real in 22 samples in the upper part of the section. Its mean direction (after tilt correction) is D = 188°, I = −29°, k = 6.3, α_{95} = 13.5°. This is very close to the remanence directions of 1674 Ma mafic intrusions of the Chaya complex (Vodovozov *et al.* pers. comm.).

The medium-temperature component was also found in 29 samples of the lower part of the section, but it is very scattered there. However, suggesting the same direction of remagnetization (D = 188°, I = −29°) we applied the test for a random distribution of Shipunov *et al.* (1998). It revealed $\rho = 0.244$ ($\rho = n^{-1}\Sigma \cos \phi_i$, where n is the number of samples and ϕ_i is the angle between the ith unit vector and the suggested direction of the remagnetization). The critical value $\rho_c = 0.176$ is less than the dataset value of 0.244, so the partial influence of a 1674 Ma remagnetization is suggested.

Conglomerate test. We collected oriented pebbles from two conglomerate layers (sites D04-3 and D04-11, Fig. 1c) to constrain the age of our characteristic remanence by the conglomerate test of Graham (1949). The first set of pebbles includes 17 porphyritic clasts of the underlying Khibelen suite (circles in Fig. 5c, e). Their thermal demagnetization in most cases isolated just one high-temperature remanence component (Fig. 5a) with a chaotic direction (Fig. 5e), and a few samples also contain a low-temperature component with a PDF direction (Fig. 5c). A random distribution of the high-temperature component is indicated with $r/r_c = 0.161/0.388$, where $r = R/n$ and r_c is its critical value for n = 17 at the 95% confidence level (Mardia 1972). This confidence level has been chosen for all tests mentioned hereafter. The conglomerate test of Shipunov *et al.* (1998), with suggested direction of remagnetization of D = 192.5°, I = 18.6° (our site mean direction for the Chaya suite) is also positive: $\rho/\rho_c = 0.112/0.230$.

Seven pebbles from the same set (site D04-3) are sandstones and siltstones of the Chaya suite (triangles in Fig. 5e). Their thermal demagnetization also revealed a randomly distributed high-temperature component, according to both Rayleigh ($r/r_c = 0.371/0.597$) and Shipunov's ($\rho/\rho_c = 0.141/0.362$) tests. Hence this intra-formational conglomerate test is also positive.

Twenty-nine pebbles of siltstones and fine sandstones of the Chaya suite were collected in site D04-11 (Fig. 1c). Most of their thermal demagnetization plots show low- and high-temperature components (Fig. 5b). The low-temperature component is generally distributed around the PDF (Fig. 5d), probably being a recent viscous remanence. The distribution of the high-temperature component

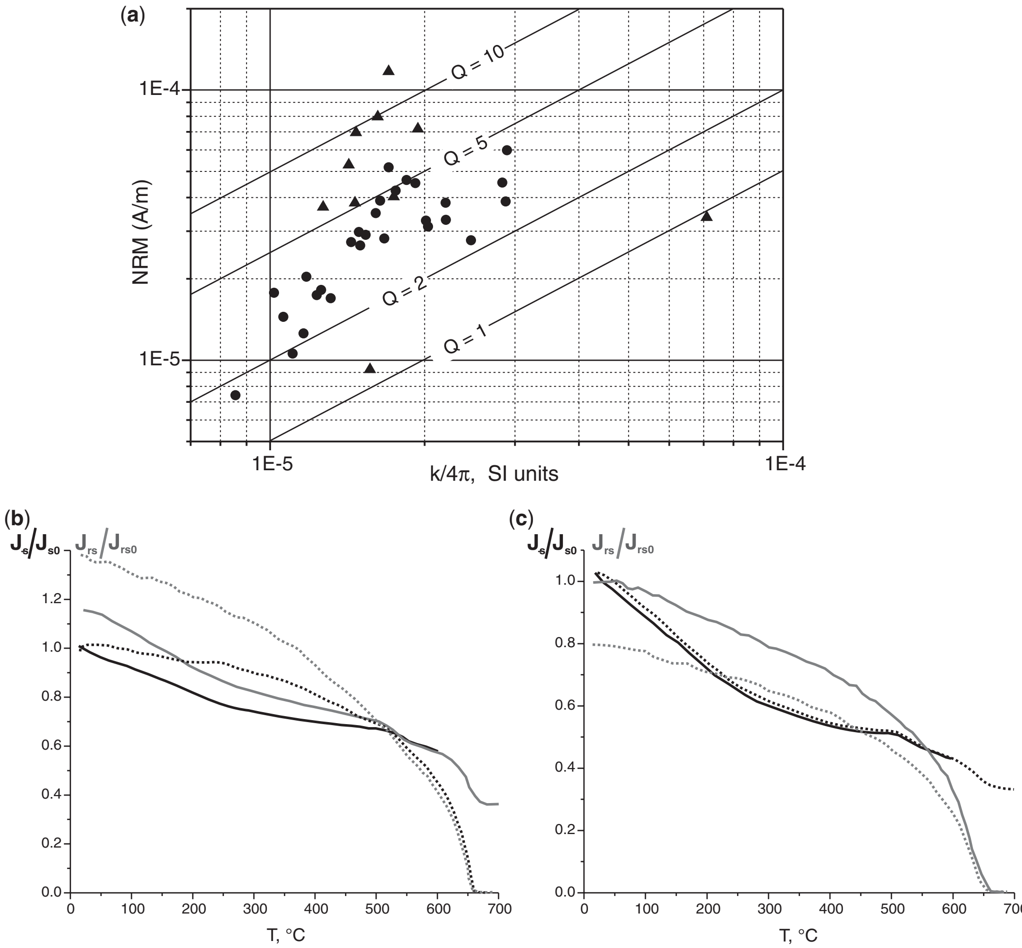

Fig. 3. Rock magnetism of the upper part of the Akitkan Group: (**a**) remanence-susceptibility relationships for igneous (triangles) and sedimentary (circles) rocks; (**b–c**) J_s-T (black) and J_{rs}-T (grey) curves (dotted lines – second heating): (b) site D04-13 (porphyric andesite); (c) site D04-15 (siltstone.)

