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Spatial distribution of mean winter air temperatures in Siberian permafrost at 20–18 ka BP using oxygen isotope data

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Palaeotemperature reconstruction for the period of 20–18 ka BP in Siberia is here based on $\delta^{18}O$ analysis and ^{14}C dating of large syngenetic ice wedges. Dozens of yedoma exposures, from Yamal Peninsula to Chukotka, have been studied. Snow meltwater is considered to be the main source of ice-wedge ice. The modern relationship between $\delta^{18}O$ composition of ice-wedge ice and winter temperature is used as a base for reconstruction. In modern ice wedges (elementary veins that have accumulated during the last 60–100 years) $\delta^{18}O$ fluctuates between −14 and −20‰ in western Siberia and between −23 and −28‰ in northern Yakutia. The trend in $\delta^{18}O$ distribution in ice wedges dated at 20–18 ka BP is similar to the modern one. For example, the $\delta^{18}O$ values in Late Pleistocene wedges are more negative going from west to east by 8–10‰, i.e. from −19 to −25‰ in western Siberian ice wedges to −30 to −35‰ in northern Yakutia. However, values are as high as −28 to −33‰ in north Chukotka and the central areas of the Magadan Region and even as high as −23 to −29‰ in the east of Chukotka. The same difference between the oxygen isotope composition of ice wedges in the eastern and western regions of Siberian permafrost (about 8–10‰) is also preserved from 20–18 ka BP to the present: $\delta^{18}O$ values obtained from large ice wedges from the Late Pleistocene vary from −19 to −25‰ in western Siberia to −30 to −35‰ in northern Yakutia. We conclude that, at 20–18 ka BP, mean January temperatures were about 8–12°C lower (in Chukotka up to 17–18°C) than at present.

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This study focuses on the palaeotemperature interpretation of stable isotope records obtained from syngenetic ice wedges (width up to 3–4 m, height up to tens of metres) contained within yedoma sediments that formed 20–18 ka BP in northern and central Siberia. The isotope record obtained is comparable with those obtained from foraminiferans in deep-sea cores and glacier ice-cores with respect to palaeoclimatic value, authenticity, reliability, and replication of results.

'Yedoma', or ice-rich syngenetic permafrost, accumulated in northern and central Siberia during the Pleistocene. The thickest yedoma sequences accumulated during the Last Permafrost Maximum (LPM), a period between *c.* 20 000 and *c.* 18 000 years ago. During that time, much of Siberia experienced cold permafrost conditions. An area of 'super permafrost' extended almost to the Black Sea, Caspian Sea, and Azov Sea (Fig. 1). The time-span between 20–18 ka BP is a global key period for the LPM. Yedoma formation stopped at the Late Pleistocene–Holocene transition.

As a rule, yedoma sediments have a cyclic structure as shown by alternations of organic and inorganic layers accompanied by a cyclic structure of syngenetic ice wedges. The presence of autochtonous organic material allows ^{14}C dating of the time of formation of the ice wedges (Vasil'chuk 2006, 2013). However, the potential admixture of old allochthonous organic material requires careful selection of material to be dated and for the validation of ^{14}C dates. As the main source for ice-wedge formation is melted snow water, there is a good possibility that the stable isotope record can be connected to winter temperatures. The stable frozen state of yedoma over many thousands of years in Siberia suggests that yedoma is one of the best palaeotemperature archives.

The objectives of this paper were to reconstruct mean January air temperatures based on $\delta^{18}O$ analyses of ice-wedge ice, and mean annual ground temperatures from syngenetic permafrost in Eurasia at the time of the LPM. Application of the oxygen isotope method to ground ice within permafrost permits tentative quantitative reconstructions of palaeoclimatic and palaeogeocryological conditions. This paper is based on our own detailed $\delta^{18}O$ analysis, palaeotemperature reconstruction, and ^{14}C chronology of large syngenetic ice wedges in Siberia that have been dated to 20–18 ka BP (Vasil'chuk 1992, 2006, 2013; Vasil'chuk & Vasil'chuk 1997, 1998a, b; Vasil'chuk *et al.* 2005) and a review of other published data.

Data sources

Considerable information has become available on about 21 sequences of yedoma of northern Eurasia (Fig. 1). We have studied Late-Pleistocene age (*c.* 20–18 ka BP) syngenetic ice wedges throughout the Yamal to the Chukotka and Magadan Region, and from northern Yakutia to the Aldan and Vilyui River valleys (Vasil'chuk 1988, 1990, 1992, 1993, 2006, 2013). We have also summarized the results of many international studies in the Russian permafrost regions

