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North Siberian Lakes: A Methane Source Fueled by Pleistocene Carbon

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The sizes of major sources and sinks of atmospheric methane (CH_4), an important greenhouse gas, are poorly known. CH_4 from north Siberian lakes contributes ~ 1.5 teragrams $\text{CH}_4 \text{ year}^{-1}$ to observed winter increases in atmospheric CH_4 concentration at high northern latitudes. CH_4 emitted from these lakes in winter had a radiocarbon age of 27,200 years and was derived largely from Pleistocene-aged carbon.

The highest concentration and greatest seasonal amplitude of atmospheric CH_4 occurs at 65° to 70°N . Concentrations are highest in March to April and lowest in summer (1). Photochemical oxidation of CH_4 contributes to the low summer levels (2) but does not explain why the seasonal amplitude of atmospheric CH_4 is twice as high in the Northern as in the Southern Hemisphere, given large summer effluxes from North American bogs and tundra (3, 4) and modest CH_4 fluxes from Siberian wetlands (5). Between August and April, 5.8 Tg ($1 \text{ Tg} = 10^{12} \text{ g}$) of CH_4 accumulate in the atmosphere north of 60°N (6). High-latitude winter fluxes measured in a muskeg and a peatland were only 10 to 12% of the annual total (4, 7), an insufficient flux to explain a winter maximum in atmospheric CH_4 . Here we provide evidence for a large winter CH_4 source from Siberian lakes.

In the Pleistocene, most of the northern Siberian plains were unglaciated and accumulated $\sim 400,000 \text{ Tg}$ of organic C in sediments (8) (mainly derived from plant roots), similar to the total C in the terrestrial biosphere (9). These sediments contained abundant ice (40 to 70% of soil volume) (10–12), which began melting during the Holocene to form thermokarst (thaw) lakes that now make up $\sim 30\%$ of the landscape. These lakes migrated across the north Siberian plains during the Holocene (10), releasing to the atmosphere an average of 170 to $220 \text{ g C m}^{-2} \text{ year}^{-1}$, including $\sim 16 \text{ g CH}_4 \text{ m}^{-2} \text{ year}^{-1}$; we esti-

mate that half of this CH_4 was derived from Pleistocene C (13). Siberian lake sediments produce CH_4 bubbles in lakes throughout the year (14), particularly near shores with active erosion. During winter, the bubbles form koshkas, which are flat bubbles of CH_4 in lake ice separated by ice films that periodically sublimate and release CH_4 to the atmosphere. In areas where CH_4 ebullition (bubbling) is most active, channels through the ice remain open all winter.

To evaluate the significance of this source, we incubated Pleistocene sediments from an eroding lakeshore with lake water. The yield was $65 \pm 3 \text{ mg CH}_4 \text{ g}^{-1} \text{ sediment}$ at 15°C (mean $\pm \text{SE}$, $n = 3$) over 12 months, equivalent to 5% of the C originally present in the soil; $26 \pm 2 \text{ mg CH}_4 \text{ g}^{-1}$ were emitted at 3.5°C , and $19 \pm 2 \text{ mg CH}_4 \text{ g}^{-1}$ were emitted at 0°C . These data indicate that the C in Pleistocene sediments is sufficiently labile to support methanogenesis and that, although methanogenesis is temperature-sensitive, it occurs at substantial rates at 0° to 3.5°C .

To determine whether methanogenesis in lake sediments is currently fueled by Pleistocene-aged organic matter, we measured stable

and radiocarbon isotopes of CH_4 emitted by ebullition from two representative thaw lakes near Cherskii, Republic of Sakha (Yakutia), Russia (69°N , 161°E). CH_4 collected from these lakes in winter (April) had an average ^{14}C age of 27,200 years (Table 1). This age indicates that Pleistocene sediments deposited 20,000 to 40,000 ^{14}C years ago (11) contributed 68 to 100% of CH_4 flux from these lakes. In contrast, CH_4 emitted in the summer (July) had an average ^{14}C age of 9,200 years, indicating that Pleistocene C fueled 23 to 46% of summer methanogenesis and thus that more CH_4 was produced in the younger surface sediments, which are warmer in summer than winter (10). Thus, about half of current annual methanogenesis is fueled by Pleistocene C. In contrast, CH_4 from Alaskan lakes was only 200 years old (15) because Alaska lacks extensive Pleistocene sediments.