(Fig. 5f) is not entirely random. The Rayleigh uniformity test (Mardia 1972) is negative ($r/r_c = 0.499/0.296$) and the mean direction (D = 172°, I = 20°, k = 1.93, $\alpha_{95} = 27.4°$, after the tilt correction) is close to the characteristic remanence direction for the Chaya suite (Table 2).

Thus we have three conglomerate tests in the Chaya suite. Two of them, including one intra-formational test (site D04-3) are positive, but the second intra-formational test (site D04-11) is negative. This probably means that the characteristic remanence of igneous and some sedimentary rocks of the Chaya suite is primary, but that of some other sediments may be partly remagnetized. However, there is no systematic difference between the remanence directions of igneous and sedimentary parts of our collection (Table 2, Fig. 4j). In addition, the calculated palaeopole (Table 2) is close to the *c*. 1850 Ma Shumikhin granites pole of Didenko *et al.* (2003, 2005) that is supported by the presence of two polarities (Table 4, Fig. 9) and to our *c*. 1880 Ma Malaya Kosa pole that proved to be primary by positive conglomerate and fold tests (see below). This puzzle could be explained if some minor remagnetization event occurred after the deposition of the Chaya suite and the direction of this overprint was similar to the characteristic remanence direction. The 1674 Ma mafic intrusions intruding the Chaya succession could cause such an overprint. We should also note that our Chaya pole lies close to the late Ordovician–Silurian part of the Siberian APWP (Didenko & Pechersky 1993; Smethurst *et al.* 1998). As mentioned above there is evidence of the Early–Middle Palaeozoic

