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Carbon accumulation in peatlands of West Siberia over the last 2000 years

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[1] We use a network of cores from 77 peatland sites to determine controls on peat C content and peat C accumulation over the last 2000 years (since 2 ka) across Russia's West Siberian Lowland (WSL), the world's largest wetland region. Our results show a significant influence of fossil plant composition on peat C content, with peats dominated by *Sphagnum* having a lower C content. Radiocarbon-derived C accumulation since 2 ka at 23 sites is highly variable from site to site, but displays a significant N–S trend of decreasing accumulation at higher latitudes. Northern WSL peatlands show relatively small C accumulation of 7 to 35 kg C m^{-2} since 2 ka. In contrast, peatlands south of 60°N show larger accumulation of 42 to 88 kg C m^{-2}. Carbon accumulation since 2 ka varies significantly with modern mean annual air temperature, with maximum C accumulation found between −1 and 0°C. Rates of apparent C accumulation since 2 ka show no significant relationship to long-term Holocene averages based on total C accumulation. A GIS-based extrapolation of our site data suggests that a substantial amount (~40%) of total WSL peat C has accumulated since 2 ka, with much of this accumulation south of 60°N. The large peatlands in the southern WSL may be an important component of the Eurasian terrestrial C sink, and future warming could result in a shift northward in long-term WSL C sequestration.

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1. Introduction

[2] The rate of carbon (C) sequestration in northern ecosystems is an important element in understanding the global C cycle and global climate change. The land C sink of the northern extratropics, estimated to be about 1.5 Pg C a^{-1}, may have varied from 1 to 2 Pg C a^{-1} over recent decades [*Schimel et al.*, 2001; *Gurney et al.*, 2002; *Houghton*, 2007; *Stephens et al.*, 2007]. Variations in the rate of C sequestration in the northern latitudes has been largely attributed to changes in forest biomass [*Houghton*, 2005, 2007]. However, nonforest ecosystems may also play an important role in C exchange and storage, particularly those that represent long-term C sinks and very large C pools, including lakes and peatlands [*Dean and Gorham*, 1998].

[3] At present, northern peatlands cover about 4 million km^2, and store 270–450 Pg C [*Gorham*, 1991; *Turunen et al.*, 2002; *MacDonald et al.*, 2006], equivalent to as much as 1/3 of the world's soils C and about 2/3 of today's atmospheric C. Unlike forest ecosystems, northern peatlands sequester and store large amounts of dead C belowground over long time periods, typically about 130 kg C m^{-2} or 2.3 m of peat [*Gorham*, 1991] but often more than 180 kg C m^{-2} or 3.2 m of peat [e.g., *Sheng et al.*, 2004]. On the basis of total C stocks and the Holocene ages of deposits, peatlands globally have been estimated to be a long-term average atmospheric C sink of ~0.07–0.10 Pg C a^{-1} [*Clymo et al.*, 1998; *Dean and Gorham*, 1998]. However, it has been observed for more than a decade that interannual C balances at the site level can switch from net sink to net source in response to climate [e.g., *Shurpali et al.*, 1995]. Variations in climate can exert a control over short- and long-term rates of C sequestration through their influence on rates of peatland net primary productivity (NPP) and rates of litter/peat decomposition. Improved understanding of the factors, including climate, that affect rates of peat C sequestration and the distribution of these long-term fluxes is required to assess the role of peatlands in the global C cycle.

[4] Paleoenvironmental methods based upon study of C in radiocarbon-dated cores can reveal much about the nature of long-term C dynamics in northern peatlands over the Holocene [e.g., *Clymo et al.*, 1998; *Vitt et al.*, 2000; *Yu et al.*, 2003; *Gorham et al.*, 2007]. However, a number of pitfalls exist toward the interpretation of such histories. First, large changes in Earth-Sun geometry have reduced top-of-atmosphere summer insolation at northern high latitudes from the Holocene maximum at 9 ka (9000 calendar years before 1950 A.D.) to present values, e.g., −41 W m^{-2} at 60°N [*Berger and Loutre*, 1991]. Second, pollen and tree-fossil records show that climate and continental-scale vegetation

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distributions in areas such as northern Eurasia have changed substantially since 9 ka [*Bigelow et al.*, 2003; *MacDonald et al.*, 2000, 2008]. Third, a peatland's intrinsic C sink potential diminishes over time quite independently of external factors, including climate variation, if accumulated deep C has even a very slow turnover rate [*Clymo*, 1984; *Belyea and Baird*, 2006], which prevents direct comparison of recent and distant past apparent C accumulation rates. Fourth, postglacial lateral expansion of peatlands was rapid between 11.5 and 8.5 ka and has since been slower across much of the northern hemisphere [*MacDonald et al.*, 2006], which makes early- to mid-Holocene three-dimensional C accumulation over the landscape difficult to assess from single cores.

[5] The West Siberian Lowland (WSL) is the world's largest wetland region [*Fraser and Keddy*, 2005] and holds a large C pool as peat of ~70 Pg C [*Sheng et al.*, 2004]. Despite its significance, relatively few studies have investigated the long-term patterns of C accumulation and the sensitivity of WSL peatlands to past climate [*Botch et al.*, 1995], and these display substantial variation. For example, observations range from evidence for the near shutdown of C accumulation processes in arctic permafrost peatlands over the past few thousand years [*Peteet et al.*, 1998] to the recent rapid accumulation of taiga peat deposits that are now 5 to 10 m deep [e.g., *Turunen et al.*, 2002; *Borren et al.*, 2004]. In addition, peat C content has been assessed in a few site level studies only [e.g., *Turunen et al.*, 2001; *Borren et al.*, 2004] and WSL-wide C storage assessments have relied on assumed constant C content values [*Sheng et al.*, 2004].

[6] In this paper we address C accumulation patterns in WSL peatlands over recent millennia, with a focus on the last 2000 years (2 ka before present) to make baseline observations regarding present C dynamics and potential global warming impacts. The last 2000 years provides a time period that minimizes the potential confounding influences discussed above. By 2 ka insolation patterns were closely similar to today [*Berger and Loutre*, 1991]. Climate and broad-scale vegetation distributions in northern Eurasia were also closely similar to modern conditions [*MacDonald et al.*, 2000, 2008; *Bigelow et al.*, 2003]. On a continental scale, northern peatlands including the WSL were at their approximate modern extent by 2 ka [*Smith et al.*, 2004; *MacDonald et al.*, 2006]. At finer spatial scales, peatland expansion appears also to have been limited over recent millennia; basal ages from peatland margins at a site in southern WSL suggest that lateral expansion has been only a few meters since 3 ka [*Turunen et al.*, 2001]. We investigate four questions: (1) How does the C content of WSL peatlands vary across peat types, locations, and peat ages? (2) How much peat C has accumulated in the WSL since 2 ka across a network of study sites? (3) How does the accumulation since 2 ka at these sites compare to that of the very recent past (the last ~100 years) and over the Holocene (since 11 ka)? and (4) What geographic patterns are evident in rates of apparent C accumulation since 2 ka across the WSL?