The $\delta^{13}\text{C}$ value of CH_4 collected from Siberian lakes was -71 to -73 (Table 1). This value is less than that produced in summer by Alaskan tundra lakes ($\delta^{13}\text{C} = -61 \pm 2$) (15) or North American wet tundra ($\delta^{13}\text{C} = -66$ to -63) (15, 16). These values imply that the Siberian winter-collected CH_4 was not as oxidized as in these other environments, or that there was an isotopic difference in substrate or a different pathway of methanogenesis (17). The hydrogen isotopic composition of the CH_4 was variable, but most samples from the Siberian lakes were low ($\delta\text{D} = -370$), indicative of a biotic source for CH_4 , low oxidation rates in the water column, and CH_4 production by fermentation (17, 18).

We measured CH_4 ebullition fluxes from two thaw lakes using large funnels suspended beneath the ice (19). CH_4 fluxes were generally highest from October to January (Fig. 1), when deep sediments had their annual thermal maximum (10). Fluxes were highly variable within a season; fluxes were highest at

Table 1. Isotopic data for CH_4 collected from sediments in two thaw lakes in Pleistocene sediments in the Kolyma lowlands. Results are given as percent modern C, or 100 times the ratio of $^{14}\text{C}/^{12}\text{C}$ in the sample divided by the $^{14}\text{C}/^{12}\text{C}$ ratio in 1895 wood (corrected for $^{13}\text{C}/^{12}\text{C}$ differences) (27, 28).

Lake no.	Lake depth (m)	$\delta^{13}\text{C}$	$^{14}\text{CH}_4$ age	Modern C (% of $\text{CH}_4\text{-C}$)	δD
<i>Summer</i>					
13	5	-69.5	*	*	
	7	-69.6	$11,731 \pm 360$	23.2 ± 1.0	
	10	-71.8	$8,330 \pm 240$	35.3 ± 1.0	
	10†	-72.1	$8,370 \pm 180$	35.1 ± 0.8	
	10†	-	$8,330 \pm 100$	35.3 ± 0.4	
Average:		-70.8 ± 0.7	$9,200 \pm 800$	32 ± 3	
<i>Winter</i>					
13	2	-75.1	*	*	-169
	10	-80.0	38,000	0.2	-430
14	4†	-72.7	$28,670 \pm 850$	2.8	-479
	4†	-	$27,230 \pm 730$	3.4	-
	10	-65.3	$15,000 \pm 150$	15.3	-391
Average:		-73.3 ± 3.1	$27,200 \pm 4700$	5 ± 3	-370 ± 70

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*Not determined. †Subsamples analyzed separately.

Table 2. Summary of CH₄ fluxes from Siberian lakes during times of increasing (August to April) and decreasing (May to July) atmospheric CH₄, based on values presented in the text.

Flux component	Area (% of lake)	Area-averaged CH ₄ flux (g CH ₄ m ⁻² year ⁻¹)		
		May–July	August–April	Annual
Ebullition flux				
Lake center*	80	0.3	1.0	1.3
Lake shore*	20	1.0	3.1	4.1
Overflow	100	0	0.5	0.5
Diffusion	100	0.5	0.4	0.9
Lake total (g CH ₄ m ⁻² year ⁻¹)		1.8	5.0	6.8
Siberian total (Tg CH ₄ year ⁻¹)		0.5	1.5	2.0

*Averaged over the entire lake area. Ebullition flux near the lake shore is a conservative estimate that does not include CH₄ released through koshas.

times of low atmospheric pressure, as in north temperate lakes (20). The average CH₄ ebullition flux in centers of lakes (80% of the lake area) was 4.7 ± 2.3 mg CH₄ m⁻² day⁻¹ (90 measurements). Near eroding lake shores, fluxes were so high that they frequently overturned the collection funnels; the flux (2 cm³ min⁻¹, $n = 25$; 90% CH₄) emitted from open holes in the ice (300 ha⁻¹) was 56 mg CH₄ m⁻² day⁻¹. In addition to open holes, there were koshas (400 to 1000 ha⁻¹) containing 1 to 100 liters of 50% (25 to 75%) CH₄ ($n = 8$) that vent CH₄ several times each winter and provide an additional unquantified CH₄ source. Sediments in these lakes released 22 g CH₄ m⁻² (38 ± 12 liter m⁻² of 80% CH₄, $n = 3$) in September, when we disturbed the sediments (15), but <5 g CH₄ m⁻² a month later. This large CH₄ release over 1 month provides independent evidence for a large CH₄ ebullition flux.