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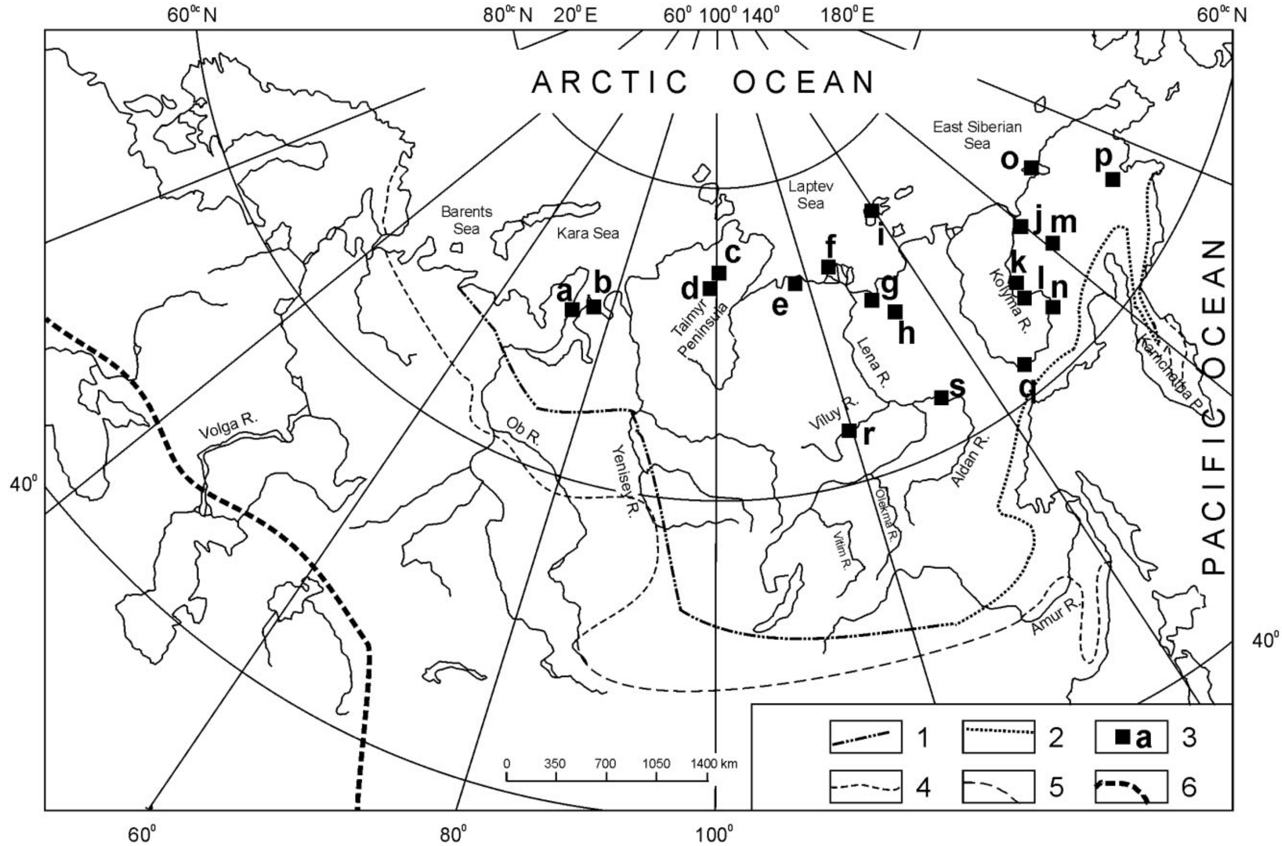

Fig. 1. Map showing locations of key yedoma exposures in Siberia where ice wedges have been dated to about 20–18 ka BP. The exposure localities are indicated as follows: a = Seyaha; b = Era-Maretayakha River estuary; c = Sabler Cape; d = Labaz Lake; e = Mamontov Klyk; f = Mamontova Khayata; g = Chekanovsky Ridge; h = Kular; i = Kotel'ny Island; j = Zelyony Mys; k = Plakhinskii Yar; l = Duvanny Yar; m = Krasivoe; n = Alyoshkinskaya terrace; o = Ayon Island; p = Mayn River; q = Phoenix; r = Tyalychima; s = Mamontova Gora. Also shown are the following: 1, 2 = the southern boundary of modern ice-wedges (1 = known, 2 = estimated); 3 = other yedoma sites mentioned in the text; 4, 5 = the southern boundary of modern permafrost zone (4 = known, 5 = estimated); 6 = the southern boundary of the Late Pleistocene 'super permafrost' zone.

(Fukuda 1993; Chizhov *et al.* 1997; Fukuda *et al.* 1997; Siegert *et al.* 1999; Kotov 1998, 2002; Dereviagin *et al.* 1999, 2002, 2007, 2010; Meyer *et al.* 2002a, b; Schirrmeister *et al.* 2002, 2003, 2008, 2010, 2011; Popp *et al.* 2006; Nikolaev *et al.* 2010; Strauss 2010; Wetterich *et al.* 2011; Oblogov *et al.* 2012).

A specialized methodology of yedoma study was developed about 30 years ago (Vasil'chuk & Trofimov 1984); it includes detailed isotope horizontal and vertical sampling of large syngenetic ice wedges, and radiocarbon dating of surrounding sediments and of organic microinclusions directly in the wedges. This allows the oxygen isotope curves to be placed in a chronological framework.

The present reconstruction of January palaeotemperatures is based on the methodology and relationship established by Vasil'chuk (1992). Special attention is paid here to the period 20–18 ka BP, and thus all calculations are made only for ice wedges that formed during this period. One of the key points of this review is a precise chronology of yedoma and ice wedges.

Syngenetic ice wedges in Siberia

The northern Eurasian permafrost is situated in a vast region (>10 000 000 km^2) with diverse physiographical conditions and Quaternary history (Fig. 1). Late Pleistocene syngenetic ice wedges are widespread in the arctic coastal plains, river valleys, and intermontaneous depressions of northern Eurasia. The vast Arctic coastal plains that occur in western Siberia, Yakutia, and Chukotka have not been glaciated during the past 40 ka.