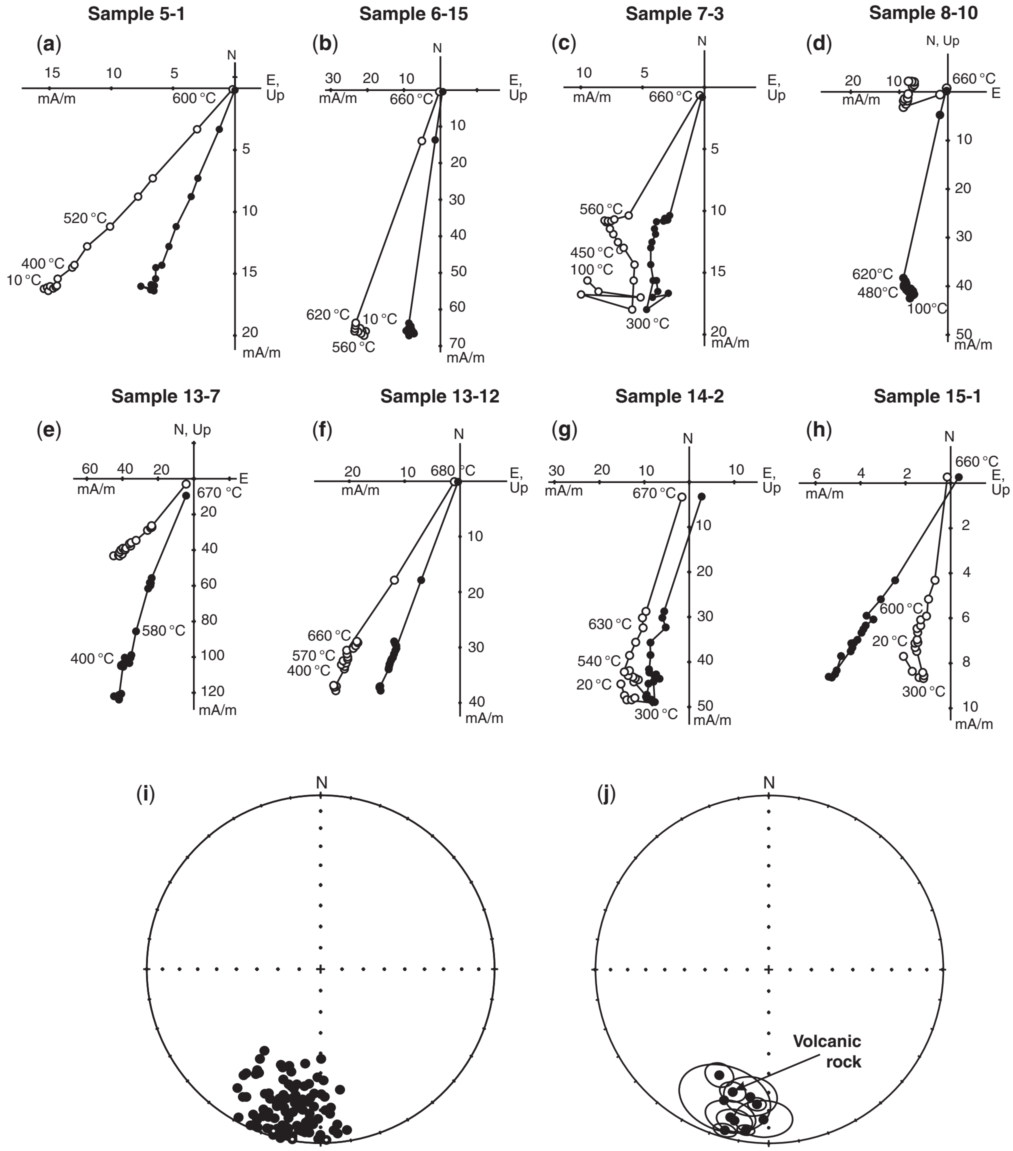

Fig. 4. Thermal demagnetization of the upper part of the Akitkan Group. Orthogonal Zijderveldt plots for: (**a**–**d**, **g**–**h**) sedimentary rocks; (**e**–**f**) igneous rocks (site D04-13); solid and open symbols denote horizontal and vertical projections, respectively. Stereographic projections of the high-temperature component: sample means (**i**) and site means (**j**); solid and open symbols represent downward and upward directions.

tectonic activity in the region and the 420 Ma synkinematic granites have been found in the Chaya river section (Zorin *et al.* 2008), so some Silurian remagnetization is also possible. The nature of remagnetization in both scenarios is not clear, because igneous and some sedimentary rocks were not affected. We think that this rules out a thermal remagnetization, but leaves a possibility for some chemical process selectively influencing some parts of the studied section.

Lower unit of the Akitkan Group

Geochronology. The geochronological sample 05100 has been collected from Khibelen dacite (Fig. 1d). This dacite conformably rests on the

Table 2. *Mean palaeomagnetic directions for high temperature component from the upper part of the Akitkan Formation (Chaya River, 57.6 °N; 110.8 °E)*

Site	N/n	Dip dir./Dip	*In situ*				Tilt corrected				Palaeomagnetic pole				
			D, °	I, °	k	α_{95}, °	D, °	I, °	K	α_{95}, °	Plat, °N	Plong, °E	d_p, °	d_m, °	φ_a, °
D04-5	6/8	348/22	185.3	7.0	40.4	10.7	188.2	26.9	40.4	10.7					
D04-6	15/15	348/28	183.0	−2.7	115.7	3.6	185.0	23.3	115.4	3.6					
D04-7	14/15	345/60	197.2	−14.0	43.3	6.1	205.1	33.4	44.2	6.0					
D04-8	15/15	335/22	188.3	−9.3	150.5	3.1	188.1	7.5	157.0	3.1					
D04-1	6/11	359/42	198.0	−16.4	14.6	18.2	198.9	21.7	14.6	18.2					
D04-2	8/12	327/39	182.0	−14.9	25.7	11.1	182.0	14.8	27.4	10.8					
D04-13	10/12	355/30	193.0	0.4	83.5	5.3	196.4	27.5	83.7	5.3					
D04-14	13/13	340/25	192.3	−9.2	40.6	6.6	192.8	11.8	41.3	6.5					
D04-15	5/5	348/27	194.1	−9.4	67.6	9.4	194.6	13.4	67.6	9.4					
D04-16	16/16	333/25	197.2	−14.9	89.0	4.7	195.3	4.7	86.7	4.0					
Sample mean:	108/128		191.1	−8.8	35.0	2.3	192.5	17.8	26.6	2.7	22.5	277.4	1.5	2.8	9.1
Site mean:	10/11		191.0	−8.4	70.2	5.8	192.5	18.6	50.5	6.9	22.1	277.5	3.7	7.2	9.6

Note: N/n, number of samples or sites used/collected; D/I, mean declination/inclination; k, Fisher's precision parameter; α_{95}, the semi-angle of the 95% cone of confidence; Plat/Plong, latitude and longitude of palaeomagnetic poles; d_p/d_m, semi-axes of the cone of confidence about the pole at the 95%; φ_a, palaeolatitude.

