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Holocene Paleohydrology and Paleoclimate at Treeline, North-Central Russia, Inferred from Oxygen Isotope Records in Lake Sediment Cellulose

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Lake-water oxygen-isotope histories for three lakes in northern Russia, derived from the cellulose oxygen-isotope stratigraphies of sediment cores, provide the basis for preliminary reconstruction of Holocene paleohydrology in two regions along the boreal treeline. Deconvolution of shifting precipitation $\delta^{18}O$ from secondary evaporative isotopic enrichment is aided by knowledge of the distribution of isotopes in modern precipitation, the isotopic composition of paleo-waters preserved in frozen peat deposits, as well as other supporting paleoclimatic information. These data indicate that during the early Holocene, when the boreal treeline advanced to the current arctic coastline, conditions in the lower Yenisey River region were moist compared to the present, whereas greater aridity prevailed to the east near the lower Lena River. This longitudinal moisture gradient is consistent with the suggestion that oceanic forcing (increased sea-surface temperatures in the Nordic Seas and reduced sea-ice cover) was a major contributor to the development of a more maritime climate in western Eurasia, in addition to increased summer insolation. East of the Taimyr Peninsula, large tracts of the continental shelf exposed by glacial sea-level drawdown may have suppressed maritime climatic influence in what are now coastal areas. In contrast, during the late Holocene the two regions have apparently experienced coherent shifts in effective moisture. The similarity of the records may primarily reflect reduced North Atlantic influence in the Nordic Seas and southward retreat of coastline in eastern Siberia, coupled with declining summer insolation.

Key Words: **stable isotopes; sediment cellulose; paleoclimate; paleohydrology; treeline; Russia.**

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INTRODUCTION

The boreal forest in northern Eurasia responded dramatically to Holocene climate change. Radiocarbon dating of tree macrofossils found in the present-day tundra (MacDonald *et al.*, 2000) and palynological studies on lake-sediment cores from the tundra near the lower Yenisey (Hahne and Melles, 1997) and Lena Rivers (Pisaric *et al.*, in press) indicate that the boreal forest expanded to the arctic coast between 9000 and 7000 yr B.P. and retreated to its present location between 4000 and 3000 yr B.P., representing a northward vegetation shift of hundreds of kilometers in some sectors. Comparison with modern temperatures near the treeline suggests that mean summer temperatures may have been 2.5° to 5.0°C warmer than today between 9000 and 4000 yr B.P. (MacDonald *et al.*, 2000), results that compare well with general circulation model simulations for the Holocene (Kutzbach *et al.*, 1993).

MacDonald *et al.* (2000) suggest that the early Holocene

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expansion of the boreal forest was probably linked to several factors including: (1) elevated summer insolation; (2) Eurasian ice-sheet decay; (3) reduced sea-ice cover; (4) greater continentality resulting from lower sea level; and (5) increased penetration of North Atlantic waters into the Arctic, which led to enhanced flow of warm and moist air into northern Eurasia. Climate models indicate that oceanic forcing (i.e., increased sea-surface temperatures (SSTs) and reduced sea ice) and increased summer insolation may have been the major factors contributing to high-latitude summer warming at 6000 yr B.P. (Kerwin *et al.*, 1999). Despite the apparent complexity of climate change, evaluation of Holocene vegetation dynamics across northern Eurasia has focused primarily on temperature (Khotinsky, 1984; Foley *et al.*, 1994; Velichko *et al.*, 1993; 1995; 1997; TEMPO, 1996; Kremenetski *et al.*, 1998). In comparison, current understanding of Holocene moisture history is rather limited, having been derived from relatively few pollen-based reconstructions (Koshkarova, 1995; Hahne and Melles, 1997; Velichko *et al.*, 1997), which may contain large uncertainties due to poor pollen representation of important Siberian treeline species (cf. Clayden *et al.*, 1996). Monserud *et al.* (1998) used a bioclimatic model to develop more comprehensive precipitation reconstructions, but these are restricted to the 6000 to 4600 yr B.P. interval. Furthermore, only general consideration has been given to the relationship between Holocene moisture conditions at treeline and possible climate forcing mechanisms (Velichko *et al.*, 1997).

Here we use lake-water oxygen-isotope histories, inferred from stratigraphic analysis of cellulose in the fine-grained fraction of lake-sediment cores, to reconstruct Holocene paleohydrology at the boreal treeline near the lower Yenisey and Lena Rivers of northern Russia. These studies are supplemented by isotopic analysis of samples of modern meteoric water, including precipitation, groundwater, and ground-ice, which provide insight into the modern isotope hydrology. The isotopic records obtained from porewaters extracted from frozen peat obtained at nearby sites contribute additional information for paleohydrologic reconstruction. Integration of these data establishes a framework for separating variations in lake-water evaporative enrichment from changes in the isotopic composition of precipitation in the cellulose-inferred lake-water oxygen isotope records. Our assessment of changing moisture conditions provides independent support for the importance of paleoceanographic factors that likely contributed to early Holocene warming in northern Eurasia and subsequent cooling during the late Holocene.

SETTING

Climate

A monsoonal climate characterizes the entire arctic coast of western Siberia (Lydolph, 1977). Cool, humid winds commonly blow inland from the Arctic Ocean during the summer, although North Atlantic air masses generally provide the predominant source of moisture to this region. During winter, air is carried northward from the cold land surface to the Arctic Ocean. The unusually mild winters of the 1980s that were centered in this region were apparently associated with increased intensity of cyclones originating over the North Atlantic, which traversed the Barents and Kara Seas to bring strong westerly flow into Siberia (Rogers and Mosley-Thompson, 1995).

The lower Yenisey River region experiences large annual temperature variations with average January temperatures ranging from −32° to −28°C. Average July temperatures vary between 4° and 12°C (Lydolph, 1977). Greater influence of continental air masses during the summer is mainly responsible for higher mean annual temperature in the northern boreal forest compared to the tundra, where the climate is more directly modulated by the cold Arctic Ocean. Mean relative humidity of surface air in July is 70 to 80%. Annual precipitation ranges from about 250 mm along the coast to 350 mm further inland. Snowfall constitutes about 45% of the annual precipitation in the tundra zone and 25–33% of the annual precipitation in the northern boreal forest.

Eastward toward the lower Lena River, the winter climate becomes increasingly continental due to strong influence from the Asiatic High, which impedes intrusions of marine air (Lydolph, 1977). In summer, cyclonic storms along the arctic coast bring limited moisture from the Arctic Ocean to the continent. Average January and July temperatures range from −40° to −32°C and 4° to 12°C, respectively. Mean relative humidity of surface air in July is roughly 65%. The majority of precipitation falls during the summer, and the total annual accumulation is roughly 200 mm.