Northern Eurasia comprises two permafrost zones: (i) continuous permafrost in the north and (ii) discontinuous permafrost in the south. The distribution of permafrost is broadly related to air temperature. The mean annual ground temperature of permafrost varies between slightly below 0°C (in discontinuous permafrost) and between −10 and −13°C (in the northern areas of continuous permafrost) (see also Vandenberghe *et al.* 2014). The boundary between the continuous and discontinuous permafrost zones corresponds to a mean annual air temperature of −5.5°C in eastern Europe, −7.0°C in western Siberia, −7.5°C in Yakutia, and −3.5°C in Chukotka.

Syngenetic ice wedges are a direct indicator of the existence of 'cold' permafrost (French 2012), with mean annual ground temperatures not above −1 to −3°C (Vasil'chuk 2013). Syngenetic ice wedges are formed in such a way that ice becomes vertically stratified.

Large syngenetic ice wedges are widespread in the Russian permafrost zone, including for example: the Seyaha exposure (height of the ice-wedge system is >20 m and depth is 24 m) in the north of western Siberia, near the Zelyony Mys settlement (height about 40 m), Duvanny (height about 45 m), Vorontsovsky (height about 50 m), Stanchikovsky (height about 35 m), and Oiyagossky Yar (height about 40 m) in Yakutia (see Fig. 1 for locations; Kaplina 1986; Vasil'chuk 1992, 2006, 2013; Fukuda 1993; Fukuda *et al.* 1997; Meyer *et al.* 2002a, b; Schirrmeister *et al.* 2010, 2011; Wetterich *et al.* 2011). In North America, similar thick ice wedges were found in the valley of the Titaluk and Itkillik River (height about 30 m), northern Alaska (Carter 1988; Kanevskiy *et al.* 2011). The width of large syngenetic ice wedges varies from 1.5 to 3.0 m. In northern Russia, large syngenetic ice wedges are the dominant form of ground ice. Syngenetic ice-wedge growth proceeds subaerially during the accumulation of peat or peaty sediments. Periodically, when gravel, sand, silt, and clay are deposited under subaqueous conditions, ice-wedge growth decreases or stops. When the subaerial regime returns, ice-wedge growth recommences. If the subaqueous strata is thin enough (<3–4 m), the toes of younger and stratigraphically higher ice wedges penetrate into buried ice wedges of the previous phase. On the contrary, if the thickness of subaqueous sediment exceeds about 4–5 m, the stratigraphically higher ice wedges do not penetrate into the lower wedges. This model of syngenetic ice-wedge growth is supported by the distribution of ice wedges on higher and lower levels of sediment aggradation. For example, the polygonal ice wedges on high flood-plains of northern rivers tends to be widespread, whereas it is rare that on low flood-plain they are found. This confirms that ice-wedge growth occurs preferentially under subaerial conditions (Vasil'chuk 1992; Vandenberghe & Kasse 1993). Oxygen isotope data correspond with long-term subaerial stages of polygonal ice-wedge formation, and radiocarbon determinations relate to subaerial stages of peat accumulation.

The main source of water for syngenetic ice wedges is snow-melt. Minor sources comprise hoarfrost and the melt of active-layer ice (<5–10%). Within high terraces and divides, syngenetic ice wedges are formed exclusively of atmospheric water that freezes within the frost cracks (those of the epigenetic type). On flood-plains and coastal plains, small ice wedges also form from atmospheric water flowing into the frost cracks (if the crack is open to the surface) or from water from the seasonally thawed layer (in the case of intrastratal frost cracks). Highly mineralized water may penetrate into cracks when either there is a salty lake nearby or as a result of an extremely active tide or a surge (Vasil'chuk 2006). Such tides and storm surges usually occur in summer only when the sea or bay surface is free of ice. By this time, the majority of the frost cracks are already closed and thus, in only a few instances does this water freeze within the ice wedge.

Methodology

Sampling method strongly determines the accuracy of the age determination and representativity of the isotope measurements. Sampling comprised both large ice wedges (15–20 m high and 3–3.5 m wide), and small, deeply buried ones (1–3 m high and several centimetres wide).

The most reliable information was obtained from uniformly thick ice-wedge systems, and from small buried ice wedges formed simultaneously. Although $\delta^{18}O$ data from large and small wedges tend to be similar, the larger wedges provide a more complete palaeoclimatic record. The main advantage of sampling small ice wedges is that their age is more accurately known; they are similar to, or slightly less than, the age of the host sediments. However, small ice wedges are rare, often located at the same depth, sometimes isolated within tiered ice-wedge systems. As such, they may have an anomalous $\delta^{18}O$ composition, reflecting the supply of stagnant bog water or water from other non-atmospheric sources (Vasil'chuk 1992). Thus, the fragmentary occurrence of small wedges precludes measurement of a complete isotopic record of the growth of the whole complex. The ranges of isotope data obtained for small and large ice wedges are nearly coincident and the palaeotemperature reconstructions from both give similar results (Vasil'chuk 1990).

Vertical sampling of ice wedges is favoured over horizontal sampling because with the latter it is impossible to establish the exact sequence of ice-wedge formation. Vertical sample spacing was typically 50–100 cm. The isotopic data from horizontal profiles were used as a control. The coincidence of the $\delta^{18}O$ ranges of vertical and horizontal sampling, observed in almost all cases, indicates that the isotope information in the case of vertical sampling of ice wedges is sufficiently complete. Sample size was typically 10×10×10 cm, which made it possible to obtain about 1 L water, from which samples were taken for chemical, palynological, and isotope analyses.