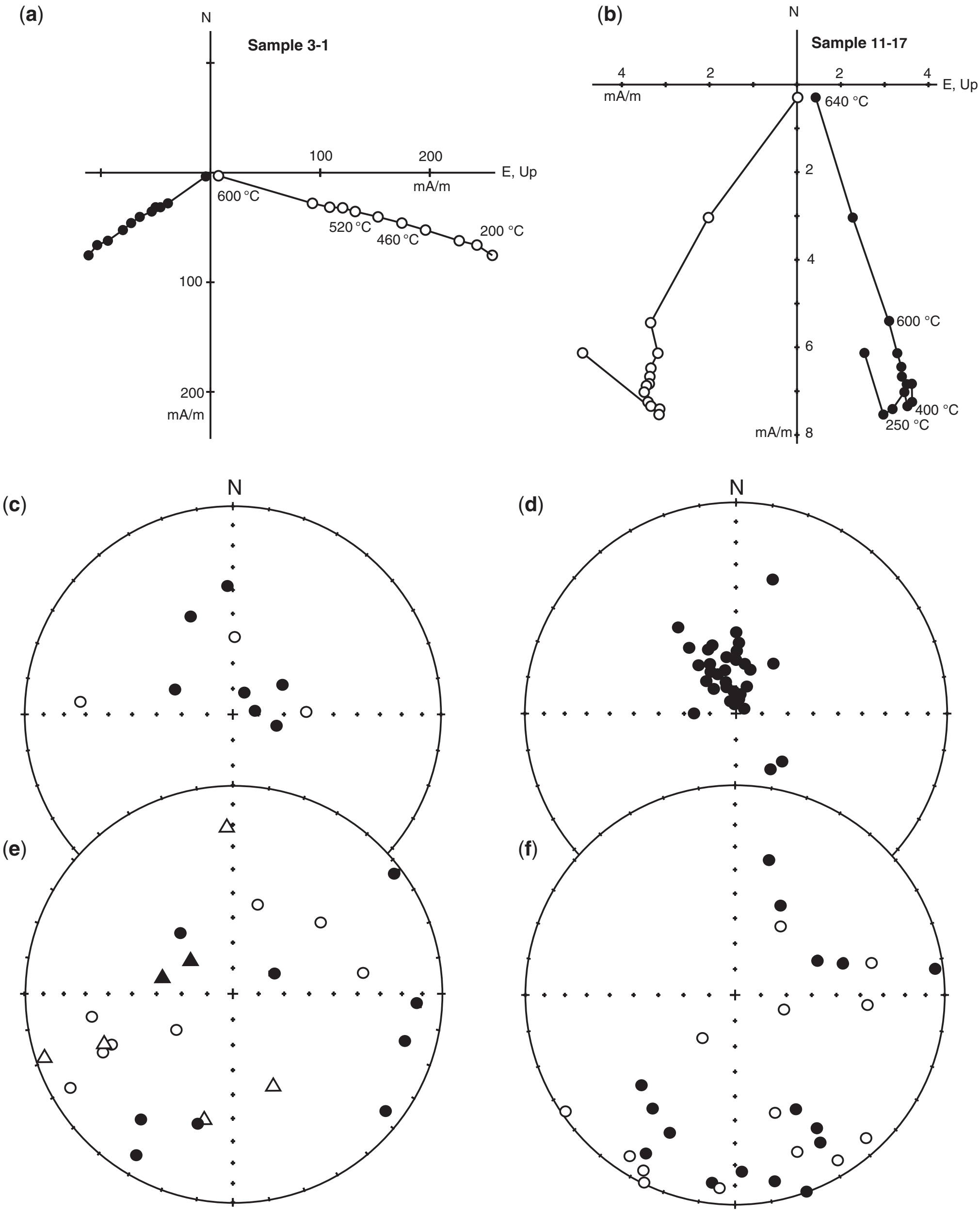

Fig. 5. Conglomerate tests for the upper part of the Akitkan Group. Left column, site D04-3; right column, site D04-11; (**a**–**b**) orthogonal plots; (**c**–**d**) stereoplots for the low-temperature component; (**e**–**f**) stereoplots for the high-temperature component. Triangles denote seven intra-formational pebbles of the Chaya suite. See also caption to Figure 4.

sedimentary sequence (Malaya Kosa suite) sampled for our palaeomagnetic study. The separated accessory zircons are short prismatic semi-transparent yellow crystals with smoothed terminations and sometimes visible inner zonation. The sample for analysis was handpicked avoiding grains containing inclusions or cracks. The U–Pb analyses have been done for three size fractions and one residue after selective decomposition (SD). One point containing cores (Table 1; +100) was not used in age calculations. The lower intercept age of 145 ± 29 Ma points to relatively recent lead loss. The upper intercept age is 1877.7 ± 3.8 Ma (Fig. 2b). Morphology and inner structure of zircons suggest their magmatic genesis, so we interpret the age of 1878 ± 4 Ma as the most reliable estimate for the time of the parent melt crystallization. The age of granulites near the Kaltygey Cape of 1876 ± 6 Ma (Poller *et al.* 2005) is overlapping with our new date. These granulites are overlain by the Akitkan Group, suggesting that its lower part (unmetamorphosed volcanics and sediments) has been accumulated just after the end of the collisional event accompanying the basement metamorphism. According to geological observations, the analysed dacite lies directly on the earliest sediments of the Akitkan Group. The age of this dacite could therefore be interpreted as the earliest volcanic event within the Akitkan Group. The latest volcanic activity of the lower and middle part of the Akitkan Group could be constrained by a U–Pb age of 1866 ± 6 Ma (Neymark *et al.* 1991).

Magnetic minerals and rock magnetism. The natural remanent magnetization (NRM) of red sediments ranges from $0.07–2.0 \times 10^{-5}$ A m^{-1}, and their magnetic susceptibility from about $0.5–2 \times 10^{-5}$ SI units with Q >1 for the most part of the collection (Fig. 6a), indicating high palaeomagnetic stability. The J_s-T curves (both the first and the second heating) indicate Curie temperatures between 660 and 680 °C typical for the hematite