Study Sites

Sediment-cellulose oxygen-isotope records have been obtained from 2 lakes near the lower Yenisey River and one lake near the Lena River Delta (Fig. 1; Clayden *et al.*, 1997; Laing *et al.*, 1999; Wolfe *et al.*, 1999; Pisaric *et al.*, in press). Peatlands from which frozen porewater for isotope analysis was extracted are located in forested regions of the lower Yenisey (68°10′N, 87°09′E) and Lena Rivers (69°23′N, 125°08′E; Fig. 1). General limnological characteristics of lakes in these two regions are reported in Duff *et al.* (1999).

Middendorf Lake (informal name; 70°22′N, 87°33′E) is roughly 90 m by 100 m and has a maximum depth of about 11.5 m. The water is currently alkaline (pH 9.2), oligotrophic (total phosphorus = 12 μg/L), and chemically dilute (specific conductivity = 91 μS). The lake is a headwater basin draining a small catchment. At the time of sampling, it appeared to be closed hydrologically, with evidence of limited melt-season overflow via a moist meadow at the southwestern shore. Local vegetation is dominated by *Betula nana*, with shrub *Salix* and *Alnaster fruticosa*. The lake lies approximately 50–75 km north and west of the mapped northern limits of *Larix sibirica* forest-tundra.

Derevanoi Lake (informal name; 69°14′N, 86°34′E) is sim-

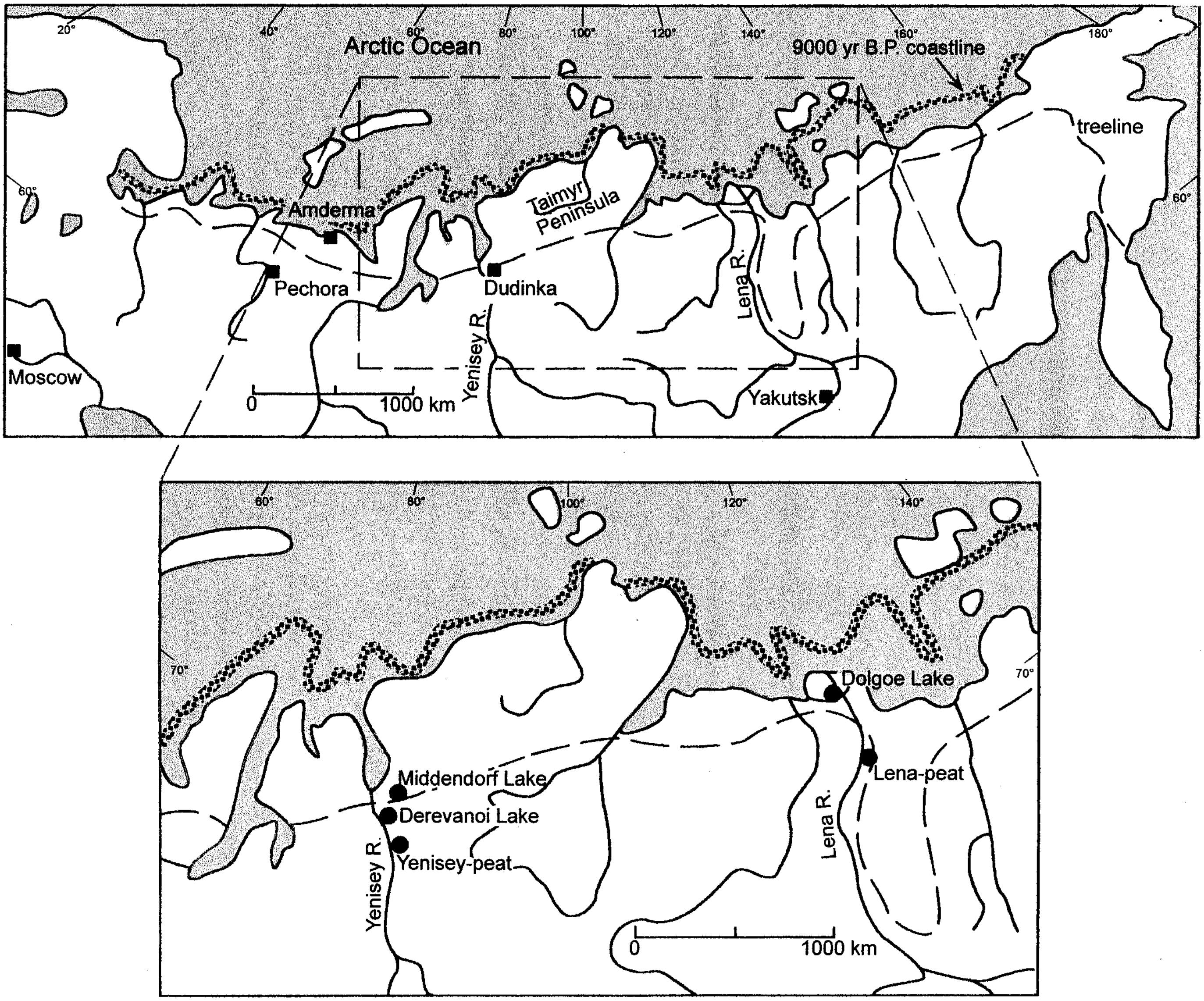

FIG. 1. Maps of northern Eurasia showing inferred coastline at 9000 yr B.P. (from MacDonald *et al.*, 2000), modern position of treeline, and locations of study sites. Conifer treeline on the northern Russian Plain and western Siberia is formed by *Picea obovata* (Russian spruce) and *Larix sibirica* (Siberian larch). In eastern Siberia (from the Taimyr Peninsula east) the conifer treeline is formed by *Larix dahurica* (Dahurian larch). In the very far east it is formed by the shrubby species *Pinus pumila.* Several species of arboreal *Betula* (birch) are also found in the treeline zone across the Russian Federation.

ilar in surface area to Middendorf Lake, though shallower (4.5 m) and somewhat lower in pH (7.9), total phosphorus (9 μg/L), and specific conductivity (30 μS). The more dilute chemistry is consistent with rapid hydrologic flushing in this open basin, which is fed by several lakes and streams draining an extensive catchment. Trees in this area of forest-tundra are mostly *L. sibirica* and *Larix dahurica*. *Betula verrucosa* and *Betula pubescens* are rare, while *B. nana, A. fruticosa, Juniperus communis*, Ericaceae, Cyperaceae, and herbs occur in open areas.