Ice-wedge dating

Each sample of ice-wedge ice is thought to represent a time interval of 100 to 300 years. We estimated this time interval by studying young syngenetic ice wedges with respect to the rate of deposition of their host sediments on late Holocene flood-plains. Small syngenetic ice wedges (8–12 cm wide) are typically 1–1.2 m thick and comprise 32–45 laminae. As the flood-plains periodically flood, once every 3–4 years or more, the age of laminated flood-plain sediment and therefore of the syngenetic ice wedges are unlikely to exceed 100–200 years. The ^{14}C age of 1 m of flood-plain sediment was determined to be 200–300 years based on vertical sequences of ^{14}C dates (for example, sites in the Tanama river and Yenisei river valleys; Vasil'chuk 1992). Although Mackay (1986, 2000) reported that a 10-cm-wide ice wedge can form in one year, our calculations indicate a mean rate of ice-wedge growth of 10 cm in 100–300 years (Vasil'chuk 1992). To construct $\delta^{18}O$ diagrams, samples were arranged in their stratigraphical sequence, from the bottom upwards. However, it cannot be certain that two adjacent samples are not coeval. In some cases, when active postgenetic deformation was noted in the ice wedges, the upper sample may even be somewhat older than the lower one. However, such cases are uncommon and in general the ice in large ice wedges becomes younger from the bottom upwards.

The age of larger syngenetic ice wedges was established by radiocarbon dating of organic matter from the host sediments that accumulated synchronously with the ice wedges (Table 1). The organic matter includes peat, roots, woods, bones, tusks, and dispersed organic plant material. Age control was provided by ^{14}C dating of different kinds of organic matter from the same layer.

As re-deposition of organic material is common in permafrost (Vasil'chuk 2006, 2013), ^{14}C dates should be carefully evaluated, especially those beyond the range of radiocarbon dating; these usually correspond to redeposited organic material in yedoma. Lachniet *et al.* (2012) showed that the dated organic material from ice wedges can yield very different ^{14}C ages: in particular, organic carbon provides ^{14}C ages up to 11 170 years older than dissolved organic carbon or occluded CO_2 gas. Therefore, the youngest ^{14}C date from the data set in a particular horizon is most likely the closest to the actual time of accumulation and freezing of the yedoma sediment (Vasil'chuk 2006, 2013). Based on the principle of the choice of the youngest ^{14}C date, we selected for this paper the horizon formed approximately during 20–18 ka BP in every yedoma outcrops.

Dating of ice wedges is difficult. Dating is either carried out on sediments surrounding the wedges or on organic fragments embedded in the ice. In arctic conditions wind-transported organic material may have an age that is clearly different from the time of formation of the wedge. Connecting the age of the surrounding sediment to the formation of an ice wedge is also far from straightforward. When dating is obtained from interpolation, the time window may not be so narrow (20–18 ka BP). As in Siberia today, the winter climatic conditions were stable for a long period including the time interval 20–18 ka BP; therefore, the exact limits of the time window are not very significant for the accuracy of the palaeoreconstruction.

One of the first attempts at radiocarbon dating of an ice wedge was by Brown (1965) at the Voth polygon site on the Barrow Peninsula, Alaska. In the Russian permafrost zone, radiocarbon dates from ice-wedge complexes are provided by Kaplina (1986), Vasil'chuk (1992, 2006, 2013), and Vasil'chuk & Vasil'chuk (1997, 1998a, b, 2008). In 1998 at Seyaha, an ice wedge was, for the first time, dated by ^{14}C AMS using organic microinclusions taken directly from the ice wedge (Vasil'chuk *et al.* 2000). The total sum of radiocarbon dates used in this latter paper is about 1000. A significant contribution to the ^{14}C study of yedoma was made by the joint German–Russian investigations in northern Siberia and the Russian Arctic Islands from 1995–2012 (Chizhov *et al.* 1997; Dereviagin *et al.* 2002, 2007, 2010; Meyer *et al.* 2002a, b; Schirrmeister *et al.* 2002, 2003, 2008, 2010, 2011; Hubberten *et al.* 2004; Andreev *et al.* 2009, 2011; Wetterich *et al.* 2011).

Organic samples from the same horizon were treated in different ways here in order to determine the optimal procedure of sample preparation. This includes extraction by filtered water from simultaneous ice wedges and treatment by routine laboratory procedures. A comparison of results revealed that prompt laboratory treatment gave reliable dates. Some samples were dried in a nylon envelope in the field and found to give similar results to samples treated by the routine procedure.

We summarize all available $\delta^{18}O$ data of 20–18 ka BP old syngenetic ice wedges from northern Eurasia in Fig. 2B and Table 2. These data are based on a similar methodology of ice-wedge sampling and study. Vasil'chuk (1992, 1993) has previously published the distribution of mean January air temperatures for the period 22–14 ka BP. For the present paper we recalculated data more specifically for the period between 20 and 18 ka BP, and added new data yielded in the last two decades.

The palaeotemperature reconstruction

The $\delta^{18}O$ composition of ice-wedge ice is a function of the isotopic signatures of the contributing sources, the proportions contributed from each of them, and the isotopic changes (by mixing, evaporation, or fractionation) during either freezing or by diffusion after freezing. The parameter most strongly related to the $\delta^{18}O$ composition of ice-wedge ice is winter air temperature. In order to establish this relationship, we compared data on winter air temperature and the $\delta^{18}O$ composition of modern ice wedges <100 years old, each com-

Table 1. Radiocarbon ages obtained by different authors for different types of organic material collected from the Siberian ice-wedge complex (dated at about 20–18 ka BP; fragments of outcrops only).