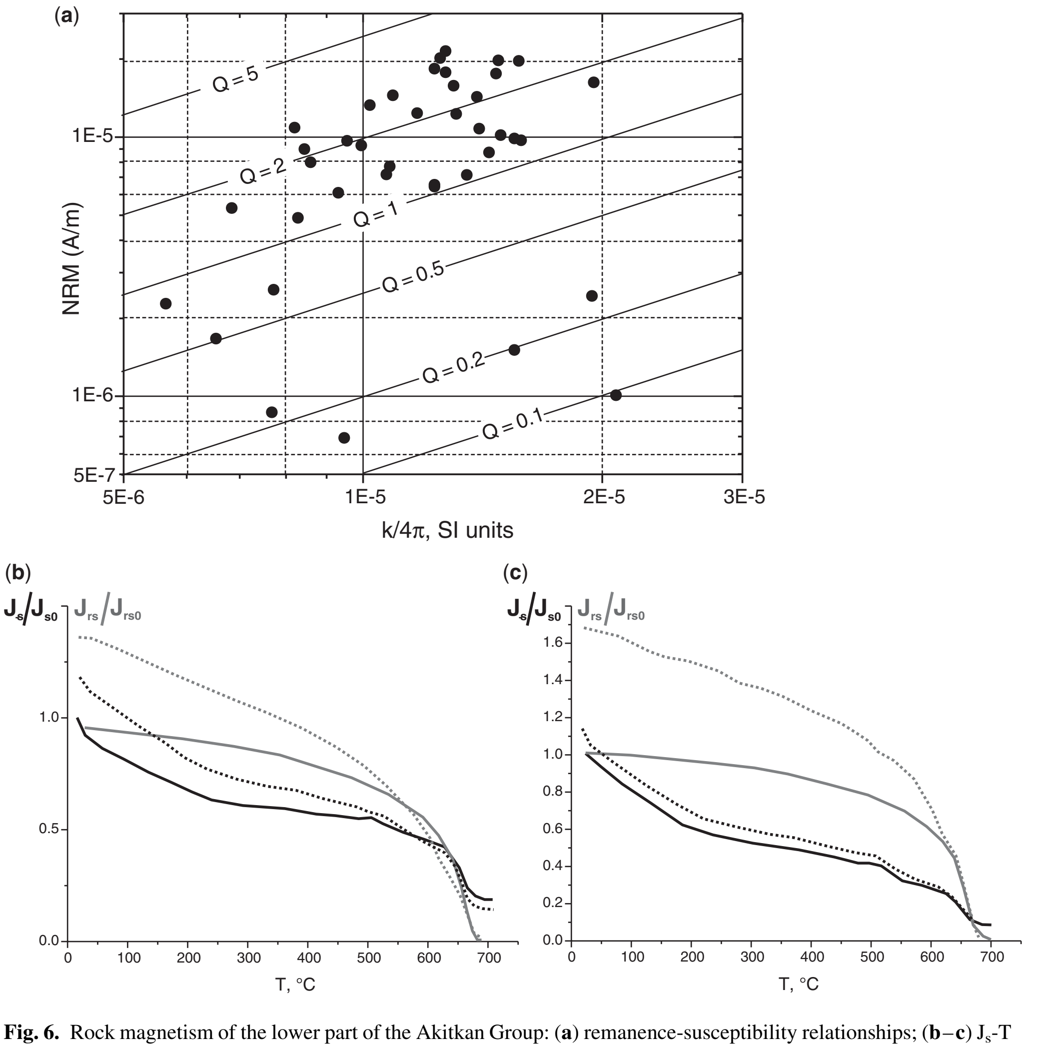

Fig. 6. Rock magnetism of the lower part of the Akitkan Group: (**a**) remanence-susceptibility relationships; (**b–c**) J_s-T (black) and J_{rs}-T (grey) curves (dotted lines–second heating): (b) site D05-9 (siltstone); (c) site D05-9 (fine sandstone).