2. Methods

2.1. Study Sites and Cores

[7] This study uses cores of organic deposits raised from 77 peatland sites across the WSL (Figure 1) that range in depth

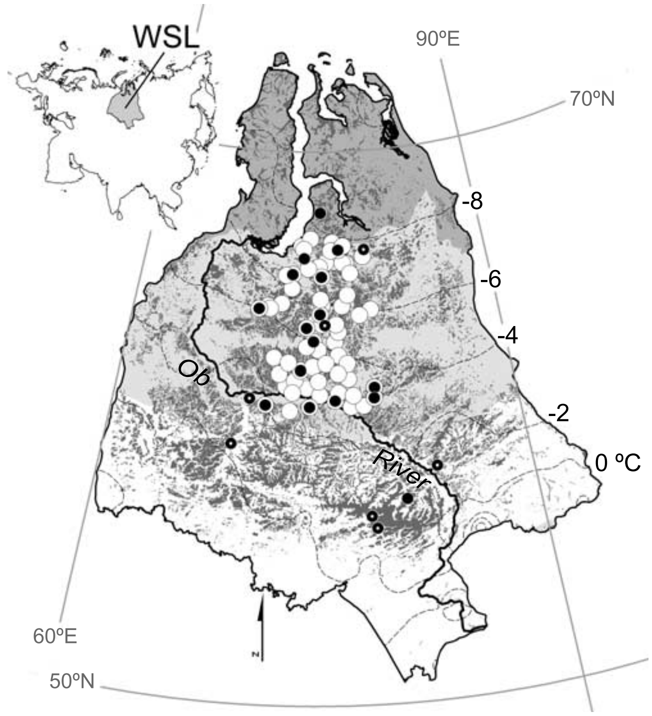

Figure 1. The West Siberian Lowland and location of peatland sites and cores used for study of peat C content (white circles) and C accumulation since 2 ka (black circles) and recent centuries (white dots). Shaded polygons show peatland extent [*Sheng et al.*, 2004]. Modern mean annual air temperature isotherms are shown by dashed lines (1931–2000) (K. Matsuura and C. J. Willmott, Arctic Land-Surface Air Temperature: 1930–2000 Gridded Monthly Time Series, version 1.01, 2004, Center for Climate Research, University of Delaware, Newark; available at http://climate.geog.udel.edu/~climate/html_pages/archive.html). The Continuous Permafrost Zone (darker shaded) and the Discontinuous, Isolated, and Sporadic Permafrost zones (lighter shaded) are from the Data and Information Working Group, International Permafrost Association (IPA Circumarctic Permafrost and Ground Ice Map. UNEP/GRID-Arendal. Digital Version: Circum-arctic Map of Permafrost and Ground Ice Conditions, version 1.0, 1998).

from 40 to 542 cm. The sites in this network, distributed widely across the WSL, were selected to span important environmental gradients that affect peatland development and C sequestration, i.e., temperature, precipitation, permafrost status, and peatland age. The basal ages, depths, and organic matter densities of cores from 71 of these sites are summarized by *Smith et al.* [2004] and *Sheng et al.* [2004]. Here we present an analysis of the C content using these 71 cores, as well as new data for 6 additional southern WSL sites and new radiocarbon data for 23 selected sites. Permafrost conditions are an important aspect of the WSL terrain, with the Continuous Permafrost Zone spanning 15%, and the Discontinuous, Isolated, and Sporadic Permafrost Zones spanning collectively 39% of the WSL by area (Figure 1). Cores were raised from permafrost peatlands (n = 40) using a modified Cold Regions Research and Engineering Lab

(CRREL) corer, and from nonpermafrost sites (n = 37) using a 5-cm diameter Russian side-cut corer. Cores were either subsampled in the field at 10-cm resolution (71 cores), or retained as complete profiles (6 cores).

2.2. Peat Type and Organic C Content

[8] Peat C content is affected by organic matter quality (e.g., varying amounts of cellulose and lignin-like compounds), and is thus influenced by fossil plant composition (e.g., the proportion of vascular plant versus bryophyte litter) and by changes in organic matter chemistry owing to decomposition. In our cores, the plant composition of peat was determined by semiquantitative macrofossil analysis for peat subsamples taken at 10-cm intervals. Each subsample was dispersed in distilled water and picked for fossil plants identified to the species level when possible. Relative abundances were quantified with the aid of a stereomicroscope, reference collections, and various taxonomic keys. We determined macrofossil-derived peat types by agglomerative cluster analysis using City-Block distances and average linkage [*Legendre and Legendre*, 1998] and named the groups after their dominant macrofossil component.

[9] To investigate other potential factors affecting peat C content, we further separated the macrofossil-based groups by geographical location (either north or south of 63°N on the basis of the bimodal latitude distribution of peatland area and peat C identified by *Sheng et al.* [2004]) and by age (younger or older half of each core). Five replicate peat samples were drawn at random from each macrofossil, depth, and latitude combination (20 samples per peat type, n = 120). These samples were analyzed for total C (TC), inorganic C (IC), and organic matter content (OM). The controls on variation in peat organic C (OC) were explored using ANOVA to investigate fixed effects (peat macrofossil type, depth, latitude) and all possible interactions with the PROC MIXED procedure in SAS 9 (SAS Institute, Cary, North Carolina, USA).

[10] Peat OC was determined by the indirect method [*Bisutti et al.*, 2004; *Nelson and Sommers*, 1996] by subtraction of IC from TC. Subsamples of bulk peat for TC analysis were dried at 70°C for 24 h, ground, and measured on a Carlo Erba NA1500 elemental analyzer at the Institute of Ecology Stable Isotope Laboratory at the University of Georgia. Peat OM was determined by loss-on-ignition analysis (LOI) at 550°C for 4 h [*Heiri et al.*, 2001; *Nelson and Sommers*, 1996] and peat IC content was determined on these same samples by LOI at 950°C for 2 h (IC = $LOI_{950} \times 0.273$ on the basis of the stoichiometry of C in carbonate and CO_2). We confirmed the LOI_{950}-derived IC content on a subset of samples (three random samples from each of the six macrofossil-derived peat groups) using a Thermo EA1112 elemental analyzer (EA) at the ^{14}CHRONO Centre at Queen's University Belfast and measuring C on paired samples of HCl-fumigated and untreated homogenized peat [*Harris et al.*, 2001]. In this paper OC values are expressed per mass of organic matter, i.e., on an ash-free basis.

2.3. Deposition Age and Peat C Accumulation

[11] Carbon sequestration since 2 ka in 23 sites (Figure 1) was assessed by determining peat deposition age by the ^{14}C-AMS dating of plant macrofossils or bulk peat. The 2000-year level was targeted by selecting two peat subsamples estimated to bracket the 2 ka level on the basis of peatland basal age [*Smith et al.*, 2004], deposit thickness, and an assumed linear accumulation rate.

[12] Peat subsamples were dispersed in distilled and deionized water overnight then separated into large and small size fractions by wet sieving at 150 μm. Photosynthetic tissue of plant macrofossils was handpicked from the large fraction, taking care to exclude unrecognizable organic debris, roots, and fungal hyphae. Macrofossil remains were cleaned of carbonates and exogenous humic acids by standard Acid-Base-Acid pretreatment at 65°C. The samples were dried, sealed under vacuum in quartz tubes containing CuO, and combusted at 900°C for 4 h. The purified CO_2 gas sample was graphitized for ^{14}C-AMS measurement at the Lawrence Livermore National Laboratory Centre for Accelerator Mass Spectrometry. Where sample size permitted, an aliquot was measured for $\delta^{13}C$ at the Stable Isotope Laboratory at UC Davis. These $\delta^{13}C$ values (−30.3 to −23.7‰) were used to correct ^{14}C measurements for biological and physical discrimination effects rather than using assumed values in the calculation of the conventional radiocarbon age [*Stuiver and Polach*, 1977]. For small samples, a mean value of n = 91 $\delta^{13}C$ measurements (−26.8 ± 1.4‰, ±SD) was used from our other dated WSL peat macrofossil samples. All radiocarbon ages were calibrated to calendar ages using the IntCal04 calibration curve [*Reimer et al.*, 2004a] with CALIB 5.0.1 [*Stuiver and Reimer*, 1993].