Radiocarbon age (years BP)	Laboratory number	Depth (m)	Organic material
Seyaha (Vasil'chuk 2006)			
14 720±100	GrA-10539	12.0	Organic microinclusions from ice
20 960±140	GrA-10536	20.6	Organic microinclusions from ice
Era-Maretayakha and Mongatalyangyakha (Vasil'chuk 1992; Oblogov *et al.* 2012)			
3900±100	GIN-2468	0.7	Peat (authothonous)
9100±90	Lu-6534	1.0	Roots
21 930±370	Lu-6542	2.2	Detrital peat
21 900±900	GIN-2469	4.0	Detrital peat
Sabler Cape (Andreev *et al.* 2003)			
12 310±170	AWI-96-3	5.5	Peat with small twigs
18 220±320	AWI-96-3	10.0	Peat with small twigs
19 520±270	AWI-96-6	15.0	Plant remains
26 750±650	AWI-96-7	21.0	Peat
Labaz Lake (Chizhov *et al.* 1997; Siegert *et al.* 1999)			
8960+90/–90	KIA-1409	3.0	Mixed plant remains
20 400+300/–290	KIA-1411	3.7	Mixed plant remains
24 990+520/–480	KIA-1412	3.73	Mixed plant remains
Mamontov Klyk (Schirrmeister *et al.* 2008)			
16 510±60	KIA-25094	5.3	Twig fragments
18 560±100	KIA-25093	7.3	Plant fragments
19 500 +220/–210	KIA-25092	9.3	Grass roots
20 640±90	KIA-25091	11.2	Grass roots
24 600 +170/–160	KIA-25090	12.9	Small stems
Mamontova Khayata (Schirrmeister *et al.* 2002)			
12 020±205	KI-442901	0.6	Peat
13 920±100	KIA-9194	1.5	Plant
17 160±90	KIA-9195	7.0	Dispersed plant material
22 060±150	KIA-10357	8.2	Dispersed plant material
28 470±160	KIA-6716	14.8	Wood
Chekanovsky Ridge, Kurungnakh-Sise (Schirrmeister *et al.* 2003)			
16 980 +90/–80	KIA-12595	1.0	Small twigs
33 490 +380/–390	KIA-12594	3.5	Peat
38 020 +510/–480	KIA-12593	8.0	Peat
Kular (Vasil'chuk 1992)			
33 300±1100	GIN-4987	11.2	Peat (authothonous)
41 100±800	GIN-4977	17.6	Wood
Kotel'ny Island (Makeev *et al.* 1989)			
12 320±130	Lu-	0.5	Peat
19 900±1000	Lu-1790	4.0	Mammoth tusk
28 410±210	Lu-1751	About 8	Allochthonous peat
Zelyony Mys (Vasil'chuk 2006)			
13 600±200	SNU01-003	3.0	Organic microinclusions from ice
26 700±300	SNU01-002	6.5	Organic microinclusions from ice
28 700±500	SNU01-001	8.0	Organic microinclusions from ice
Plakhinski Yar (Vasil'chuk 2006)			
11 490±80	SNU02-130	3.9	Organic microinclusions from ice
17 390±200	SNU01-281	4.5	Organic microinclusions from ice
21 400±300	SNU02-131	8.6	Organic microinclusions from ice
Duvanny Yar (Vasil'chuk 2006)			
14 100±500	SNU02-004	7.0	Organic microinclusions from ice
16 800±800	SNU01-007	13.0	Organic microinclusions from ice
21 900±900	SNU02-136	16.8	Organic microinclusions from ice
Krasivoe (Nikolaev *et al.* 2010)			
18 700±1400	MSU-881	About 10	Dispersed plant material
22 700±1500	MSU-886	About 12	Dispersed plant material
27 300±300	GIN-3209	About 15	Dispersed plant material
Alyoshkinskaya terrace (Vasil'chuk 1992)			
14 780±300	SOAN-2307	5.0	Dispersed plant material
17 260±140	SOAN-2308	7.0	Dispersed plant material
Ayon Island (Vasil'chuk 1992)			
10 180±280	GIN-4967	0.5	Rootlets
28 100±800	GIN-4969	21.0	Rootlets
28 600±1000	GIN-4968	21.5	Rootlets
Mayn River, Ledovy Obryv (Vasil'chuk 1992)			
19 500±500	MAG-815	1.0	Rootlets (possibly in situ)
22 300±200	MAG-814	3.0	Rootlets (possibly in situ)
23 500±500	MAG-813	About 12	Rootlets (possibly in situ)
Pit Phoenix and Utinaya (Vasil'chuk 2006)			
11 000±80	SNU02-143	2.0	Organic microinclusions from ice
31 390±320	GIN-8925	12.0	Wood
Vilyui near Tyalychima River (Vasil'chuk 1992)			
7070±60	GEO-MSU-1	5.5	Peat
32 785±40	GEO-MSU-2	17.0	Dispersed plant material
Mamontova Gora (Vasil'chuk 2006)			
17 040±100	SNU01-283	2.6	Organic microinclusions from ice
19 800±600	SNU01-284	3.2	Organic microinclusions from ice
19 050±180	SNU01-285	5.0	Organic microinclusions from ice
18 400±400	SNU02-140	6.9	Organic microinclusions from ice
18 900±200	SNU02-139	7.2	Organic microinclusions from ice