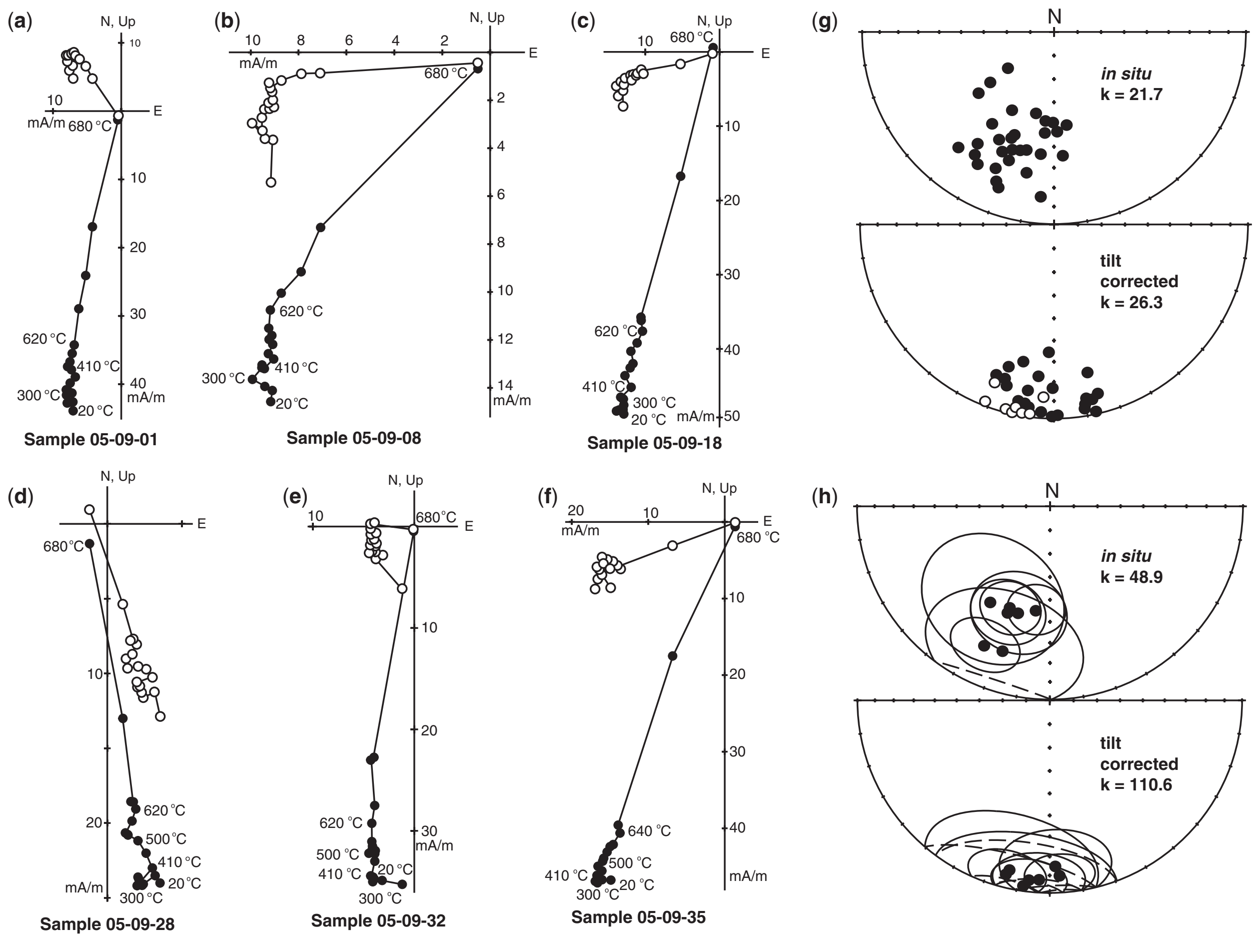

Fig. 7. Thermal demagnetization results of the lower part of the Akitkan Group: (**a**–**f**) orthogonal Zijderveldt plots; (**g**–**h**) stereographic projections of the high temperature component before and after the tilt correction: sample means (g) and site means (h). See caption to Figure 4.

(Fig. 6b, c). J_{rs}-T curves show that unblocking temperatures are mostly between 600 and 680 °C, confirming the dominance of hematite.

Palaeomagnetism. Some typical examples of the thermal demagnetization are shown in Figure 7. The random 'laboratory' viscous component and the present-day field (PDF) component have been destroyed in most samples at 200–300 °C.

The shallow SSW high-temperature hematite component was isolated in all samples (Fig. 7). The stereoplots of the high-temperature component are shown in Figure 7g (sample mean – $D = 188.0°$, $I = 8.2°$, $k = 26.5$, $\alpha_{95} = 5.2°$) and 7 h (site mean – $D = 188.7°$, $I = 8.1°$, $k = 110.6$, $\alpha_{95} = 5.0°$). Table 3 shows the corresponding means, palaeomagnetic pole and statistics.

The medium-temperature component has been isolated between 400 and 520–540 °C (sometimes up to 600 °C). Its direction after tilt correction ($D = 183°$, $I = 19°$, $k = 12$, $\alpha_{95} = 7.5°$) is close to the direction of the high-temperature component in the Chaya suite (Table 2), but it is also close to the characteristic component for the Malaya Kosa suite (Table 3). We suggest that in part it may be artificial due to the overlapping of the blocking temperature spectra of the low- and high-temperature components. However, this component may be related to an overprint (*c.* 1674 Ma or *c.* 420 Ma, see above).

Field tests. Figure 7g and h demonstrate a better grouping of the high-temperature component directions after the tilt correction. The fold test of Enkin (2003) is positive with maximum grouping after 99.2% of unfolding. This test demonstrates that the characteristic high-temperature remanence of the Malaya Kosa sediments is pre-folding, that is at least pre-Silurian, as the latest folding event in the area occurred at *c.* 420 Ma (Zorin *et al.* 2008).

Twenty-five out of twenty-nine oriented pebbles of red sediments collected from the intra-formational conglomerate near the base of the sampled section (Fig. 1d) revealed interpretable thermal demagnetization plots (Fig. 8a, b). The low-temperature component, with unblocking temperatures <350 °C, is distributed near the PDF direction (Fig. 8c), and the Rayleigh uniformity test (Mardia 1972) is negative ($r/r_c = 0.953/0.296$). The high-temperature component has been isolated in 18 pebbles. It is chaotically distributed (Fig. 8d) and the Rayleigh uniformity test (Mardia 1972) is positive ($r/r_c = 0.117/0.377$). Hence we conclude that the characteristic remanence of the Malaya Kosa sediments is primary. As mentioned above, the time of deposition of these sediments was very short, so the 1878 ± 4 Ma age is a good approximation of this new palaeopole.

Table 3. *Mean palaeomagnetic directions for high temperature component from the lower part of the Akitkan Formation (Khibelen Cape, 54.7°N; 108.8°E)*

Site	N/n	Dip dir./Dip	*In situ*				Tilt corrected				Palaeomagnetic pole				
			D, °	I, °	k	α_{95}, °	D, °	I, °	K	α_{95}, °	Plat, °N	Plong, °E	d_p, °	d_m, °	φ_a, °
D05-9(1–5)	5/5	145/50	204.6	40.3	28.3	14.6	186.5	6.9	28.3	14.6					
D05-9(6–11)	5/6	155/40	190.5	44.9	45.6	11.4	179.7	9.7	45.6	11.4					
D05-9(12–17)	4/6	155/47	214.7	41.2	11.1	28.8	196.3	9.8	11.1	28.8					
D05-6(18–23)	6/6	156/47	204.3	42.6	32.3	5.9	189.6	5.9	32.3	5.9					
D05-9(24–28)	4/5	138/42	199.4	41.3	24.8	18.8	181.0	14.5	24.8	18.8					
D05-9(29–34)	3/6	129/48	201.2	22.0	19.8	28.5	191.0	2.2	19.8	28.5					
D05-9(35–38)	3/4	130/43	208.1	21.3	96.6	12.6	196.9	7.7	96.6	12.6					
Sample mean:	30/41		203.0	38.2	21.7	5.8	188.0	8.2	26.5	5.2	30.8	279.5	2.6	5.2	4.1
Site mean:	7/8		203.4	36.4	48.9	7.6	188.7	8.1	110.6	5.0	30.8	278.7	2.5	5.0	4.0

Note: see Table 2.