Dolgoe Lake (informal name; 71°52′N, 127°04′E) is substantially larger in surface area than the Yenisey lakes, spanning 3500 m by 500 m, although maximum depth is only 4 m. Chemical characteristics are slightly more dilute (pH 7.4, total phosphorus = 7 μg/L, specific conductivity = 25 μS). The lake is a headwater basin that drains a small catchment lying within one kilometer of the Lena River, at an elevation of about 25 m above river level, beyond the influence of flooding. Outflow currently drains through a fractured sandstone ridge on the eastern side of the basin. Dolgoe Lake is located approximately 10 km north of the mapped treeline. Local vegetation is shrub tundra with *B. nana, Alnus crispa, Salix,* and Ericaceae.

FIELD AND LABORATORY METHODS

Samples of rain were collected in 30-ml high-density polyethylene (HDPE) bottles. Snow and ground-ice were sealed in polyethylene bags, where they melted completely, and then were transferred into HDPE bottles. Shallow groundwater was obtained from springs at the site of discharge or by collecting water-saturated soil from near the base of the active layer. Soil samples were sealed and triple-wrapped in heavy-gauge polyethylene bags to prevent vapor loss during transport to the laboratory for subsequent azeotropic distillation of porewater (Revesz and Woods, 1990). Samples from the lower Yenisey River region were collected between 28/07/93 and 05/08/93 and from the lower Lena River region between 23/07/94 and 07/08/94.

A 3.5-m-thick section of peat located in the forest of the lower Yenisey River region was sampled from 2.15 m below the surface to the bottom of the section in 5-cm-thick slices at 10 cm intervals for porewater isotope analysis. In the forest of the lower Lena River region, a 3.9-m peat core was obtained. Below 50 cm depth, 2- to 3-cm sections at 10- to 20-cm intervals were sampled for porewater isotope analysis. Water thawed from frozen peat samples was extracted using a centrifuge in the lab and transferred to HDPE bottles.

Analyses of $^{18}O/^{16}O$ and $^{2}H/^{1}H$ ratios in water samples were performed at the Environmental Isotope Laboratory (EIL), University of Waterloo, using CO_2-equilibration (Epstein and Mayeda, 1953) and Zn-reduction (Coleman *et al.*, 1982), respectively. Results are reported as δ values, representing deviation in per mil (‰) from the international Vienna-SMOW standard, such that $\delta = [(R_{\text{sample}}/R_{\text{V-SMOW}}) - 1]* 1000$, where R is the $^{18}O/^{16}O$ or $^{2}H/^{1}H$ ratio in the sample and standard. The $\delta^{18}O$ and $\delta^{2}H$ values are normalized to −55.5 and −428‰, respectively, for SLAP (Standard Light Antarctic Precipitation) (see Coplen, 1996). Analytical uncertainties are ±0.2‰ for $\delta^{18}O$ and ±2‰ for $\delta^{2}H$.

Sediment cores from Middendorf, Derevanoi, and Dolgoe Lakes, spanning 2.7, 4.2, and 3.5 m, respectively, were obtained using a Livingstone piston corer (Wright *et al.*, 1984). Sediment core samples generally were taken at 5-, 10-, and 7-cm intervals, respectively, from the three lakes. Samples were pretreated in a 10% hydrochloric acid solution at 70°C for 2 h to dissolve trace amounts of shell or mineral carbonate. Bulk elemental carbon and nitrogen compositions were determined on the <500-μm acid-washed residue using a Carlo Erba elemental analyzer. Additional sample preparation on this fraction, involving solvent extraction, bleaching, and alkaline hydrolysis, removed noncellulose organic components (Edwards and McAndrews, 1989; Edwards *et al.*, 1997). Nickel-tube pyrolysis was used to produce CO_2 from the sediment cellulose fraction for $^{18}O/^{16}O$ determination and results were calibrated to in-house water and cellulose standards (Edwards *et al.*, 1994; Buhay *et al.*, 1995; Elgood *et al.*, 1997; see Table 1). Oxygen isotope ratios from cellulose-derived CO_2 are reported in δ-notation with respect to Vienna-SMOW. Estimated uncertainty in sediment cellulose $\delta^{18}O$ results are ±0.4, 0.4, and 1.2‰ for the Middendorf, Derevanoi, and Dolgoe samples, respectively, based on repeated analyses (Table 1).

RESULTS AND DISCUSSION

Modern Isotope Hydrology

Nonevaporated water (snow, rain, groundwater, and ground-ice) sampled in the lower Yenisey and Lena River regions plot close to the global meteoric water line (GMWL) of Craig (1961; $\delta^{2}H = 8\ \delta^{18}O + 10$; Fig. 2). The distribution of data points along the GMWL mainly reflects seasonal variation in temperature-dependent fractionation during condensation of atmospheric vapor and other aspects of air mass history (Dansgaard, 1964; Rozanski *et al.*, 1993). As a result, summer rain is predictably enriched in ^{18}O and ^{2}H relative to snow, whereas groundwater and ground-ice data are intermediate due to recharge by varying mixtures of snowmelt and rain. Previously reported $\delta^{18}O$ values of snow by Nikolayev and Mikhalev (1995) from the Severnaya Zemlya archipelago (−24.2 to −14.4‰) and near the Yenisey River (−26.4 to −18.2‰) are similar to our one-time snow sampling results from the Yenisey study area (−23.1 to −18.0‰). However, a comprehensive two-yr study (Boike, 1997) in a catchment on the Taimyr Peninsula near the Yenisey sampling region revealed a significantly broader $\delta^{18}O$ range from precipitation (−32.4 to −9.1‰), compared to our smaller data set (−23.1 to −13.6‰).

Estimated oxygen isotope composition of mean annual precipitation ($\delta^{18}O_P$) from groundwater and ground-ice data is −16.9‰ for the treeline near the lower Yenisey River, whereas the more northeasterly Taimyr Peninsula precipitation data suggest a $\delta^{18}O_P$ value of −20.1‰. The roughly 3‰ discrepancy may reflect warmer mean annual temperature (MAT) in our study region or that ^{18}O-enriched thaw-season summer precipitation may be the dominant source of the groundwater and ground-ice that we sampled, as was also found by Boike (1997) and others in tundra locations (Mikhalev, 1989; Bursey *et al.*, 1991; Vardy *et al.*, 1997). Both estimates of $\delta^{18}O_P$ are more negative than those of IAEA/WMO network stations to the west at Moscow, Pechora, and Amderma (Fig. 1), as expected because of the greater rain-out of atmospheric vapor largely originating over the North Atlantic (Table 2). Good agreement occurs between the slope coefficients of the spatial $\delta^{18}O_P$–MAT relation for Amderma, Yenisey, Taimyr (data from Boike, 1997), and Lena data ($\delta^{18}O_P = 0.59$ MAT − 11.47, $R^2 = 0.91$) and data from 47 European IAEA/WMO stations to the west ($\delta^{18}O_P = 0.59$ MAT − 14.35; Rozanski *et al.*, 1992).