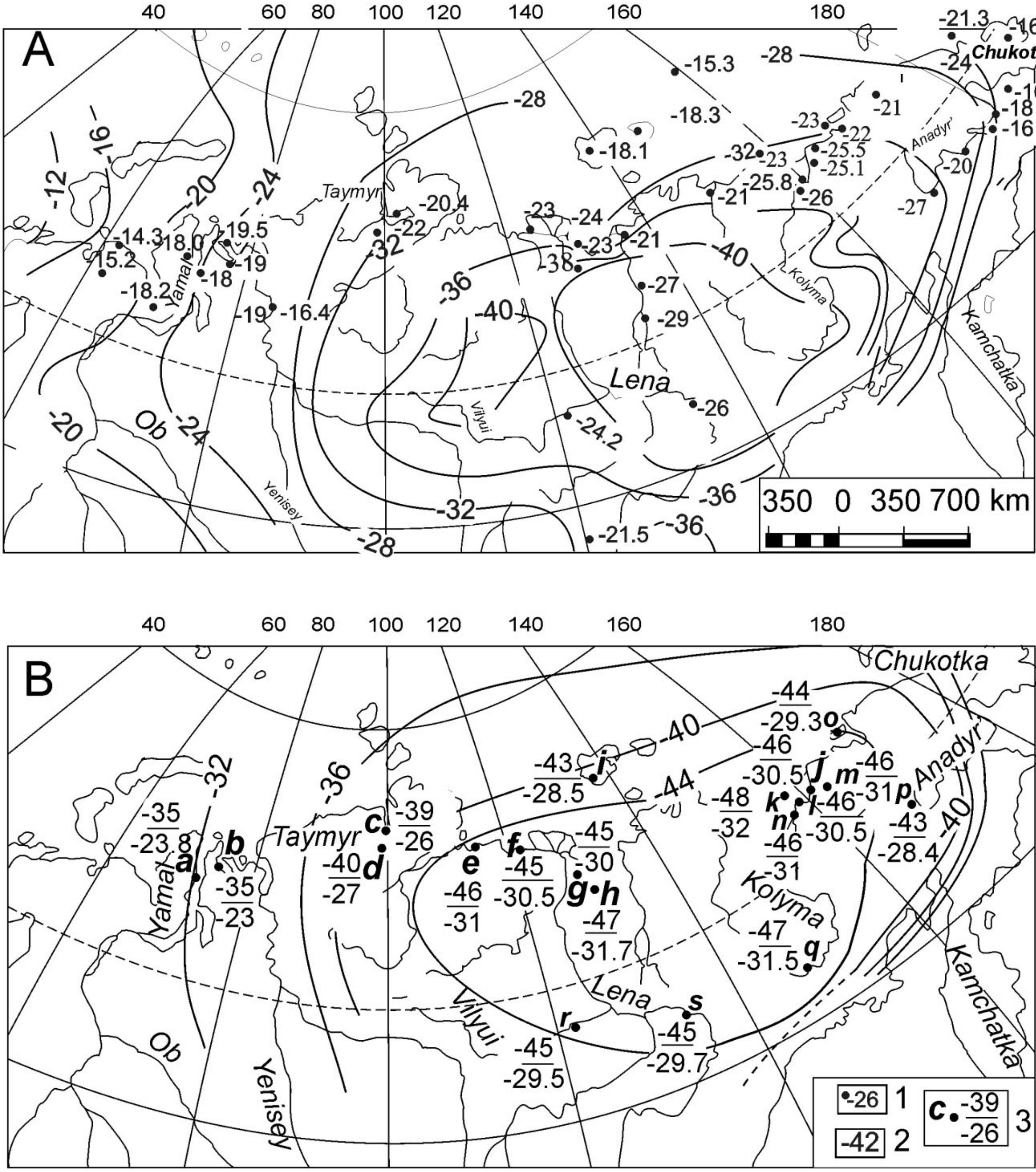

Fig. 2. Maps of northern Siberia showing (A) isotherms of modern mean January surface air temperatures and sites with mean $\delta^{18}O$ values obtained from modern syngenetic ice wedges (after Vasil'chuk 1992, 1993 with additions) and (B) isotherms of mean January surface air temperatures and sites with mean $\delta^{18}O$ values obtained from syngenetic ice wedges dated at 20–18 ka BP; also indicated are reconstructed mean January surface air temperatures using $\delta^{18}O$ data (after Vasil'chuk 1993, 2006, 2013, with corrections). 1 = Mean $\delta^{18}O$ values in modern ice wedges; 2 = isotherms of mean January temperatures; 3 = reconstructed air temperature and isotope values: the numerator is reconstructed mean January surface air temperatures, the denominator is mean $\delta^{18}O$ value of ice wedges dated at 20–18 ka BP. Letters indicate key yedoma sites where ice wedges have been dated to about 20–18 ka BP (see Fig. 1 and Tables 1, 2). All temperatures in °C.

Table 2. Mean January air temperatures in Siberian permafrost at 20–18 ka BP and comparison with modern values (after Vasil'chuk 1992 with additions and amendments).