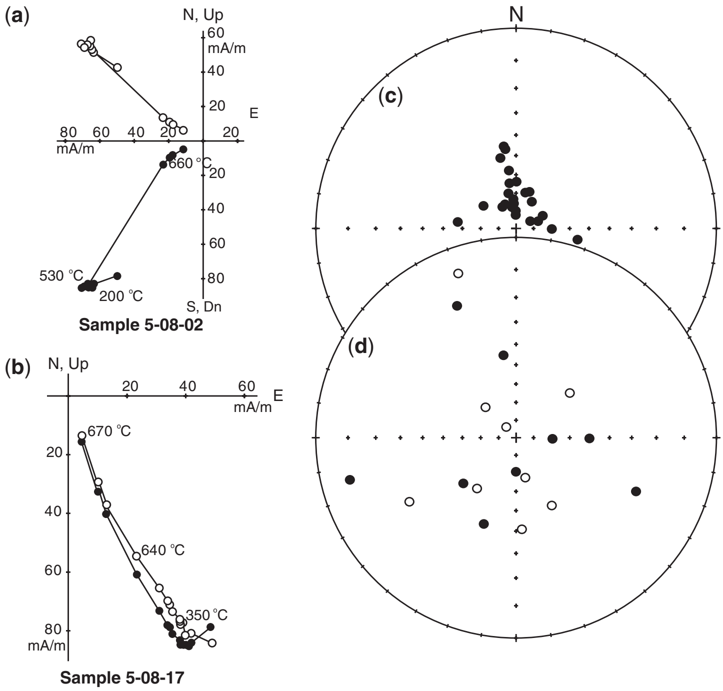

Fig. 8. Conglomerate test for the lower part of the Akitkan Group (Malaya Kosa). (**a**–**b**) orthogonal Zijderveldt plots. Stereoplots: (**c**) low-temperature component; (**d**) high-temperature component. See caption to Figure 4.

Discussion

The results of this study provide two reliable Siberian Palaeoproterozoic palaeopoles. The reliability index Q of Van der Voo (1990) for the 1878 Ma Malaya Kosa pole is 6, but we are more cautious about the Chaya pole due to less convincing conglomerate tests. These two poles, together with the Shumikhin pole (Didenko *et al.* 2003, 2005), provide robust constraints on the palaeopositions of Siberia at 1850–1880 Ma. Lack of reliable Mesoproterozoic palaeopoles prevents a construction of the pre-Neoproterozoic Siberian APWP, and we must consider two polarity options for any reconstruction. However, there are some direct tectonic applications of our new results.

The age of the Malaya Kosa pole is remarkably close to the age of the highly reliable 1880 Ma Molson dykes B-pole (Halls & Heaman 2000). Such closeness in age of two reliable poles from different continents is rare for pre-Neoproterozoic times (Cawood *et al.* 2006; Evans & Pisarevsky 2008). As the latest Mesoproterozoic Laurentia-Siberia reconstruction is palaeomagnetically constrained (Gallet *et al.* 2000; Pavlov *et al.* 2000, 2002; Pisarevsky & Natapov 2003), our study provides an important test for the differential movements of continental blocks in the late Palaeoproterozoic–Mesoproterozoic, challenged by Stern (2005). Stern argued that the modern-style subduction-related plate tectonics did not exist before 1 Ga, as in his opinion such evidence for an ancient subduction as blueschists, high-pressure metamorphics and ophiolite complexes are apparently absent in pre-Neoproterozoic. High-quality palaeomagnetic data are crucial to test this suggestion, because they may indicate the differential movements of different continents. Two pairs of coeval poles from two continental blocks allow a comparison of their mutual positions at two different time slices. Cawood *et al.* (2006) and Evans & Pisarevsky (2008) found only a few examples of such evidence of differential movements between Superior and Kalahari, Baltica and Australia, Superior and Karelia. Now we have an opportunity to test late Palaeoproterozoic–Mesoproterozoic movements between Siberia and Superior/Laurentia. Figure 9a and b demonstrate possible Superior–Siberia reconstructions at 1880 Ma in two polarity options based on the Malaya Kosa and Molson B poles. Multiple images of Siberia for each option highlight longitudinal uncertainty in craton position.