Peatland Development and Porewater Isotope Stratigraphy

Paleoecological investigations on a 3.5-m peat section near the lower Yenisey River (Fig. 1) indicate that peat, characterized by Cyperaceae-dominated fen vegetation, began to accumulate at about 6000 yr B.P. (B. G. Warner and A. A. Andreev, unpublished data; Fig. 3a). The vegetation remained largely unchanged until about 2000 to 1500 yr B.P., suggesting that

TABLE 1
Bulk Organic Carbon/Nitrogen Weight Ratios and Sediment Cellulose δ^{18}O Data

Middendorf Lake				Derevanoi Lake				Dolgoe Lake			
Depth (cm)	C:N	$\delta^{18}O_{cell}$ [a] (‰ V-SMOW)		Depth (cm)	C:N	$\delta^{18}O_{cell}$ [a] (‰ V-SMOW)		Depth (cm)	C:N	$\delta^{18}O_{cell}$ [a] (‰ V-SMOW)	
8.5	7.1	16.0		4.5	10.7	12.9		0.5	15.2	13.9	12.6
12.5	7.5	16.6		14.5	13.8	11.6		5.5	15.2	14.2	
14.5	8.2	11.1		25	19.1	12.1	11.0	10.5	15.2	7.2	
18.5	9.1	15.5		40	12.5	11.1		15.5	16.6	9.9	
20.5	8.9	13.9		50	14.6	12.1		20.5	16.2	6.9	
22.5	9.4	11.0		60	13.5	12.5		25.5	14.5	8.1	
26.5	9.0	9.4		70	16.5	11.6		30.5	16.4	8.5	
28.5	10.4	11.2	11.7	80	11.4	13.6		35.5	16.1	11.4	
32.5	9.6	9.5	7.8	90	13.8	12.8	12.4	42.5	14.9	4.2	
38.5	10.2	12.0		100	13.5	14.0		49.5	16.0	7.9	
40.5	9.8	14.6		110	13.5	15.3		56.5	14.9	11.0	
44.5	10.4	13.0		120	11.4	15.5		61	15.8	3.2	
48.5	9.8	10.1	10.1	135	11.3	15.5		68	13.6	15.4	17.4
55.0	9.0	17.5		145	11.9	16.5	15.5	75	13.5	9.2	
62.5	9.5	16.0		155	11.9	16.0		82	12.7	14.5	
67.5	9.7	16.1		165	10.7	16.4		89	15.5	8.6	
72.5	9.9	16.4		175	11.5	15.4		96	10.3	8.7	
77.5	9.9	14.5		185	11.6	14.6		103	12.5	16.2	12.5
82.5	10.4	13.0		195	12.9	15.4		110	14.2	14.5	
87.5	9.8	17.3		205	13.0	16.0		117	15.5	12.2	
92.5	10.6	13.6	12.6	232	13.1	14.0	14.3	124	13.5	8.3	
97.5	10.3	17.5		240	13.0	14.8		131	12.0	14.0	
102.5	9.3	16.4		250	11.6	15.2		138	13.0	9.0	
107.5	9.6	12.0	12.4	260	11.9	16.4	15.7	145	14.6	16.0	12.7
112.5	10.2	16.6		270	11.6	14.2		152	12.9	10.8	
117.5	11.7	17.1		280	12.1	13.5		159	13.7	15.1	16.3
127.5	10.7	17.7		290	11.3	15.6	16.9	166	12.9	12.4	
132.5	11.0	17.8		300	11.3	14.2		173	10.8	15.3	
137.5	10.5	18.9		310	9.3	12.6		180	13.6	18.3	
142.5	10.4	17.4		320	10.5	13.3		187	15.1	20.2	18.4
152.5	10.4	16.6		332	11.2	13.6		194	13.2	17.6	
157.5	10.2	14.2		340	10.0	15.1	15.9	201	12.1	16.4	
162.5	10.7	17.3		350	10.8	14.3		208	11.5	16.0	
167.5	11.5	18.1		360	10.7	14.0	13.7	215	11.9	17.6	19.3
172.5	11.6	15.0		375	9.9	16.9		222	12.3	12.9	
177.5	12.6	17.5		385	11.9	15.3		229	12.0	10.8	14.0
182.5	12.3	14.7		395	15.3	12.5		236	13.9	10.9	
187.5	13.5	13.0		415	15.3	12.9					
190.0	11.7	17.3	18.2								
192.5	14.4	15.8									
196.5	15.6	13.8									
199.5	18.6	8.8									
202.5	17.3	9.1									
207.5	17.9	10.9									
212.5	16.3	7.7									
218.5	10.0	11.9									

[a] $\delta^{18}O_{cell}$ results were calibrated to standards of known values including waters BW-5-8 (−13.86‰), SWAT (−10.7‰), and TT071 (0.15‰) and cellulose IAEA-C3 (31.3–32.7‰), which were analyzed routinely with sediment samples. Mean $\pm$ 1σ (n) results were −13.8 $\pm$ 1.4 (10), −9.9 $\pm$ 0.9 (18), 0.1 $\pm$ 1.2 (4), and 32.7 $\pm$ 1.0 (8), respectively.

permafrost activity affected the peatland only intermittently during the mid- to late Holocene. Porewater δ^{18}O values from the lower 1.5 m of the section, which likely represents ice that formed sometime during this interval, vary between −18 and −16‰ (Fig. 3a). By 1500 yr B.P., accumulation of ice had elevated the surface sufficiently to cause a reduction in water and nutrient supply and the development of *Sphagnum*-dominated bog vegetation.

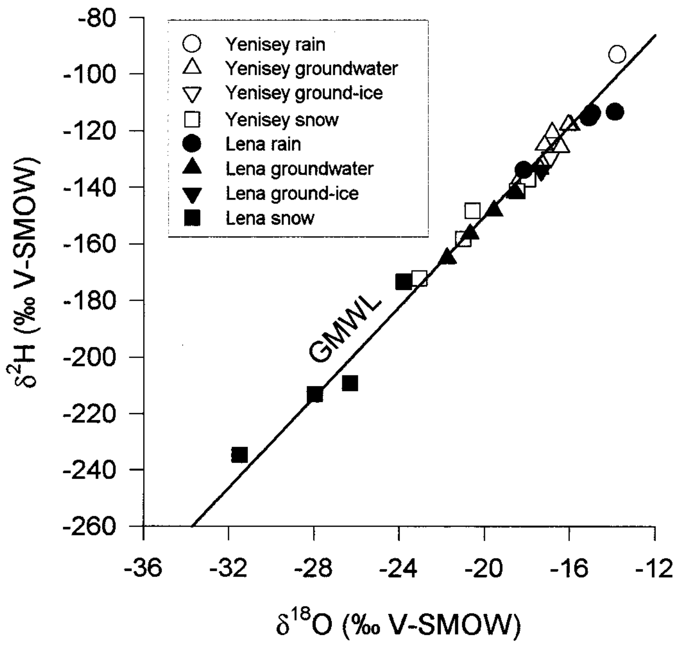

FIG. 2. Isotopic data obtained from samples of precipitation, groundwater, and ground-ice from near treeline, lower Yenisey and Lena Rivers, northern Russia. Data plot along the Global Meteoric Water Line (GMWL: $\delta^2H = 8\,\delta^{18}O + 10$; Craig, 1961).