Site location on Fig. 1	Coordinates	$\delta^{18}O_{ice\ wedge}$ (‰)		T_J (°C)		References
		20–18 ka BP	Modern	20–18 ka BP	Modern	
a. Seyaha	70°10′N, 72°12′E	–23.8	–18	–35	–23	Vasil'chuk (1992, 2006)
b. Era-Maretayakha	71°39′N, 75°25′E	–23	–19	–35	–27	Oblogov *et al.* (2012), Vasil'chuk (1992)
c. Sabler Cape	74°33′N, 100°32′E	–26	–20.4	–39	–33	Dereviagin *et al.* (2002, 2010)
d. Labaz Lake	72°18′N, 99°40′E	–27	–22	–40	–34	Chizhov *et al.* (1997), Siegert *et al.* (1999)
e. Mamontov Klyk	73°36′N, 117°11′E	–31	–21.3	–46	–33	Schirrmeister *et al.* (2008), Boereboom *et al.* (2013)
f. Mamontova Khayata	71°61′N, 129°28′E	–30.5	–23	–45	–33	Meyer *et al.* (2002a)
g. Chekanovsky Ridge	72°53′N, 125°11′E	–30	–23	–45	–41	Schirrmeister *et al.* (2003)
h. Kular	70°38′N, 131°53′E	–31.7	–26	–47	–41	Vasil'chuk (1992)
i. Kotel'ny Island	75°26′N, 138°49′E	–28.5	–18.1	–43	–29	Vasil'chuk (1992)
j. Zelyony Mys	69°N, 161°E	–30.5	–25.5	–46	–33	Vasil'chuk (1992)
k. Plakhinski Yar	68°40′N, 160°17′E	–32	–25.8	–48	–32	Vasil'chuk (1992), Fukuda *et al.* (1997)
l. Duvanny Yar	68.63°N, 159.14°E	–30.5	–25.1	–46	–35	Vasil'chuk (2006), Strauss (2010)
m. Krasivoe	68°18′N, 161°44′E	–31	–26	–46	–35	Nikolaev *et al.* (2010)
n. Alyoshkin-skaya terrace	68.72°N, 158.4°E	–31	–26	–46	–35	Vasil'chuk (1992)
o. Ayon Island	69°38′N,168°35′E	–29.3	–23	–44	–29	Vasil'chuk (1992)
p. Mayn River, Ledovy Obryv	64°06′N, 171°01′E	–28.4	–20	–43	–27	Vasil'chuk (1992)
q. Pit Phoenix	62°15′N, 150°45′E	–31.5	–27	–47	–37	Vasil'chuk (1992)
r. Vilyui near Tyalychima River	64°N, 126°E	–29.5	–24.2	–45	–37	Vasil'chuk (1992)
s. Mamontova Gora	63°N, 134°E	–29.7	–26	–45	–43	Vasil'chuk (1992), Popp *et al.* (2006)

prising 8–12 elementary ice veins, in different regions of the Eurasian permafrost zone (Vasil'chuk 1990, 1992, 2006, 2013).

The $\delta^{18}O$ values from modern syngenetic ice wedges on flood-plains, low coastal plain (marshes), and syngenetic peat in northern Eurasia are as follows: north-eastern Europe –15 to –12‰, north of western Siberia –20 to –16‰, northern Yakutia –28 to –24‰, central Yakutia –27 to –25‰, Chukotka –17 to –15‰, central Magadan Region –27 to –25‰, Trans-Baikal Region –22 to –20‰. These data have been averaged and compared with winter surface air temperatures for each region (Vasil'chuk 1990, 1992, 2006, 2013). For comparison, $\delta^{18}O$ values from modern ice wedges in the Mackenzie Delta area and the Yukon Coastal Plain in Canada vary from –26 to –17‰, although 90% fall within a smaller range of –24.5 to –22‰ (Mackay 1983; Michel 1990). Lauriol *et al.* (1995) measured a range of about –27 to –23‰ from ice wedges along the Porcupine River, Yukon.

$\delta^{18}O$ values in modern syngenetic ice wedges ($\delta^{18}O_{iw}$) show a strong empirical relationship with winter air temperatures. These relationships are expressed in the following simplified regression equation (Vasil'chuk 1990, 1992):

$$t_{\text{mean January}} = 1.5\,\delta^{18}O_{iw}(\pm 3°C). \quad (1)$$

where $t_{\text{mean January}}$ is mean January temperature of the period of modern ice-wedge formation during last 60–100 years; $\delta^{18}O$ is oxygen isotope composition of ice-wedge ice formed during last 60–100 years.

There is a good correlation between $\delta^{18}O$ in present syngenetic ice-wedge ice and mean January air temperatures in Eurasia. This is mapped in Fig. 2A. Both of these properties demonstrate similar isoline distribution.

We applied Equation 1 to reconstruct mean January temperatures from ice wedges formed at 20–18 ka BP using mean values of $\delta^{18}O$ for this period (Table 2). Although we understand that there are uncertainties in palaeotemperature reconstructions caused by short-term variations of air temperatures and stable isotope records (i.e. modern ones vary by 4‰ during 2–3 decades), we conclude that mean January temperatures at 20–18 ka BP were similar to modern ones. This includes the character of the trends but not the values. The latter ranges from about –33 to –35‰ in western Siberia, –45 to –48‰ in northern Yakutia, and –31 to –43‰ in Chukotka (Table 3).

Winter air temperature is the main factor controlling the permafrost ground temperature in northern Eurasia. The duration of winters is about 8–10 months in these regions, summer duration is only 2–4 months; hence, even a two- or threefold increase in total summer temperatures would not significantly influence annual

Table 3. Mean January temperature at 20–18 ka BP in Siberia in comparison with modern temperatures.