Pisarevsky & Natapov (2003) made a Siberia–Laurentia late Mesoproterozoic reconstruction using a best fit between the *c.* 950–1050 Ma set of

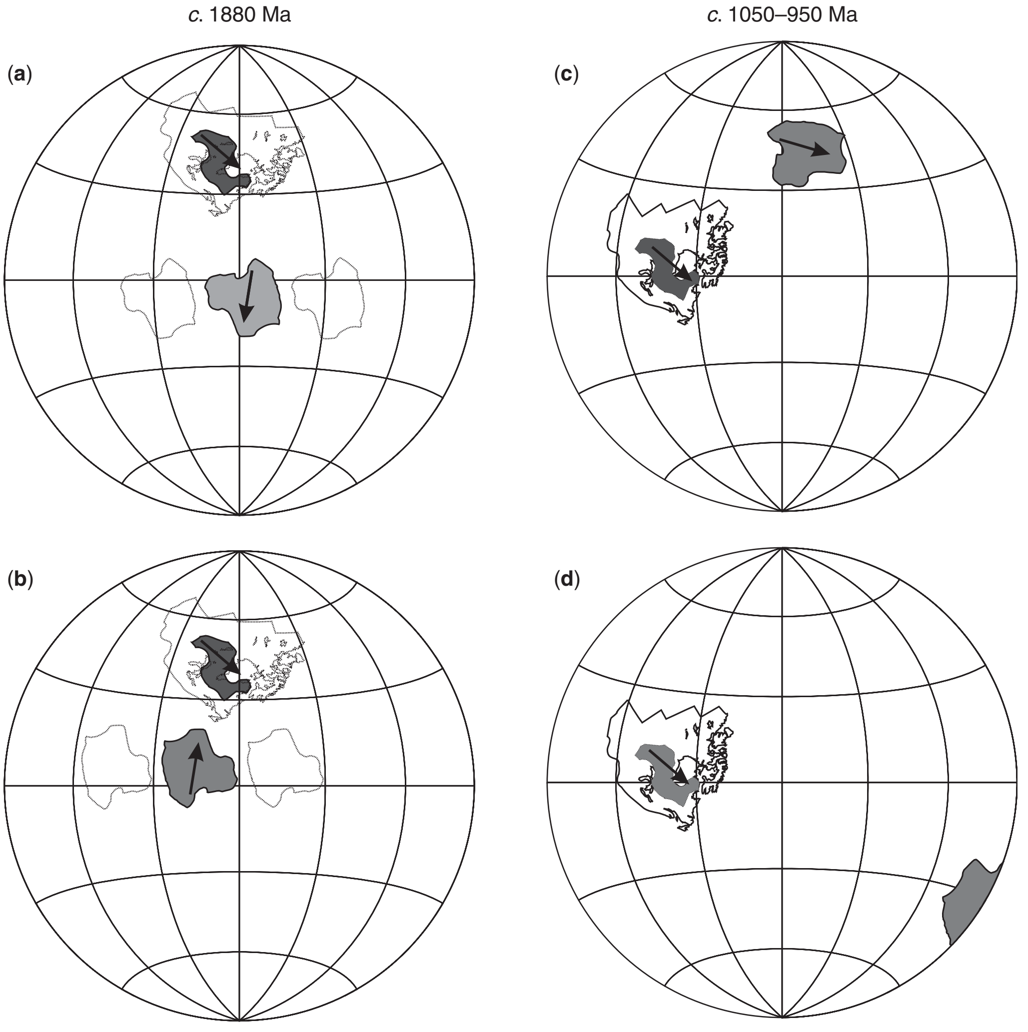

Fig. 9. Superior–Siberia reconstructions in two polarity options at (**a**–**b**) 1880 Ma; (**c**–**d**) 1050–950 Ma (after Pisarevsky & Natapov 2003). The arrows indicate to the modern 'North'.

poles from the Uchur-Maya area (Gallet *et al.* 2000; Pavlov *et al.* 2000, 2002) and the coeval fragment of the Laurentian APWP (Pisarevsky & Natapov 2003 for reference). Two polarity options are shown in Figure 9c and d. In this case we do not have a longitudinal uncertainty, because APWP fragments were used rather than individual poles. Figure 9 demonstrates a significant difference between relative palaeopositions of Superior/Laurentia and Siberia between *c*. 1880 and *c*. 1000 Ma, suggesting independent drift of two continents in the Mesoproterozoic, which is possible only in the present-day style tectonic environments with generation and consumption of lithosphere between these blocks on a constant-radius Earth. However, this should be considered just as a reasonable suggestion, because most of 950–1050 Ma Siberian and Laurentian poles cannot be considered as 'key-poles' for various reasons, mainly their poor age constraints (Evans & Pisarevsky 2008).

Conclusions

(1) Two parts of the Palaeoproterozoic Akitkan Group of south Siberia are dated at 1863 ± 9 and 1878 ± 4 Ma by the U–Pb zircon conventional method, confirming the earlier published ages (Neymark *et al.* 1991; Larin *et al.* 2003).

(2) A stable high-temperature remanence has been isolated in sedimentary rocks of the lower part of the Akitkan Group. Positive conglomerate and fold tests confirm that this magnetization is primary. The corresponding Malaya Kosa pole is highly reliable and may be considered as the 1878 Ma key-pole for Siberia.

(3) A stable high-temperature remanence has been isolated in sedimentary rocks of the upper part of the Akitkan Group. Conglomerate tests in two conglomerate layers gave contrasting results, suggesting that this remanence is primary in some studied rocks, but it might be contaminated in other rocks by a secondary remagnetization of uncertain nature. However, all remanence directions are coherent and resemble the direction of coeval rocks in a distal area. We suggest that the overprint direction could be close to the primary direction and the palaeopole is reliable.

(4) A comparison of coeval Malaya Kosa pole and Laurentian Molson B pole demonstrate that the Siberian craton changed its position with respect to the Superior craton during the Mesoproterozoic.

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