At the peatland in the forest of the lower Lena River (Fig. 1), a shallow lake or open-water wetland existed at the site around 7000 yr B.P. (Jasinski *et al.*, 1998; Fig. 3b, Table 3). By 5000 yr B.P., this had evolved into a fen characterized by *Carex*, *Comarum palustris*, and *Drepanocladus*. Similar to the Yenisey site, permafrost activity is thought to have contributed to the development of a *Sphagnum*-dominated fen. However, the timing of ice accretion at this site is more tightly constrained, as considerable accumulation appears to have occurred between 5000 and 4500 yr B.P. (Jasinski *et al.*, 1998). Porewater $\delta^{18}O$ values in this part of the core span a narrow range (-20 to -19‰). A trend toward more ^{18}O-enriched values approaching -18‰ at ca. 4000 yr B.P. is consistent with increasing surface elevation (primarily due to ice accumulation) and greater influence of ^{18}O-enriched summer precipitation, relative to more ^{18}O-depleted groundwater input. Development of a high-centered polygon bog by 3000 yr B.P., which was likely elevated above the water table, created a drier peatland surface that has since limited permafrost accretion.

Notably, the porewater $\delta^{18}O$ records from the Yenisey and Lena peatlands provide independent estimates of $\delta^{18}O_P$ at about 6000–1750 and 5000–4500 yr B.P., respectively, which contribute key additional information for the interpretation of the lake sediment cellulose $\delta^{18}O$ records, as discussed below.

Isotope-Inferred Lake Paleohydrology

C/N values from Middendorf, Derevanoi, and Dolgoe Lakes are generally low (Fig. 4), indicating that the organic matter is derived primarily from aquatic sources (cf. Meyers and Lallier-Vergès, 1999). The larger uncertainty and variability in the $\delta^{18}O_{lw}$ values from Dolgoe Lake, compared to the Middendorf and Derevanoi $\delta^{18}O_{lw}$ records, are attributed to the greater inherent isotopic heterogeneity expected for sediments formed in a large, shallow lake having a small catchment area. To focus on longer-term isotopic signals, three-point running means have been drawn through the $\delta^{18}O_{lw}$ raw data.

Broadly similar trends occur in the upper part of the smoothed $\delta^{18}O_{lw}$ profiles, especially between the Middendorf and Dolgoe records, including small fluctuations between about 3500 and 2000 yr B.P. and a trend toward more ^{18}O-depleted values between 2000 and 1000 yr B.P., followed by ^{18}O-enrichment. Except for the shift toward more ^{18}O-depleted values at about 1000 yr B.P., these variations are less clearly visible in the more complacent Derevanoi $\delta^{18}O_{lw}$ profile. The trends over the past two millennia correspond to slightly higher bulk organic C/N ratios at Derevanoi and Dolgoe Lakes. However, low C/N ratios at Middendorf Lake suggest that this $\delta^{18}O$ oscillation is likely not related to local changes in the source of the organic matter. Differences in the smoothed profiles are also apparent, including ^{18}O-depletion prior to 4000 yr B.P. at Middendorf Lake and opposite $\delta^{18}O_{lw}$ oscillations at Derevanoi and Dolgoe Lakes between about 9000 and 5000 yr B.P.

Interpretation of $\delta^{18}O_{lw}$ records requires deconvolution of signals deriving from changes in the isotopic composition of source water, reflecting the integrated signature of surface and subsurface inflow and precipitation, from changing hydrologic factors (often primarily evaporative enrichment) that may subsequently modify the isotopic content of the lake water. To separate these hydrologic components from the $\delta^{18}O_{lw}$ records, we have attempted to constrain the $\delta^{18}O_P$ histories for the Yenisey and Lena study regions. The $\delta^{18}O_P$ profiles have been reconstructed from a combination of: (1) estimated changes in MAT during the early to mid-Holocene assuming the modern relation 0.59‰ $\delta^{18}O_P$/°C; (2) mid-Holocene paleo-$\delta^{18}O_P$ in-

TABLE 2

Modern Measured and Inferred Mean Annual Precipitation $\delta^{18}O$ ($\delta^{18}O_P$) and Mean Annual Temperature (MAT) for Stations and Regions in Northern Eurasia

Location	Measured/inferred $\delta^{18}O_P$	MAT (°C)
Moscow	-11.0^a	$+5.4^a$
Pechora	-14.7^a	-0.5^a
Amderma	-15.8^a	-6.7^a
Yenisey-treeline	-16.9^b	-10.7^d
Taimyr Peninsula	-20.1^c	-13.8^e
Lena-treeline	-19.6^b	-13.8^e

[a] Rozanski *et al.* (1993).
[b] This study (estimated from groundwater and ground-ice $\delta^{18}O$ data).
[c] Boike (1997).
[d] At Dudinka (Lydolph, 1977).
[e] Estimated from WMO (1981).