[13] The 2 ka level was estimated by linear interpolation between ^{14}C-dated core levels. In cases where the dates bracketing the 2 ka age were 2000 years apart and separated by more than 50 cm, additional ^{14}C ages were obtained from intervening samples. Net C sequestration since 2 ka was then determined as the mass of OC accumulated above the 2 ka level in each core. Apparent rates of C accumulation were calculated as the quotient of accumulated C mass and the duration of the accumulation period. In each of the 23 cores, OM content was determined at either 2-cm or 10-cm resolution. We assigned a C content to each level on the basis of its macrofossil composition and the results of the C content analysis described in section 2.2. The total Holocene C accumulation (since peatland initiation) was calculated from the total OC mass above basal peat.

[14] Post-2-ka apparent rates were compared to C accumulation of the last ~100 years by determining the deposition age of near-surface peat at seven sites. Deposition ages were determined by ^{14}C dating *Sphagnum* plant remains as described above. Owing to the more recent timescale considered for surface samples, we calculate the near-surface rates on the basis of calendar years before the year of field collection rather than years before AD 1950.

3. Results

3.1. Variation and Controls of Peat C Content

[15] The peat macrofossil data set consists of 71 cores and a total of 1157 macrofossil assemblages that includes 61 bryophyte species (25 *Sphagnum* and 36 non-*Sphagnum* species) and 26 vascular plant species (9 woody and 17 non-woody species), as well as lichens and unidentifiable

remains. To generalize these data for identifying broad trends in peat C content, we condensed all taxa into 14 macrofossil types: *Sphagnum* spp., sedges (*Eriophorum* and *Carex* spp.), *Scheuchzeria*, *Thelipteris*, *Equisetum*, other herbs, brown mosses (family Amblystegiaceae), other bryophytes, lichens, Ericaceae, large shrub remains (*Betula*), tree remains (*Larix*, *Pinus*), and unidentifiable remains.

[16] Cluster analysis of the condensed macrofossil data showed that the four groups with the largest group membership contained 83% of all assemblages; *Sphagnum* (38%), sedge (18%), *Scheuchzeria* (15%) and herbaceous peat (other herbs; 12%). The remaining samples were dominated mainly by either woody shrub remains (9%) or brown moss remains (4%). These six groups therefore comprise the basis of our peat classification for the purpose of investigating plant composition–peat C relations.

[17] The TC content of peat in our cores varied between 41.5 and 62.8% with a mean of 52.0 ± 0.3% (±SE). Peat IC content measured by LOI_{950} was very low and often negligible, with a maximum value of 0.27% and a mean of 0.06 ± 0.01%. The subset of peat IC measured by EA supports these results and IC was present in a measurable amount in only one sample of brown moss peat at 0.14%. Peat OC content varied between 0.44 and 0.64 g-OC g-OM^{-1}, which yielded an overall mean value of 0.54 ± 0.003 g-OC g-OM^{-1}. This mean value is similar to that used by *Sheng et al.* [2004] (52%), and supports the conservative peat C estimate of 70 Pg C for the entire WSL.

[18] Three-way ANOVA revealed that peat macrofossil type had a significant influence on OC (F = 14.02, P < 0.0001), but neither latitude nor age had a significant influence (P = 0.6244 and P = 0.7554, respectively). All interaction terms were nonsignificant. One-way ANOVA of peat C content (F = 7.79, P < 0.0001), followed by a between-group comparison analyzing patterns in least squares means following Bonferroni adjustment, revealed that *Sphagnum* peat is of significantly lower C content than other groups (Figure 2). The mean OC content of *Sphagnum* peat was 0.51 ± 0.005 g-OC g-OM^{-1} and non-*Sphagnum* peat averaged 0.55 ± 0.003 g-OC g-OM^{-1} likely owing to relative differences between litter types in the abundance of various plant biopolymers. These values are used to refine our subsequent C estimates for the cores.

3.2. Peat C Accumulation Since 2 ka

[19] The depth of peat accumulated since 2 ka ranged from as shallow as 11 cm to as deep as 272 cm (Table 2). A clear geographic pattern is evident in the depth of accumulation since 2 ka with southern sites showing the greatest rates of net accumulation, and a significant linear relationship with latitude (P < 0.0001, r^2 = 0.75). In general, sites in the northern WSL show peat of mid-Holocene age (as old as 5500 calendar years BP) within the top 50 cm of the current peatland surface (Table 1). In contrast, at a southern taiga site 285 cm of net peat deposition has occurred since about 2060 calendar years BP (Site SIB04; see Table 1). Peat depths and ages reported by other authors for southern WSL sites are in general agreement with our observed patterns (Table 2), including 110-cm-deep peat of 2 ka age as reported by *Turunen et al.* [2001] and ~2-m-deep peat of 2 ka age reported by

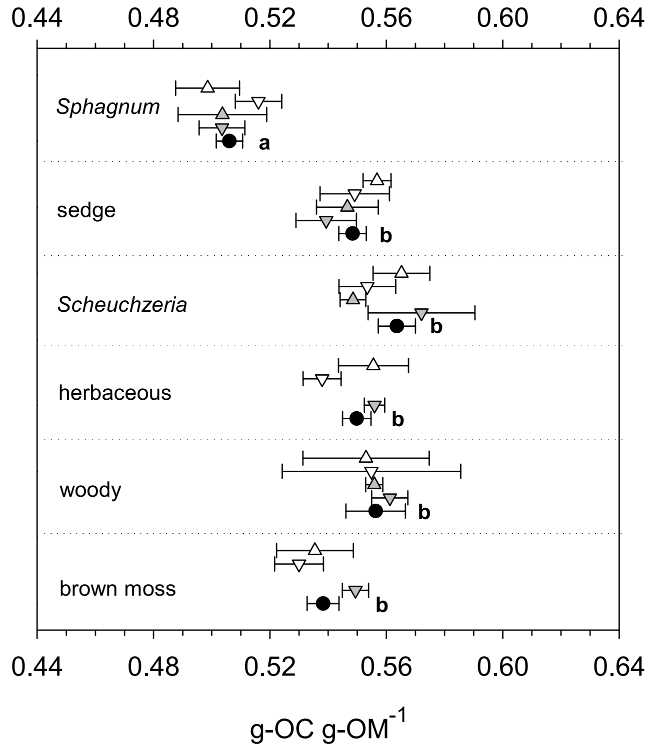

Figure 2. Mean organic C (OC) content of West Siberian Lowland (WSL) peat (see Figure 1) shown by fossil plant composition, latitude, and age. Error bars indicate ±1 SE. Peat from the younger half of cores is shown by upward pointing triangles, peat from the older half is shown by downward pointing triangles, peat from north of 63°N is shown as open symbols, and peat from south of 63°N is shown as filled symbols. Means for macrofossil groups that include all locations and depths are shown by solid circles; the means that are not significantly different (ANOVA, P < 0.05) are shown by the same letters.

Borren et al. [2004]. Across all sites, peat deposition since 2 ka accounts for between 5% and 80% of total peat depth (Table 2).

[20] The net C accumulated since 2 ka averages 34 ± 4 kg m^{-2} (a peat depth of 86 ± 15 cm; ±SE) but varies between sites by an order of magnitude (8–88 kg m^{-2}; see Figure 4). Rates of apparent long-term average C accumulation since 2 ka range from 3.8 to 44.1 g C m^{-2} a^{-1}. However, apparent rates do not reflect the true rate of C accumulation because they neglect C lost from deep peat layers by slow long-term decomposition. *Clymo et al.* [1998] suggested a plausible value for the proportional decay rate of deep peat of 3.7 × 10^{-5} a^{-1} for northern peatlands based on data from 310 peatland sites in Finland. Modeled losses based on this value and the C mass deposited before 2 ka suggest possible values between 1.7 and 10.7 kg C m^{-2} in our sites (Figure 2). If this decay rate is roughly accurate and if turnover of deep peat C in permafrost peatlands is negligible these 23 WSL sites have acted as historic long-term net sinks of atmospheric C, albeit of likely variable strength over time within individual sites, since 2 ka.

