

# 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<sub>4</sub>), an important greenhouse gas, are poorly known. CH<sub>4</sub> from north Siberian lakes contributes ~1.5 teragrams CH<sub>4</sub> year<sup>-1</sup> to observed winter increases in atmospheric CH<sub>4</sub> concentration at high northern latitudes. CH<sub>4</sub> 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<sub>4</sub> occurs at 65° to 70°N. Concentrations are highest in March to April and lowest in summer (1). Photochemical oxidation of CH<sub>4</sub> contributes to the low summer levels (2) but does not explain why the seasonal amplitude of atmospheric CH<sub>4</sub> 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<sub>4</sub> fluxes from Siberian wetlands (5). Between August and April, 5.8 Tg (1 Tg = 10<sup>12</sup> g) of CH<sub>4</sub> 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<sub>4</sub>. Here we provide evidence for a large winter CH<sub>4</sub> source from Siberian lakes.

In the Pleistocene, most of the northern Siberian plains were unglaciated and accumulated ~400,000 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 ~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 g C m<sup>-2</sup> year<sup>-1</sup>, including ~16 g CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup>; we esti-

mate that half of this CH<sub>4</sub> was derived from Pleistocene C (13). Siberian lake sediments produce CH<sub>4</sub> 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<sub>4</sub> in lake ice separated by ice films that periodically sublimate and release CH<sub>4</sub> to the atmosphere. In areas where CH<sub>4</sub> 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 ± 3 mg CH<sub>4</sub> g<sup>-1</sup> sediment at 15°C (mean ± SE, n = 3) over 12 months, equivalent to 5% of the C originally present in the soil; 26 ± 2 mg CH<sub>4</sub> g<sup>-1</sup> were emitted at 3.5°C, and 19 ± 2 mg CH<sub>4</sub> g<sup>-1</sup> 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<sub>4</sub> emitted by ebullition from two representative thaw lakes near Cherskii, Republic of Sakha (Yakutia), Russia (69°N, 161°E). CH<sub>4</sub> collected from these lakes in winter (April) had an average <sup>14</sup>C age of 27,200 years (Table 1). This age indicates that Pleistocene sediments deposited 20,000 to 40,000 <sup>14</sup>C years ago (11) contributed 68 to 100% of CH<sub>4</sub> flux from these lakes. In contrast, CH<sub>4</sub> emitted in the summer (July) had an average <sup>14</sup>C age of 9,200 years, indicating that Pleistocene C fueled 23 to 46% of summer methanogenesis and thus that more CH<sub>4</sub> 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<sub>4</sub> from Alaskan lakes was only 200 years old (15) because Alaska lacks extensive Pleistocene sediments.

The δ<sup>13</sup>C value of CH<sub>4</sub> collected from Siberian lakes was -71 to -73 (Table 1). This value is less than that produced in summer by Alaskan tundra lakes (δ<sup>13</sup>C = -61 ± 2) (15) or North American wet tundra (δ<sup>13</sup>C = -66 to -63) (15, 16). These values imply that the Siberian winter-collected CH<sub>4</sub> 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<sub>4</sub> was variable, but most samples from the Siberian lakes were low (δD = -370), indicative of a biotic source for CH<sub>4</sub>, low oxidation rates in the water column, and CH<sub>4</sub> production by fermentation (17, 18).

We measured CH<sub>4</sub> ebullition fluxes from two thaw lakes using large funnels suspended beneath the ice (19). CH<sub>4</sub> 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<sub>4</sub> 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 <sup>14</sup>C/<sup>12</sup>C in the sample divided by the <sup>14</sup>C/<sup>12</sup>C ratio in 1895 wood (corrected for <sup>13</sup>C/<sup>12</sup>C differences) (27, 28).

Lake no.	Lake depth (m)	δ <sup>13</sup> C	<sup>14</sup> CH <sub>4</sub> age	Modern C (% of CH <sub>4</sub> -C)	δD
<i>Summer</i>					
13	5	-69.5	*	*	
	7	-69.6	11,731 ± 360	23.2 ± 1.0	
	10	-71.8	8,330 ± 240	35.3 ± 1.0	
	10†	-72.1	8,370 ± 180	35.1 ± 0.8	
	10†	-	8,330 ± 100	35.3 ± 0.4	
Average:		-70.8 ± 0.7	9,200 ± 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 ± 850	2.8	-479
	4†	-	27,230 ± 730	3.4	-
	10	-65.3	15,000 ± 150	15.3	-391
Average:		-73.3 ± 3.1	27,200 ± 4700	5 ± 3	-370 ± 70

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\*Not determined. †Subsamples analyzed separately.