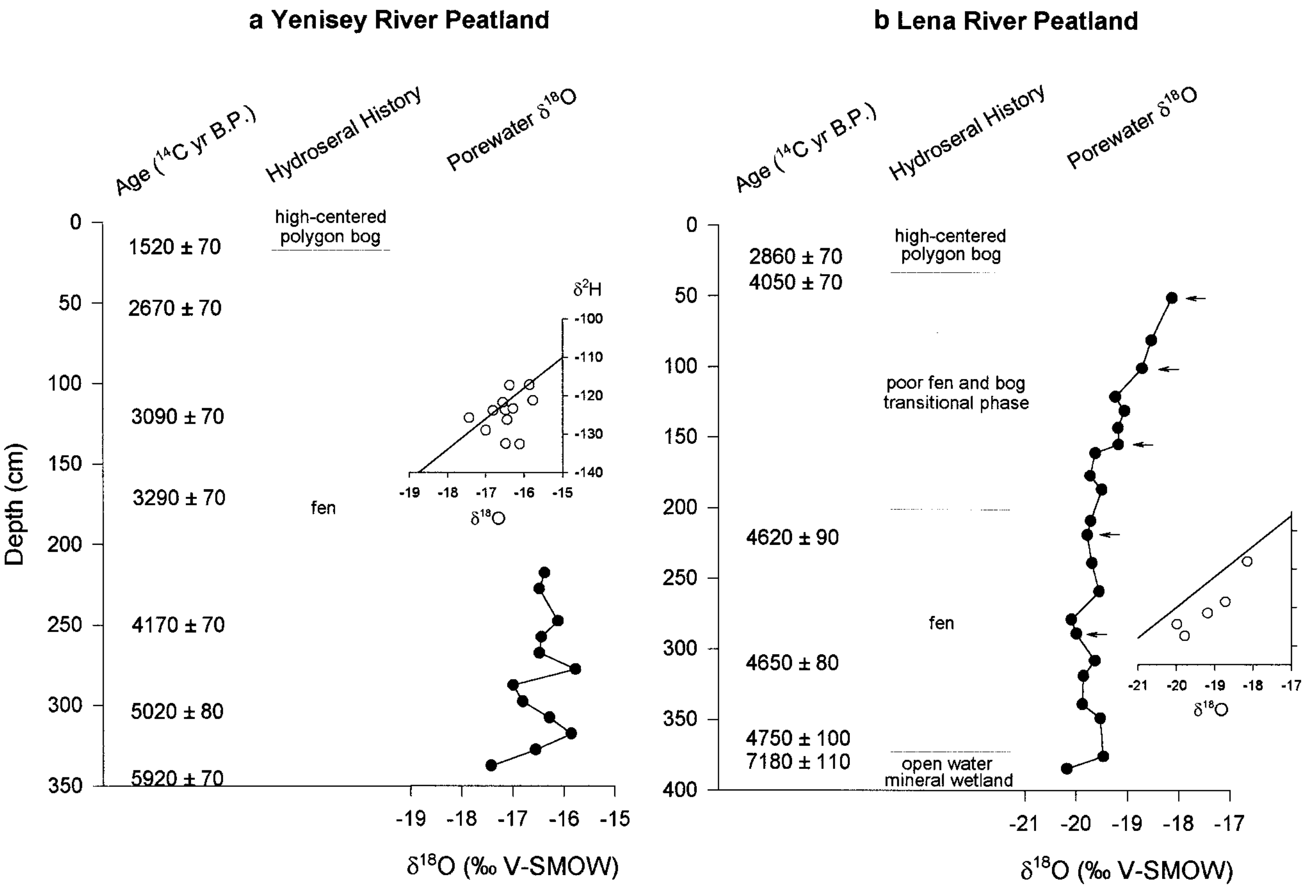

FIG. 3. Hydroseral history and porewater δ^{18}O for (a) the Yenisey River peatland (B. G. Warner and A. A. Andreev, unpublished data) and (b) the Lena River peatland (Jasinski *et al.*, 1998; see Table 3 for ^{14}C data). Insets show porewater δ^{18}O versus δ^2H (arrows in the Lena porewater δ^{18}O profile identify samples for which δ^2H values have been determined). Values plot close to the GMWL, indicating that porewater has not undergone significant evaporation.

ferred from peat porewater (described above); (3) minimum $\delta^{18}O_{lw}$ values during the mid- to late Holocene at Middendorf and Dolgoe Lakes, which likely reflect values that have not been strongly influenced by evaporation; and (4) modern estimated $\delta^{18}O_P$ for each site. The varying $\delta^{18}O_{lw}$ offsets from $\delta^{18}O_P$ values provide a measure of changing evaporative enrichment in the three lakes, which is a primary function of changing local hydrologic balance in response to shifting moisture regimes.

In the lower Yenisey River region, the $\delta^{18}O_{lw}$ record from Derevanoi Lake is largely controlled by $\delta^{18}O_P$, suggesting that this basin has likely been continuously open during the Holocene and buffered by inflow from its large catchment. However, subtle changes in the $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset provide evidence for millennial-scale changes in water balance. After 9000 yr B.P., $\delta^{18}O_{lw}$ values decline to about 7000 yr B.P., probably reflecting increasingly moist conditions. This corresponds with the arrival in the catchment of *Picea* (Clayden *et al.*, 1997), a genus that is favored by warm and moist conditions (Pisaric *et al.*, in press). A relatively wet regional climate during the early Holocene has also been inferred from other paleobotanical evidence (Velichko *et al.*, 1995, 1997; Koshkarova, 1995; Hahne and Melles, 1997). From 7000 to 1500 yr B.P., increasing $\delta^{18}O_{lw}$ values may be in part related to evaporation becoming a more important component of the water balance of Derevanoi and other lakes upstream due to increasing aridity. At about 1000 yr B.P., a shift to lower $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset indicates a return to moister conditions.

The Middendorf $\delta^{18}O_{lw}$ record shows large oscillations between minimal $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset, indicative of rapid through-flow, and high $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset, consistent with steady-state enrichment when the basin was seasonally closed, as indicated by isotope-mass balance calculations (Wolfe, 1997). In spite of this high-frequency hydrologic sensitivity, the changing evaporative ^{18}O-enrichment at Middendorf Lake appears to be largely compatible with the $\delta^{18}O_{lw}$ record from Derevanoi Lake, suggesting that common changes in moisture conditions have affected the long-term water balance for these lakes. For instance, dominantly closed-basin conditions occur at Middendorf Lake between 4000 and 2000 yr B.P. and persistent $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset occurs at Derevanoi Lake between 5500 and 1500 yr B.P. The Middendorf Lake record also shows a decline in the $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset between 2000 and 1000 yr B.P., similar to the Derevanoi Lake record.