Table 1. ^{14}C-AMS Measurements of Peat Macrofossil Age at 23 Sites, Listed by Decreasing Latitude

Site	Location (dd)	Total Depth (cm)	Lab Number	Dated Depth (cm)	Material Dated	δ^{13}C[d] (‰ PDB)	^{14}C Age (years BP)	Median Age (calendar years BP[e])	2σ Age Range (calendar years BP[e])
E115[a]	67°48′34″N, 75°26′04″E	99	BETA-168558	20–21	Bulk peat	−26.4	440 ± 40	550	330–540
			BETA-168599	55–56	Bulk peat	−25.6	5800 ± 40	6600	6490–6720
E110[a]	66°28′11″N, 76°59′39″E	223	CAMS-107574	28–30	*Sphagnum* remains	*−26.8*	360 ± 35	410	320–500
			CAMS-105578	49–50	*Sphagnum* remains	−28.0	2275 ± 25	2320	2160–2350
E113[a]	66°26′59″N, 79°19′24″E	329	CAMS-105759	47–48	Cyperaceae epidermis	−26.7	1885 ± 30	1830	1730–1890
			CAMS-105760	78–79	Bryophyte remains	−23.8	2400 ± 40	2440	2340–2700
P131[a]	66°09′59″N, 73°59′19″E	127	CAMS-119798	23–24	Cyperaceae epidermis	*−26.8*	145 ± 30	150	0–280
			BETA-168564	40–41	Bulk peat	−27.0	5690 ± 50	6480	6320–6640
D122[a]	65°34′59″N, 73°00′20″E	108	CAMS-105608	18–20	*Sphagnum* remains	−29.2	1000 ± 30	930	800–970
			CAMS-105579	28–30	Cyperaceae epidermis	−30.3	2075 ± 30	2050	1950–2130
E119[a]	65°29′59″N, 75°30′08″E	312	CAMS-105580	49–50	Cyperaceae epidermis	*−26.8*	2940 ± 30	3110	3000–3210
			CAMS-105581	79–80	Cyperaceae epidermis	*−26.8*	6055 ± 30	6910	6800–6990
D127[a]	64°18′25″N, 70°17′47″E	228	CAMS-107576	18–20	Brown moss remains	*−26.8*	3355 ± 30	3600	3480–3690
			CAMS-105609	48–50	Bryophyte remains	−27.7	4735 ± 30	5500	5330–5580
G136[a]	64°08′51″N, 75°21′39″E	166	CAMS-107582	28–30	*Sphagnum* remains	*−26.8*	115 ± 30	120	0–270
			CAMS-105347	49–50	Cyperaceae epidermis	−27.6	2470 ± 30	2570	2370–2710
G137[a]	63°45′01″N, 75°45′58″E	177	BETA-168560	20–21	Bulk peat	−25.9	730 ± 40	680	570–730
			BETA-168561	60–61	Bulk peat	−23.7	5240 ± 40	5990	5920–6180
N015[b]	63°39′00″N, 74°16′09″E	213	CAMS-107577	27–30	*Sphagnum* remains	*−26.8*	1735 ± 35	1650	1550–1720
			CAMS-105588	58–60	Cyperaceae epidermis	−27.9	6980 ± 35	7820	7710–7930
N001[b]	63°09′40″N, 74°49′23″E	242	CAMS-105589	37–40	Ericaceae leaves	−27.1	4755 ± 30	5520	5330–5590
			CAMS-107581	60–63	*Sphagnum* remains	*−26.8*	6805 ± 25	7640	7590–7680
S009[b]	62°07′22″N, 73°50′28″E	225	CAMS-105590	35–40	*Sphagnum* remains	−28.5	885 ± 40	800	730–920
			CAMS-105591	65–70	*Sphagnum* remains	−28.5	1110 ± 30	1010	940–1070
V034[b]	61°28′03″N, 79°27′36″E	327	CAMS-105593	45–48	*Sphagnum* remains	−25.7	postbomb: F^{14}C = 1.0035 ± 0.0034		
			CAMS-105594	75–80	*Sphagnum* remains	−27.9	2415 ± 35	2440	2350–2700
V039[b]	61°05′22″N, 79°22′50″E	315	CAMS-105596	75–80	Cyperaceae epidermis	−27.0	1985 ± 30	1930	1880–2000
			CAMS-105736	145–148	Cyperaceae epidermis	−26.6	4800 ± 30	5520	5470–5600
SIB02[c]	61°03′19″N, 70°03′31″E	280	CAMS-119904	95–96	*Sphagnum* remains	*−26.8*	1755 ± 25	1660	1570–1730
			BETA-165121	266–276	Bulk peat	−27.2	7720 ± 50	8500	8420–8590
V026[b]	61°01′43″N, 76°28′06″E	447	CAMS-105599	65–70	Cyperaceae epidermis	−26.4	925 ± 30	850	770–920
			CAMS-105600	115–120	*Sphagnum* remains	−28.3	2470 ± 30	2570	2370–2710
S022[b]	60°50′24″N, 71°15′20″E	360	CAMS-105601	75–80	*Sphagnum* remains	−25.2	970 ± 35	860	790–950
			CAMS-105602	115–120	Cyperaceae epidermis	−26.8	2185 ± 30	2240	2120–2310
V038[b]	60°48′14″N, 74°32′29″E	308	CAMS-105574	65–70	Cyperaceae epidermis	−28.5	410 ± 30	480	330–520
			CAMS-105603	95–98	*Sphagnum* remains	−27.8	1160 ± 40	1080	970–1180
SIB01[c]	59°21′36″N, 68°59′05″E	180	CAMS-119903	134–135	*Sphagnum* remains	*−26.8*	1115 ± 30	1020	940–1080
			BETA-165120	173–183	Bulk peat	−27.0	6090 ± 90	6970	6740–7240
SIB06[c]	58°26′09″N, 83°26′03″E	395	CAMS-105740	179–180	Cyperaceae epidermis	−26.3	1765 ± 30	1670	1570–1810
			CAMS-107823	188–189	Ericaceous leaves	*−26.8*	2425 ± 35	2460	2350–2700
SIB05[c]	57°21′15″N, 81°09′52″E	165	CAMS-105607	95–96	*Sphagnum* remains	−27.6	1175 ± 30	1100	990–1180
			CAMS-119826	133–134	Cyperaceae epidermis	−27.7	2430 ± 30	2460	2350–2700
SIB04[c]	56°48′14″N, 78°44′13″E	400	CAMS-105733	258–259	*Sphagnum* remains	−29.0	1995 ± 30	1940	1880–2000
			CAMS-119838	284–285	*Sphagnum* remains	−26.8	2085 ± 35	2060	1950–2150
SIB03[c]	56°21′19″N, 79°04′08″E	300	CAMS-119823	223–224	Brown moss remains	−26.2	1980 ± 30	1930	1870–1990
			CAMS-119824	255–256	Brown moss remains	−27.0	2245 ± 30	2230	2160–2340

[a]Collected in 2000.
[b]Collected in 1999.
[c]Collected in 2001.
[d]The δ^{13}C values were measured on ^{14}C gas sample aliquots except for values shown in italics that are assumed on the basis of the mean of 91 measurements from other West Siberian Lowland (WSL) peat ^{14}C gas samples.
[e]Calibrated with the IntCal04 calibration curve [*Reimer et al.*, 2004a].