CH₄ can also move to the atmosphere in winter in overflow, when the weight of winter snow pushes the ice below the equilibrium water level of the lake. The CH₄ concentration in overflow water decreases from 1.7 mg CH₄ liter⁻¹ (21) to <0.01 mg CH₄ liter⁻¹. The 30 cm of overflow that typically accumulate on lakes of the forest zone would thus release 0.52 g of dissolved CH₄ m⁻² year⁻¹.

Fig. 1. Time course of CH₄ flux from the center of a thaw lake near the forest-tundra border 100 km south of the Arctic Ocean near the Kolyma River Valley (69°N, 161°W) in northern Siberia (19).

The average summer diffusive flux measured in 19 lakes along a climate transect inland from the Arctic Ocean was 7.6 ± 1.4 mg CH₄ m⁻² day⁻¹ (60 measurements) (21), a value similar to that in Alaskan lakes (6.8 ± 1.3 mg m⁻² day⁻¹) (22). The 19 lakes had 3.1 ± 0.7 mg m⁻² of dissolved CH₄ in March, indicating that winter accumulation of CH₄ is typical of north Siberian lakes.

We estimate the total annual flux of CH₄ for the lakes in our study region to be at least 7 g CH₄ m⁻² year⁻¹ (Table 2), ~50% of the potential flux we estimated (16 g CH₄ m⁻² year⁻¹) from regional C inputs to lakes. Approximately 75% of this flux occurs in winter. If these fluxes are typical of Siberian lakes, these lakes would emit ~1.5 Tg CH₄ in winter (2 Tg CH₄ annually). This is small relative to global sources (18) but is 25% of the high-latitude winter accumulation of CH₄ in the atmosphere. If high-latitude warming trends (23) continue, thawing of permafrost would increase, and methane flux from Siberian thaw lakes would act as a positive feedback to climate warming.

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Sizes and Ages of Seamounts Using Remote Sensing: Implications for Intraplate Volcanism

Paul Wessel

Satellite altimetry was used to identify and characterize Pacific intraplate seamounts. The gravimetric amplitudes of seamounts appear to be related to the age difference between the sea floor and seamounts; by inverting this relation, pseudo ages can be obtained for undated seamounts. These pseudo ages imply that excursions in seamount volcanism generally correlate with times of formation of large oceanic plateaus.

The Pacific plate may support more than 50,000 seamounts taller than 1 km, yet ~50% of these undersea volcanoes are uncharted because of sparse bathymetric coverage (1, 2). Even fewer (<1%) have been sampled for radiometric dating (3), making assessment of temporal fluctuations in intraplate volcanism uncertain. Because electromagnetic sensing devices cannot penetrate the oceans, we are unable to image the sea floor remotely and instead must rely on surface ships equipped with sonar. At the present rate of data acquisition, complete bathymetric coverage may take centuries. However, the density contrast between seawater and the sea floor basalt gives rise to gravity anomalies. These minute variations in Earth's gravitational pull cause seawater

to be attracted to seamounts, leading to a sea surface (which approximates the geoid) whose shape reflects these underlying fea-

tures (4). Thus, since the early 1980s, satellite altimetry has provided broad coverage of the sea surface or geoid undulations (5). Early attempts to map the seamount distribution were largely limited by the coarseness of the satellite coverage [the typical track spacing was >100 km (6)], and many seamounts went undetected. Because seamounts are typically much smaller than 100 km, it was difficult to estimate what part of the seamount had been traversed by the satellite, leading to large uncertainties in estimates of seamount height and diameter (7).

Recently, the U.S. Navy declassified its Geosat satellite altimetry, which has been combined with the European Space Agency

Fig. 1. Theoretical VGG over an isostatically compensated seamount of radius R and height H (14). The amplitude v_0 and zero-crossing distance d are the two clearest characteristics of the anomaly.

Fig. 2. Equal-area Hammer projection showing all 8882 seamounts found on the Pacific plate; the sizes of crosses reflect the VGG amplitudes. Blue crosses are small seamounts (30 to 60 Eötvös units, generally <2.5 km tall), red crosses are large seamounts (>120 Eötvös units, generally >3.5 km tall), and green crosses are of intermediate size. The Eltanin fracture zone is indicated.