TABLE 3
Radiocarbon Dates for Lake and Peatland Sediments

Depth (cm)	Material	Age (yr B.P.)[a]	Laboratory no.[b]
		Middendorf Lake	
80	aquatic moss	2500 ± 80	TO-4750
135	aquatic moss	3390 ± 80	TO-4751
196	wood fragments	4420 ± 80	TO-4749
218	wood fragments	4370 ± 60	TO-4348
		Derevanoi Lake	
19–20	wood	220 ± 50	TO-4756
65	aquatic moss	940 ± 50	TO-5179
70	wood	7580 ± 80[c]	TO-4755
114	*Picea* needles	3870 ± 90[c]	TO-5180
114	aquatic moss	1210 ± 50	TO-5181
132.5	wood	8210 ± 70[c]	TO-4752
172	twig	1720 ± 50	TO-5182
233	twig fragments	4490 ± 80	TO-5183
335	wood	7600 ± 90	TO-4753
373	wood	8880 ± 80	TO-4754
400	wood	9190 ± 130	WAT-2777
		Dolgoe Lake	
71.5	aquatic moss	1630 ± 60	TO-5719
117	aquatic moss	3780 ± 70	TO-5245
145	aquatic moss	5150 ± 70	TO-5255
192–194	moss and wood	7250 ± 80	TO-5720
229	aquatic moss	9830 ± 80	TO-5246
252.5	aquatic moss	10240 ± 120	TO-5721
287.5	aquatic moss	11520 ± 110	TO-5722
298	aquatic moss	12310 ± 100	TO-5247
		Peatland-Yenisey	
10–15	peat	1520 ± 70	WAT-2894
50–55	peat	2670 ± 70	WAT-2905
120–125	peat	3090 ± 70	WAT-2906
165–170	peat	3290 ± 70	WAT-2907
245–250	peat	4170 ± 70	WAT-2908
300–305	peat	5020 ± 80	WAT-2823
345–350	peat	5920 ± 70	TO-4907
		Peatland-Lena	
18–20	peat	2860 ± 70	WAT-2893
39–40	peat	4050 ± 70	WAT-3007
216–218	peat	4620 ± 90	WAT-2909
303–307	peat	4650 ± 80	WAT-3030
364–368	peat	4750 ± 100	WAT-3036
383–386	organic mud	7180 ± 110	WAT-2820

[a] Radiocarbon dates are corrected to a δ^{13}C value of −25‰ V-PDB and are reported with $\pm 1\sigma$ error.
[b] Laboratory designations: WAT, University of Waterloo Radiocarbon Laboratory; TO, University of Toronto Isotrace Laboratory.
[c] Rejected as reworked material (see Clayden *et al.,* 1997).

Middendorf Lake also displays a rapid shift from ^{18}O-depleted to ^{18}O-enriched values at the base at about 4000 yr B.P., corresponding to local treeline retreat (Wolfe *et al.,* 1999). Although relatively high C/N ratios occur at this horizon, sediment cellulose δ^{13}C values range between −34.2 and −29.7‰, which supports a dominantly aquatic origin (Wolfe *et al.*, 1999). This $\delta^{18}O_{lw}$ shift may simply reflect the crossing of a hydrological threshold (i.e., open- to closed-basin) at Middendorf Lake, resulting from the onset of drier conditions. Such a change is consistent with paleoclimatic conditions inferred from the physical sedimentology (A. A. Velichko, unpublished data) and carbon- and nitrogen-isotope stratigraphy of the lake-sediment core (Wolfe *et al.*, 1999). Alternatively, rapid lake-water through-flow prior to 4000 yr B.P. perhaps was fostered by thermokarst processes that released ^{18}O-depleted groundwater (melting ground-ice or buried glacier ice) during the initial formation of the basin. This latter hypothesis may also explain why several data points at this strata are more ^{18}O-depleted than the $\delta^{18}O_P$ inferred from the peat porewater at the Yenisey site.

At Dolgoe Lake, high $\delta^{18}O_{lw}$–$\delta^{18}O_P$ offset from before 9000 yr B.P. to about 6000 yr B.P. suggests dry conditions prevailed near the lower Lena River during the early Holocene, in contrast to the equivalent record from Derevanoi Lake. Additional evidence for increased aridity during the early Holocene in this region is provided by radiocarbon dating of sediments from a small closed-basin lake about 0.5 km from Dolgoe Lake. The age indicates that Holocene lacustrine deposition at this site did not begin until ca. 7000 yr B.P.[2] Wetter conditions developing after about 7000 yr B.P. are suggested by a decline in the diatom-inferred alkalinity from Dolgoe Lake, possibly resulting from increased supply of organic acid residues to the lake via runoff (Laing *et al.,* 1999) and the local presence of *Picea* (Pisaric *et al.,* in press). After 4000 yr B.P., close correspondence between the Dolgoe and Middendorf $\delta^{18}O_{lw}$ profiles suggests coherent submillennial-scale changes in moisture conditions in these two regions of northern Russia.

SUMMARY AND CONCLUDING COMMENTS

The combination of results from isotope analysis of modern meteoric water, frozen peat porewater, and lake sediment cellulose have yielded a provisional working model for Holocene paleohydrologic reconstruction in north-central Russia. These data indicate that at about 9000 yr B.P. the development of a strong moisture gradient occurred between the lower Yenisey and Lena Rivers. Moister conditions in the lower Yenisey River region during the early Holocene may have resulted from increased sea-surface temperatures (SSTs) in the Nordic Seas (Salvigsen *et al.*, 1992; Koç *et al.*, 1993; Lubinski *et al.*, 1999), resulting in increased cyclonic activity (MacDonald *et al.*, 2000). These conditions are thought also to have contributed to warming and boreal forest expansion in the high latitudes of

[2] A ^{14}C date of 6820 ± 70 yr B.P. (TO-6338) was obtained on aquatic moss sampled at 135.5–136.5 cm depth from a 178-cm sediment core from lake LS-11 (informal name; 71°53′N; 127°03′E). Organic lacustrine sediments occur to a depth of 145 cm and are underlain by stony, hard silt.