Table 2. Peat C Accumulated Since 2 ka and Since Initiation From Cores Extracted From 23 Peatland Sites in West Siberia

Site	2 ka Depth (cm)	Proportion Total Depth	Since 2 ka: C Mass (kg m^{-2})	Since 2 ka: Proportion Total C	Apparent Rate of C Accumulation (g C m^{-2} a^{-1})	Since Initation (Total): Basal Age (calendar years BP[a])	Since Initation (Total): Apparent Rate of C Accumulation (g C m^{-2} a^{-1})
E115	29	0.29	16.60	0.34	8.3	9080	5.4
E110	46	0.21	32.24	0.15	16.1	10210	21.4
E113	56	0.17	29.84	0.16	14.9	8360	21.9
P131	29	0.23	19.48	0.19	9.7	9970	10.4
D122	29	0.27	33.48	0.31	16.7	8500	12.8
E119	32	0.10	20.42	0.08	10.2	9760	24.9
D127	11	0.05	7.59	0.05	3.8	10470	15.6
G136	45	0.27	34.85	0.23	17.4	7840	19.1
G137	30	0.17	19.90	0.18	10.0	9330	11.6
N015	30	0.14	17.85	0.18	8.9	9580	10.2
N001	14	0.06	7.29	0.05	3.6	10950	13.6
S009	87	0.39	29.37	0.29	14.7	8720	11.7
V034	72	0.22	28.49	0.17	14.2	10330	16.1
V039	79	0.25	47.24	0.27	23.6	10930	16.8
SIB02	104	0.37	31.67	0.24	15.8	8500	15.6
V026	101	0.23	30.47	0.18	15.2	9650	17.1
S022	111	0.31	30.78	0.26	15.4	7170	16.6
V038	126	0.41	36.37	0.31	18.2	7540	15.4
SIB01	142	0.79	41.90	0.62	21.0	6970	9.7
SIB06	183	0.46	55.01	0.31	27.5	8610	20.5
SIB05	121	0.73	41.52	0.57	20.8	4240	17.1
SIB04	272	0.68	88.24	0.65	44.1	3770	35.9
SIB03	231	0.66	73.82	0.76	36.9	2770	34.9

[a]From *Smith et al.* [2004].

[21] In Arctic WSL peatlands, C accumulation since 2 ka has been limited to values as low as ~8 kg m^{-2}. However, between-site variation is substantial in the northern portion of WSL, and some northern permafrost sites have C accumulations similar to some nonpermafrost sites much further south (Figure 2 and Table 2). Southern WSL sites show the highest C accumulation, with particularly high rates in the very large peatlands south of 61°N including the Great Vasyugan Bog complex.

3.3. C Accumulation Since 2 ka and Over the Recent Past

[22] The pattern of peat deposition and C accumulation over the last ~100 years follows the broad regional pattern in C accumulation since 2 ka. At northern WSL sites in permafrost peatlands, near-surface peat is hundreds of years old within the top 16 cm (Table 3). Near-surface apparent C accumulation rates are on the same order of magnitude as rates since 2 ka and since initiation. In contrast to Arctic sites, profiles from the southern taiga of the WSL show young C (postbomb ^{14}C values) in peat at 24 to 58 cm below the surface. Average apparent C accumulation rates in near-surface peat varied between 15 and 32 g C m^{-2} a^{-1} in northern sites and between 42 and 269 g C m^{-2} a^{-1} in southern sites (Table 3). This latitudinal pattern suggests that the relatively fast rates of apparent C accumulation since 2 ka in the southern WSL have persisted over recent centuries and decades.

Table 3. Near-Surface Peat Macrofossil Ages and Recent C Accumulation at Seven Sites

Site	Depth (cm)	Lab Number	δ^{13}C (‰)	^{14}C Age or F^{14}C[a]	Material Dated	Age[b] (calendar year BP or years AD)	Section (cm)	Apparent Rate of C Accumulation[c] (g C m^{-2} a^{-1})
E113	7–8	CAMS-119827	−26.1	205 ± 30	*Sphagnum* remains	180 (0–300)	0–8	17 (11–79)
	15–16	CAMS-119828	−26.0	350 ± 25	*Sphagnum* remains	390 (320–470)	8–16	16 (7–172)
G137	7–8	CAMS-119829	−27.8	280 ± 35	*Sphagnum* remains	370 (290–430)	0–8	15 (14–19)
	15–16	CAMS-119830	−26.7	480 ± 30	*Sphagnum* remains	520 (510–530)	8–16	32 (20–60)
SIB01	45–46	CAMS-105604	−25.9	*1.7824 ± 0.0054*	Sphagnum remains	1963–1965 AD	0–46	256 (250–264)
SIB06	24–25	CAMS-119833	−25.0	*1.5055 ± 0.0047*	Sphagnum remains	1963 or *1970–1972* AD	0–25	143 (138–148)
	31–32	CAMS-119834	−25.1	*1.2730 ± 0.0040*	Sphagnum remains	*1959–1962* or 1979–1981 AD	25–32	148 (114–185)
SIB05	57–58	CAMS-105606	−28.0	*1.0799 ± 0.0038*	Sphagnum remains	1956–1957 AD	0–58	269 (263–269)
SIB04	42–43	CAMS-119831	−27.1	245 ± 30	*Sphagnum* remains	290 (150–310)	0–43	45 (43–77)
SIB03	44–45	CAMS-119835	−23.6	145 ± 30	*Sphagnum* remains	150 (0–270)	0–45	42 (26–169)

[a]Postbomb F^{14}C values are shown in italics.
[b]Uncertainty in the calibrated age (1σ age range) is shown in brackets. Postbomb ages were calibrated using CALIBomb [*Reimer et al.*, 2004b] and the NH Zone 1 data [*Hua and Barbetti*, 2004]. Bimodal calibrated age distributions for SIB06 were resolved on the basis of superposition of samples within the core, and the most likely age is italicized.
[c]On the basis of median age. The range shown in brackets reflects uncertainty introduced by ^{14}C measurement and calibration.

3.4. C Accumulation Since 2 ka and Over the Holocene

[23] Long-term apparent C accumulation since peatland initiation, i.e., the ratio of total OC mass and basal age, in the WSL varied from 5.4 to 35.9 g C m^{-2} a^{-1} (Table 2). These values are within the range of other values reported for northern peatlands [e.g., *Kobak et al.*, 1998; *Vardy et al.*, 2000], and are similar in range to the rates of apparent C accumulation since 2 ka (3.6 to 44.1 g C m^{-2} a^{-1}; see Table 2). However, more than half of our sites show slower rates since 2 ka than the Holocene average at these same sites (Figure 3), mainly in the northern WSL (Table 2). Only a few show substantially faster rates since 2 ka. Site-by-site comparison of rates reveals that there is poor agreement and no significant relationship between apparent rates of total Holocene C accumulation and those since 2 ka (Figure 4; $P = 0.19$ for all 23 sites, $P = 0.47$ for 21 sites that initiated before 4 ka BP). This suggests that geographical areas or individual peatlands that were substantial C sinks during the mid and/or early Holocene have not necessarily been the location of the greatest C accumulation over recent millennia. The most substantial outliers, i.e., sites with the greatest Holocene average C accumulation and the lowest relative C accumulation since 2 ka, occur between 63 and 65.5°N. However, variability between peatland sites is high, with other peatland sites within this region showing little difference between apparent rates of C accumulation over the entire Holocene and since 2 ka (Table 2).