Area	T_J (°C)	
	20–18 ka BP	Modern
Western Siberia (Yamal and Gydan Peninsula)	–33 to –35	–23 to –27
Taymyr Peninsula	–39 to –40	–33 to –34
Northern Yakutia	–45 to –48	–32 to –41
Chukotka	–31 to –43	–21 to –29

temperatures. Analysis of the $\delta^{18}O$ values of syngenetic ice wedges suggests that mean January temperatures 20–18 ka BP were about 8–12°C below modern ones. In areas with changeable environments, such as Chukotka, mean January temperatures were up to 17–18°C colder than modern ones (Fig. 2B).

A north–south distribution of the mean January air temperatures and permafrost temperatures isolines at 20–18 ka BP is characteristic of the western sector of the Eurasian permafrost zone, including the northern areas of European Russia and western Siberia. This reflects the prevailing western and northern direction of air masses and moisture transport (Vandenberghe *et al.* 2012). An east–west direction of isolines over the vast area that extends from the Taimyr Peninsula to Western Chukotka marks the influence of continentality relative to the Arctic Ocean.

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In modern ice wedges the mean $\delta^{18}O$ value decreases by 8‰ (from −18 to −26‰) from western Siberia to the east of northern Yakutia. This is over a distance of about 2000 km. The same decrease (from −22 to −30‰) is characteristic for the 20–18 ka old ice wedges in this same area (Vasil'chuk 1992).

One can see that the Late Pleistocene permafrost zone in the different regions of northern Eurasia shows close similarities with the present Yakutian type represented by low values of ground and air temperature, winter duration, and low temperature gradients from north to south. Therefore, it can be proposed that during the LPM a vast permafrost 'super zone' existed throughout Eurasia. The primarily continental Late Pleistocene western Europe climate provides support for the concept of widespread Northern Atlantic sea-ice cover.

Some investigators have speculated that the 20–18 ka BP ice sheet covered an area stretching from the Yamal and Gydan Peninsulas to the Novosibirsk Islands (e.g. Grosswald 1998). However, the areas with undisturbed Pleistocene syngenetic ice wedges have not been glaciated during or since the time of ice-wedge formation (Vasil'chuk 1990, 1992). Figure 1 shows the distribution of undisturbed syngenetic ice wedges that accumulated during the period 20–18 ka BP. These regions are all coastal and alluvial plains and plateaus found in western Siberia, Yakutia, and Chukotka. Syngenetic ice wedges also formed in mountainous areas, for example in the inter-mountainous depressions of Kular in northern Yakutia and in the Phoenix sequence in the Upper Kolyma valley in the Magadan Region (Vasil'chuk 1992). If these areas were covered by glacier ice 20–18 ka BP, syngenetic permafrost sediments and ice wedges of this age would not have survived.

Additional evidence against Late Pleistocene glaciation is provided by the distribution of $\delta^{18}O$ values in syngenetic ice-wedge ice. As discussed above, the $\delta^{18}O$ composition in syngenetic ice wedges of 20–18 ka BP and that in modern-age wedges show a similar decrease of about 8–10‰ from the Yamal and Gydan Peninsulas to northern Yakutia. This precludes the possibility of major rearrangements in the atmosphere and cryosphere at 20–18 ka BP.

Conclusions

The palaeotemperature reconstructions discussed here are based on the study of the unusual characteristics of yedoma sediments. This includes detailed isotope horizontal and vertical sampling of large syngenetic ice wedges, radiocarbon dating of surrounding sediments and organic microinclusions directly from ice wedges, and the determination of oxygen isotope values in their chronological framework. Close attention has been paid to ice-wedge fragments formed 20–18 ka BP. Vasil'chuk's (1992) relationship between the $\delta^{18}O$ value of recent ice wedges and January temperatures was used for the reconstruction of January palaeotemperatures at the LPM.

The results of our palaeoclimatic reconstructions in northern Eurasia 20–18 ka BP are as follows:

- Syngenetic ice wedges were formed in the northern areas of the Eurasian permafrost zone during 20–18 ka BP.
- The trend in the $\delta^{18}O$ distribution in ice wedges during 20–18 ka BP is similar to the modern one, i.e. $\delta^{18}O$ is more negative from west to east by 8–10‰, from −19 to −25‰ in western Siberian ice wedges to −30 to −35‰ in northern Yakutia, becoming as high as −28 to −33‰ in north Chukotka and central areas of the Magadan Region, and as high as −23 to −29‰ in east Chukotka.
- Mean January temperatures over northern Siberia were ~8–12°C less than modern ones. In areas with changeable climatic conditions, such as Chukotka, the range of mean January temperatures was up to 17–18°C less than in modern times.
- The Eurasian permafrost zone during 20–18 ka BP was similar to the modern Yakutian type.
- $\delta^{18}O$ trends in LPM syngenetic ice wedges in Siberia suggest that air-mass transport throughout Subarctic Asia was similar to today. Westerly transport was prevalent over a considerable part of northern Eurasia. Atlantic influences prevailed from the Yamal Peninsula to NE Yakutia, but these influences were possibly weaker then at present, due to more frequent cold and dry Arctic air-mass advection. It is also likely that the influence of Pacific air masses was negligible in the eastern part of northern

Asia. It appears that a continental anticyclone regime dominated there, particularly in winter.

- Our palaeoclimatic reconstruction precludes any significant change in permafrost distribution in northern Eurasia at 20–18 ka BP. The uniformity of $\delta^{18}O$ values in ice-wedge ice throughout large areas of the Eurasia permafrost zone indicates uniformity of atmospheric circulation at that time.

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