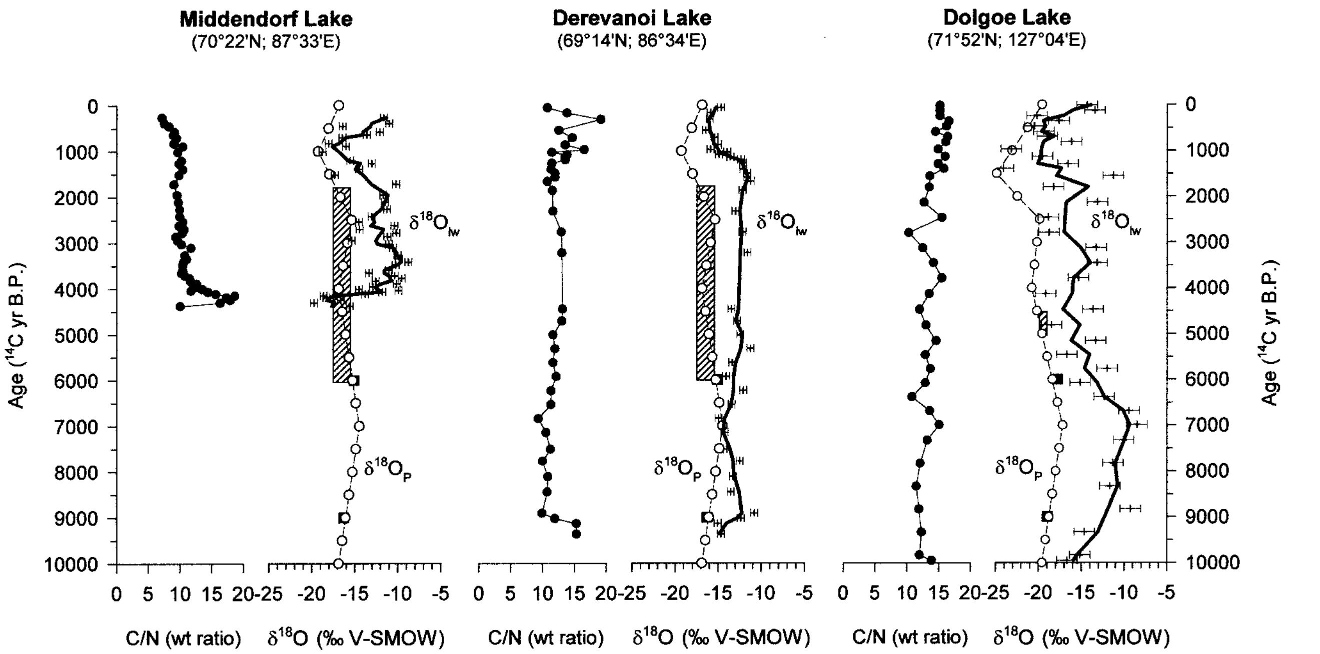

FIG. 4. Bulk organic C/N weight ratios and cellulose-inferred lake-water $\delta^{18}O$ ($\delta^{18}O_{lw}$) for the study lakes ($\delta^{18}O_{lw}$ values are calculated using a cellulose–water fractionation factor of 1.028; Edwards and McAndrews, 1989; Wolfe and Edwards, 1997). The bold line on $\delta^{18}O_{lw}$ profiles is a three-point running mean. Chronology is derived from linear interpolation between ^{14}C dates (see Table 3). Also shown are reconstructed mean annual $\delta^{18}O$ values in precipitation ($\delta^{18}O_P$) at 500-yr intervals. Early to mid-Holocene $\delta^{18}O_P$ values are constrained by using mean annual temperature (MAT) increases at 9000 (+1°C) and 6000 yr B.P. (+3°C) (estimated from Kutzbach *et al.,* 1993) and the modern relation 0.59‰ $\delta^{18}O_P$/°C, calculated relative to modern $\delta^{18}O_P$ from Table 2 (solid squares). Early Holocene maximum MAT is estimated to have occurred at about 7000 yr B.P., based on optimal conditions in the Greenland, Iceland, and Norwegian Seas (Koç *et al.,* 1993). Assuming negligible evaporative enrichment in the Derevanoi Lake $\delta^{18}O_{lw}$ record at 7000 yr B.P. places an upper limit on $\delta^{18}O_P$ at this horizon which corresponds to an increase of about 2.4‰ relative to present, equivalent to a MAT increase of 4°C. A similar 2.4‰ increase has been incorporated at the 7000 yr B.P. horizon in the $\delta^{18}O_P$ reconstruction for Dolgoe Lake. Peat porewater $\delta^{18}O$ values provide additional control on the $\delta^{18}O_P$ reconstructions during the mid-Holocene (hatched rectangles; see Fig. 3). Late Holocene trends in $\delta^{18}O_P$ are anchored by the most ^{18}O-depleted values at Middendorf and Dolgoe Lakes, which probably are not strongly altered by evaporation. Uncertainties in $\delta^{18}O_P$ values are on the order of ±1.2‰ for Middendorf and Derevanoi Lakes and ±2.5‰ for Dolgoe Lake.

Eurasia (Kerwin *et al.,* 1999; MacDonald *et al.,* 2000) and are consistent with modern circulation patterns that periodically affect this region (Lydolph, 1977; Rogers and Mosley-Thompson, 1995). A trend toward drier conditions beginning after 7000 yr B.P. and extending through the mid-Holocene in the lower Yenisey River region generally parallels reduced SSTs in the Nordic Seas. This pattern was caused by declining insolation (Koç *et al.,* 1993), leading to decreased influence of warm, moist North Atlantic air masses on western Eurasia (MacDonald *et al.,* 2000).

Causes of early Holocene warming, which perhaps extended east of the Yenisey River, include decreased albedo (resulting from high-latitude forest establishment in western Eurasia and reduced sea-ice cover), an emergent coastline, and increased summer insolation (MacDonald *et al.,* 2000). These factors likely contributed to the development of conditions drier than present in the lower Lena River region during the early Holocene. Early Holocene coastline paleogeography is presumed to be of minor influence on the Derevanoi Lake $\delta^{18}O_{lw}$ record, because: (1) this site is located more inland than Dolgoe Lake; and (2) Holocene changes in Eurasian coastline position were of greater significance east of the Taimyr Peninsula (see Fig. 1). Decreasing $\delta^{18}O_{lw}$ at Dolgoe Lake between 7000 and 5000 yr B.P. most likely reflects rising sea level, declining summer insolation, and onset of a cooler, more maritime climate.

During the last 3500 ^{14}C years, similar paleohydrologic records in both the lower Yenisey and Lena River areas suggest a weakened longitudinal moisture gradient. Reduced evaporative ^{18}O-enrichment between 2000 and 1000 yr B.P. at all three sites, perhaps coupled with regionally low MAT from 2000 to 1000 yr B.P. as suggested by the $\delta^{18}O_P$ reconstructions (Fig. 4), appears to be correlated with paleoclimatic evidence for cooler conditions elsewhere in the Russian arctic. To the northwest on Franz Josef Land and Svalbard, repeated glacier advances and retreats have been documented during the late Holocene (Lubinski *et al.,* 1999). A prominent glacier advance occurred at 1000 yr B.P. coincident with minimum $\delta^{18}O_P$ (and possibly MAT) reconstructed for the lower Yenisey River region. These climatic fluctuations do not appear to be associated with changes in summer SST near Svalbard (Lubinski *et al.,* 1999), nor is any strong relationship apparent between North Atlantic forcing and arctic temperature records for the past 400 yr

(Overpeck *et al.*, 1997). The forcing mechanism for these climatic changes remains uncertain.

Although our representations of millennial-scale precipitation $\delta^{18}O$ history for the two study areas require further testing and refinement, we note that the most reasonable reconstructions of $\delta^{18}O_P$ during the early Holocene, based on present data, are achieved using constant temporal $\delta^{18}O_P$–MAT relations, equivalent to the modern spatial relations in this large region. This contrasts with evidence from treeline areas in central Canada (Edwards *et al.*, 1996) and northern Sweden (Hammarlund and Edwards, 1998). In these regions substantial differences exist in temporal $\delta^{18}O_P$–MAT relations on the scale of centuries to millennia during the early Holocene. Such variations apparently are linked to reduced rain-out as moisture was transported across topographic barriers, perhaps enhanced by an increase in the summer–winter precipitation ratio.

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