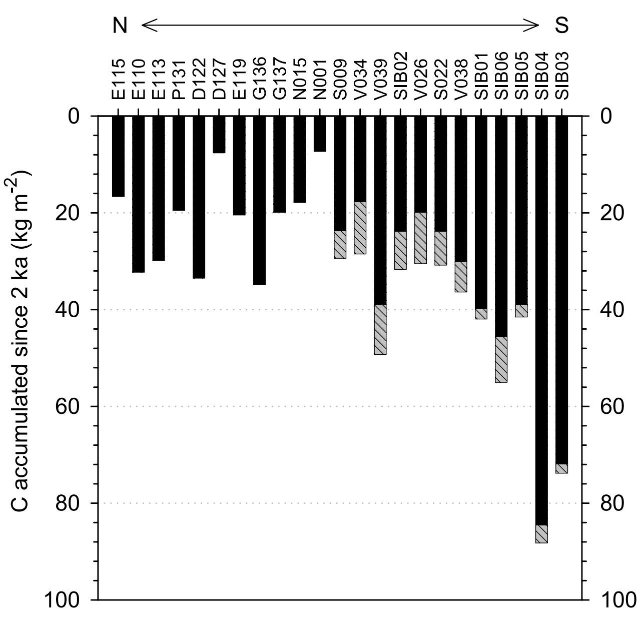

Figure 3. Organic C accumulated since 2 ka in cores from 23 WSL sites listed by latitude. Stacked bars (solid plus hatched) show C mass above the 2 ka level in each profile. Hatched bar sections show possible deep C mass losses since 2 ka by slow, long-term mineralization from peat older than 2 ka using a proportional turnover rate of 3.7×10^{-5} a^{-1} [*Clymo et al.*, 1998] or are assumed negligible for permafrost sites.

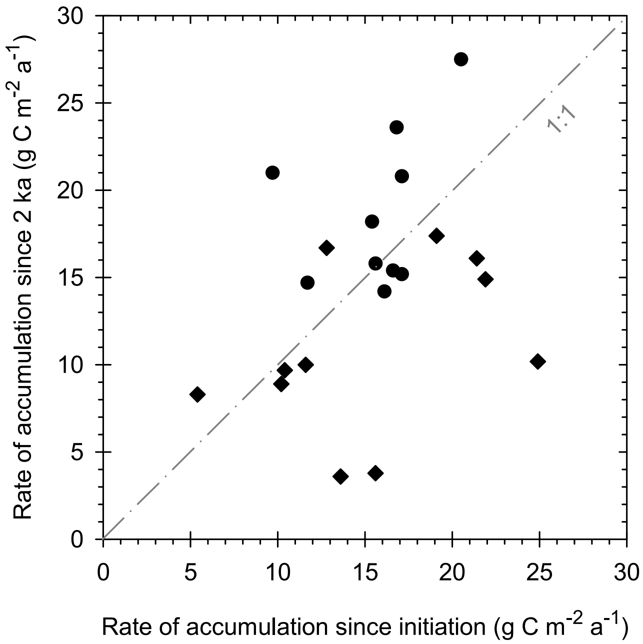

Figure 4. Relationship between average Holocene apparent rates of C accumulation (whole-core C and basal age) and rates since 2 ka for sites older than 4 ka BP. Diamonds indicate permafrost sites, and circles indicate nonpermafrost sites.

3.5. Estimation of Total WSL Peat C Accumulated Since 2 ka

[24] We used a simple model relating temperature and peat C accumulation in an exploratory analysis to extrapolate C accumulation since 2 ka across the WSL. The internal processes that determine C accumulation in peatlands, i.e., long-term balance between NPP and litter addition, and total losses by peat decomposition (mediated by microbial enzyme activity), are influenced directly by temperature [*Clymo et al.*, 1998] or indirectly by the effect of temperature on summer evapotranspiration and water table hydrology. Estimated peat accumulation since 2 ka does not reflect the pattern of total peat accumulation at our sites (Figure 4), which is in contrast to the significant nonlinear relationship with modern 70-year mean annual air temperatures in the WSL ($P < 0.0001$, $r^2 = 0.82$; see Figure 5). The fitted model is conservative at the warm end of the air temperature gradient where departures from the curve are greatest, e.g., the deepest predicted value is ~2 m, whereas observed peat of 2 ka age is as deep as 2.7 m (Table 2).

[25] *Sheng et al.* [2004] mapped in detail the geographic distribution of peatlands and peat attributes for the entirety of the WSL. To each of *Sheng et al.*'s [2004] peatland polygons, using GIS we assigned a mean annual air temperature from K. Matsuura and C. J. Willmott's 0.5° × 0.5° gridded 1930–2000 temperature data set (Arctic Land-Surface Air Temperature: 1930–2000 Gridded Monthly Time Series, version 1.01, 2004, Center for Climate Research, University of Delaware, Newark; available at http://climate.geog.udel.edu/~climate/html_pages/archive.html), and then a 2 ka peat depth using the depth-temperature relationship (Figure 5).

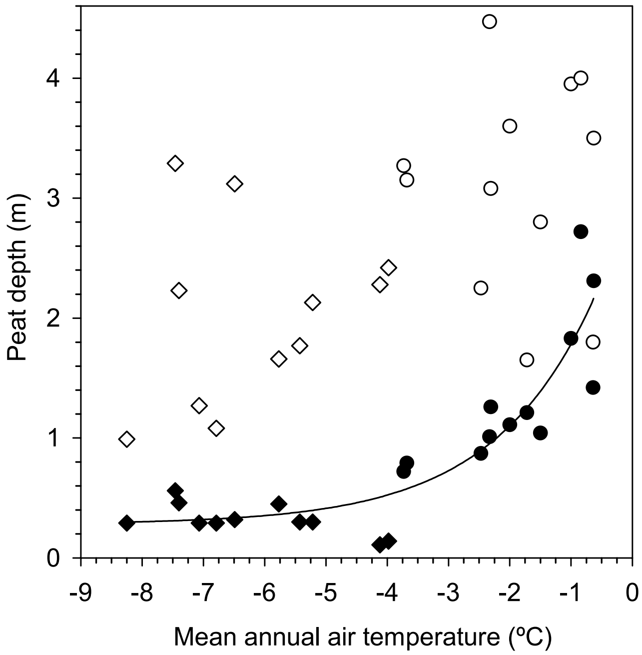

Figure 5. Relationship between modern mean annual air temperature (MAAT, 1931–2000) (Matsuura and Willmott, data set, 2004; available at http://climate.geog.udel.edu/~climate/html_pages/archive.html) and peat depth at 23 sites. Open symbols show total peat depth, solid symbols show 2 ka depths (2 ka_{depth}), and symbol shape follows Figure 4. The fitted curve yields 2 $ka_{depth} = 0.28 + 2.76\ e^{0.61*MAAT}$, $P < 0.0001$, $r^2 = 0.82$.

In total, 7307 polygons (472, 278 km^2) had temperature values within this range of inference (Figure 6). In these polygons, the independent data of *Sheng et al.* [2004] estimate 52.64 Pg organic C stored as peat. Of this total, our analysis suggests an estimated 21.81 Pg C have accumulated since 2 ka on the basis of the peat C characteristics of *Sheng et al.* [2004]; that is, 41% of the C in these peatlands is younger than 2000 years old. This proportion is highest in southern WSL (65% between 55 and 56°N) and lowest in northern WSL (17% between 70 and 71°N; see Figure 7).

4. Discussion

[26] Our analysis indicates that apparent rates of C accumulation have often been lower since 2 ka than is suggested by long-term Holocene averages, with more than half of our sites showing slower rates since 2 ka. This is particularly true of permafrost sites in the northern WSL. Apparent rates of peat C accumulation can overestimate true rates, and younger sections should show faster rates than older sections under constant conditions of C addition and turnover [*Clymo et al.*, 1998]. In this way, a slower apparent rate since 2 ka compared to the total Holocene rate reveals a clear reduction in long-term C sequestration. Although apparent and modeled true C accumulation rates based on whole-Holocene records can serve as a guide for modern peatland C sink strength [*Gorham*, 1991], the present data show that site-by-site long-term Holocene C totals and accumulation rates are poor indicators of patterns of recent accumulation over the last 100–2000 years.

[27] We find evidence for ongoing but slow apparent rates of C accumulation at northern sites, despite the potential for depressed rates of soil C turnover in such areas. Authors working in northern WSL and nearby regions have reported near-surface peat (top 40 cm) that is thousands of years old [*Peteet et al.*, 1998, and references therein; *Jasinski et al.*, 1998]. Such observations follow similar results from Arctic North America of mid-Holocene age surface peat [e.g., *Garneau*, 2000; *Givelet et al.*, 2004]. A number of the northernmost sites in our present data set also show relatively old peat ages with in the top 50 cm (Table 1). This surprisingly old near-surface C suggests the possibility of a near-shutdown in C accumulation in these Arctic regions in the late Holocene. The late Holocene environment of the WSL was typified by cooling temperatures, evident in the southward retreat of the boreal forest [*MacDonald et al.*, 2000, 2008], in addition to evidence for a long-term trend of increasing aridity [*Wolfe et al.*, 2000]. This cooling may have resulted in a reduced North Atlantic influence in the Nordic Seas after 6 ka, and contributed to cooler and drier conditions eastward to the vicinity of the northern WSL [*MacDonald et al.*, 2000].

[28] If Arctic peatland C accumulation since 2 ka has slowed greatly owing to these cooling conditions, such a near-shutdown may have implications for the C balance response of these ecosystems to future Arctic warming [*Peteet et al.*, 1998]. In our present data, however, plant fragments from 8 and 16 cm below the surface from two northern WSL sites yielded near-modern ages (Table 3) and apparent C accumulation rates on the same order of magnitude as estimated post-2-ka rates. This suggests that any shutdown has not been universal across Arctic WSL, and C accumulation has continued, albeit at reduced rates, over recent centuries at some sites.

[29] In contrast to northern sites, comparison of total C accumulated since 2 ka to the independent estimate of the total WSL peat C pool by *Sheng et al.* [2004] suggests that southern WSL peatlands have sequestered most of their current C storage since 2 ka (up to 65%; see Figure 7). High rates of apparent C accumulation in the southern WSL has been observed by other authors working in this region. For example, our southernmost sites have peat accumulations of more than 2 m since 2 ka (basal ages of 2.8 and 3.8 ka) compared to *Borren et al.* [2004] who report a peat accumulation of 2.2 m since 2 ka at a site with much deeper peat and a much older age of peat initiation (11 m; basal age of 9.7 ka).

[30] Estimates of the mean NPP of high-latitude peatland vegetation from sites in North America, Europe, and Siberia are typically about 200 to 800 g biomass m^{-2} a^{-1} [*Vitt et al.*, 2001; *Gunnarsson*, 2005; *Peregon et al.*, 2008]; about 100 to 400 g C m^{-2} a^{-1} if biomass is assumed to be 50% C. When compared to the near-surface rates of apparent C accumulation in our sites these NPP estimates suggest that in the southern WSL a substantial portion of annual net CO_2 uptake and plant litter production is stabilized as peat C over long time frames. The resulting fast accumulation may be owing to the inherent recalcitrance of *Sphagnum*-derived peat C, or by relatively depressed aerobic C losses in near surface layers.

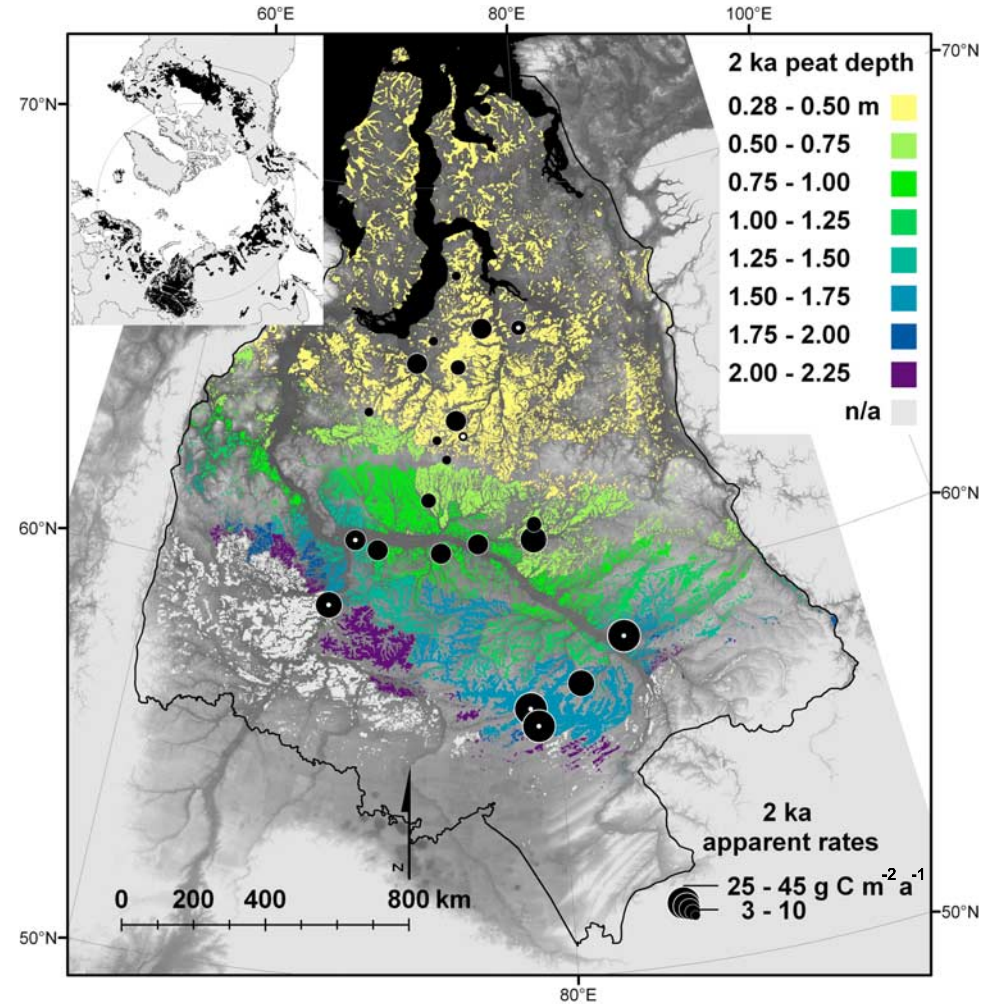

Figure 6. Map of estimated depths of peat accumulated sine 2 ka in the WSL. The 2 ka depth is based on ^{14}C-AMS age determinations of profiles from 23 sites (black circles; see Table 1) and the relationship with modern air temperature (Figure 4). Gray polygons are out of range of inference and are not estimated for 2 ka depth. Scaled black circles show rates of apparent C accumulation since 2 ka in increments of 5 g C m^{-2} a^{-1} unless shown otherwise. Black circles with white dots show sites with C accumulation data for recent centuries (Table 3). The 1-km digital terrain data show linear stretched values (0–250 m above sea level) from the NOAA GLOBE data set. Inset map shows the global distribution of northern peatlands [*MacDonald et al.*, 2006].

Alternatively, these fast rates could be the result of enhanced NPP in southern WSL peatlands. *Peregon et al.* [2008] report peatland (nonswamp) NPP for a study area in the southern taiga of the WSL of about 500 to 900 g biomass m^{-2} a^{-1}, but values as high as about 1300 g biomass m^{-2} a^{-1} in some microsites. The relative importance of NPP versus peat OM stabilization factors, and their sensitivity to climate variation, can be addressed using paleoenvironmental methods. So far this has been an understudied component of long-term C accumulation in the WSL.

[31] The variation in C accumulation and the strength the WSL C sink between regions is likely determined by a number of factors. Wildfire can play a substantial role in C accumulation when direct C losses from combustion are large [e.g., *Pitkanen et al.*, 1999]. At the scale of the entire WSL, the modern mean fire return interval is substantially shorter in the southern Eurasian taiga (estimated to be ~300–500 years) than in the Eurasian Arctic (~900 to 2000 years) [*Balshi et al.*, 2007], thus it would be expected that wildfire influence is greatest in southern WSL peatlands. However, in a peatland complex in the southern WSL, *Turunen et al.* [2001] report little evidence for large C losses from fire (macroscopic charcoal in peat) since 2 ka except very close to the margins of a large peatland complex. In contrast, a few studies of Arctic WSL peatlands report near-surface charcoal layers and charred macrofossils that indicate fire has played some role in long-term C accumulation [*Peteet et al.*, 1998; *Jasinski et al.*, 1998]. Because these Arctic sites also report very old near-surface peat ages which are not as evident in our cores, we suspect that any fire losses have likely played a secondary role in the regional pattern of C accumulation since 2 ka in our sites.

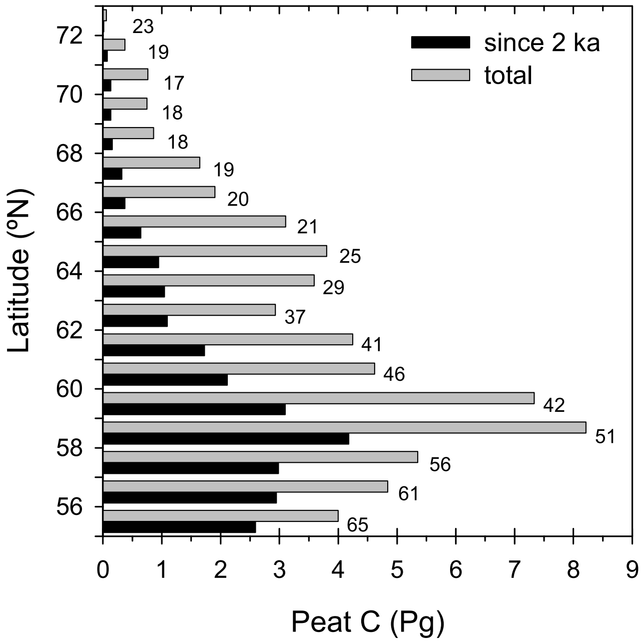

Figure 7. Latitudinal distribution of estimated peat C storage since 2 ka and total peat C storage over the Holocene [*Sheng et al.*, 2004]. Shown are totals only for polygons within range of inference for 2 ka depths. Numbers beside bars show the percentage of total C storage accumulated since 2 ka.

[32] Climate affects the NPP of peatland plant communities [*Vitt et al.*, 2001; *Gunnarsson*, 2005; *Peregon et al.*, 2008], the microbial mineralization of near-surface organic matter, and possibly the decomposition processes in deep peat [*Clymo et al.*, 1998]. In this way, a climatic optimum for peat C accumulation may occur where air-soil temperature differences optimize NPP relative to soil respiration, e.g., near 0°C MAAT [*Swanson et al.*, 2000]. At present, the approximate location of the 0°C MAAT isotherm in the WSL occurs between 56 and 60°N (Figure 1), which is roughly coincident with the greatest peatland extent and peat C stocks [*Sheng et al.*, 2004] and the greatest C accumulation since 2 ka (Figure 7). The potential relationship between temperature and peat C sequestration, and the current spatial distribution of peatland ecosystems, should be an important consideration in future attempts to anticipate the impact of climate warming on the C sink potential of the WSL region.

[33] The lack of agreement between apparent rates of C accumulation since 2 ka and over the entire Holocene supports a hypothesis that, if a climatic optimum exists for peat C sequestration, the location of this optimum may have been time-transgressive over the last 12,000 years. An extensive analysis of variations in Holocene apparent C accumulation rates across a number of sites would be needed to trace such movements. There is also, however, considerable variation in C storage since 2 ka BP not accounted for simply by climate, particularly in the southern WSL peatlands in our data set (Figures 3 and 5). Determining the mechanisms that affect this intraregional variation, including how these mechanisms may differ between peatlands of varying trophic status and vegetation type [e.g., *Trumbore et al.*, 1999], is an important future research direction since these peatlands contribute a substantial proportion of C accumulation in Siberia since 2 ka.

[34] In addition to the N-S pattern of post-2-ka peat C accumulation and its relationship with the modern MAAT gradient, our data also reflect the influence of permafrost status on long-term C accumulation. On average, cores from nonpermafrost sites have accumulated four times more peat by depth and twice as much C than cores from permafrost sites. A closer look at sites with similar MAAT (-3.5 to -4.5°C) shows that, since 2 ka, nonpermafrost sites (V034 and V039) have accumulated between four and six times more peat C than permafrost sites (D127 and N001; see Figure 3 and Table 2). Forecasts of permafrost response to future Arctic warming differ in the extent of thaw expected over the next century, but agree that substantial thaw across the Arctic can be expected [*Lawrence and Slater*, 2005; *Delisle*, 2007]. Within WSL regions where permafrost distributions may change substantially, it can be speculated on the basis of the observed pattern since 2 ka that permafrost thaw may promote a boost in peat C sequestration in affected sites, and that this response may be nonlinear with warming.

[35] Our data suggest that for WSL peatlands the long-term average C accumulation rate since 2 ka has been about 0.011 Pg C a^{-1}, on the basis of the 22 Pg C accumulated since 2 ka estimated in section 4.4. This rate is a conservative estimate for the entire WSL since it excludes those southern peatlands that are out of the range of inference (Figure 6). This rate is ~50% greater than that suggested by the total peat C stock of 53 Pg C for the same area and an average peatland initiation age of 7.3 ka (0.007 Pg C a^{-1}; mean of 186 basal dates) [*Borren et al.*, 2004; *Kremenetski et al.*, 2003; *Smith et al.*, 2004]. Ecosystem modeling estimates the Eurasian terrestrial C sink to have been about 0.3 to 0.6 Pg C a^{-1} over recent years [*Potter et al.*, 2005] but about 0.1 to 0.25 Pg C a^{-1} in some years [*Potter et al.*, 2003]. Thus, the average annual WSL peat C sink, mainly in large southern peatlands, could be as much as 10% of the total annual Eurasian terrestrial sink. Owing to the substantial magnitude of this long-term average rate for WSL peatlands, it is important to further understand the interdecadal to intercentennial variability of the southern WSL C accumulation over the recent past. This has not yet been studied in detail.

5. Conclusions

[36] Using peat C characteristics from a network of radiocarbon-dated peat cores, we quantified a strong geographical trend in the amount of C accumulated in peatland ecosystems of the WSL since 2 ka. OC content was found to be significantly lower in *Sphagnum*-derived peat, and we refined our estimates of apparent C accumulation rates by taking this difference into account. Peatlands in the southern WSL have likely been relatively strong C sinks since 2 ka. Southern sites in our network have accumulated as much as ~250 cm peat and ~88 kg C m^{-2} since 2 ka. Northern WSL peatlands have smaller C accumulations, as little as ~10 cm and 7 kg C m^{-2} over the same period. Ages of near-surface peat at seven sites shows that this general geographical pattern of peat C

accumulation since 2 ka has continued over recent centuries in the WSL peatlands. In contrast, average Holocene values, reflecting the total C accumulated since peatland initiation, showed little relationship with C accumulation since 2 ka. Extrapolation of our site data suggests that, as a first approximation, about 40% of total WSL peat C has accumulated since 2 ka, mainly in southern sites. The magnitude of these long-term accumulations suggests that the very large peatlands of the southern WSL may be an important component of the Eurasian terrestrial C sink.

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