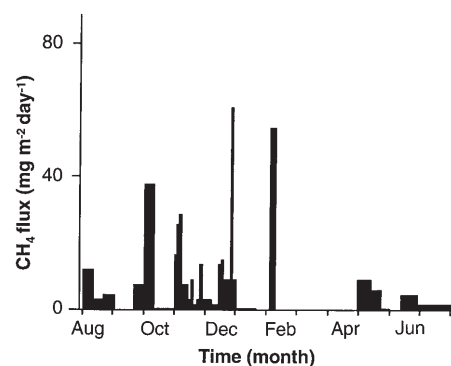
**Table 2.** Summary of CH<sub>4</sub> fluxes from Siberian lakes during times of increasing (August to April) and decreasing (May to July) atmospheric CH<sub>4</sub>, based on values presented in the text.

Flux component	Area (% of lake)	Area-averaged CH <sub>4</sub> flux (g CH <sub>4</sub> m <sup>-2</sup> year <sup>-1</sup> )		
		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 <sub>4</sub> m <sup>-2</sup> year <sup>-1</sup> )		1.8	5.0	6.8
Siberian total (Tg CH <sub>4</sub> year <sup>-1</sup> )		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 CH<sub>4</sub> released through koshas.

times of low atmospheric pressure, as in north temperate lakes (20). The average CH<sub>4</sub> ebullition flux in centers of lakes (80% of the lake area) was  $4.7 \pm 2.3$  mg CH<sub>4</sub> m<sup>-2</sup> day<sup>-1</sup> (90 measurements). Near eroding lake shores, fluxes were so high that they frequently overturned the collection funnels; the flux (2 cm<sup>3</sup> min<sup>-1</sup>,  $n = 25$ ; 90% CH<sub>4</sub>) emitted from open holes in the ice (300 ha<sup>-1</sup>) was 56 mg CH<sub>4</sub> m<sup>-2</sup> day<sup>-1</sup>. In addition to open holes, there were koshas (400 to 1000 ha<sup>-1</sup>) containing 1 to 100 liters of 50% (25 to 75%) CH<sub>4</sub> ( $n = 8$ ) that vent CH<sub>4</sub> several times each winter and provide an additional unquantified CH<sub>4</sub> source. Sediments in these lakes released 22 g CH<sub>4</sub> m<sup>-2</sup> ( $38 \pm 12$  liter m<sup>-2</sup> of 80% CH<sub>4</sub>,  $n = 3$ ) in September, when we disturbed the sediments (15), but <5 g CH<sub>4</sub> m<sup>-2</sup> a month later. This large CH<sub>4</sub> release over 1 month provides independent evidence for a large CH<sub>4</sub> ebullition flux.

CH<sub>4</sub> 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 CH<sub>4</sub> concentration in overflow water decreases from 1.7 mg CH<sub>4</sub> liter<sup>-1</sup> (21) to <0.01 mg CH<sub>4</sub> liter<sup>-1</sup>. The 30 cm of overflow that typically accumulate on lakes of the forest zone would thus release 0.52 g of dissolved CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup>.



**Fig. 1.** Time course of CH<sub>4</sub> 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  $7.6 \pm 1.4$  mg CH<sub>4</sub> m<sup>-2</sup> day<sup>-1</sup> (60 measurements) (21), a value similar to that in Alaskan lakes ( $6.8 \pm 1.3$  mg m<sup>-2</sup> day<sup>-1</sup>) (22). The 19 lakes had  $3.1 \pm 0.7$  mg m<sup>-2</sup> of dissolved CH<sub>4</sub> in March, indicating that winter accumulation of CH<sub>4</sub> is typical of north Siberian lakes.

We estimate the total annual flux of CH<sub>4</sub> for the lakes in our study region to be at least 7 g CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup> (Table 2), ~50% of the potential flux we estimated (16 g CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup>) 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 CH<sub>4</sub> in winter (2 Tg CH<sub>4</sub> annually). This is small relative to global sources (18) but is 25% of the high-latitude winter accumulation of CH<sub>4</sub> 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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13. We estimate 120 g C m<sup>-2</sup> year<sup>-1</sup> input of Pleistocene C to lakes, assuming that 400 kg C m<sup>-2</sup> were initially in Pleistocene sediments (8) and that  $0.3 \times 10^6$  km<sup>2</sup> of lakes (24) migrated across 90% of the plains during the last 10,000 years (10). We estimate Holocene C input to lakes assuming that the erosion rate for lakes 1 km<sup>2</sup> in area is 0.5 to 1.0 m year<sup>-1</sup> (10) that is, 0.05 to 0.1% of lake area) and that the lakes receive 100 kg m<sup>-2</sup> organic C from vegetation (3 to 10 kg m<sup>-2</sup>), peat (0 to 200 kg m<sup>-2</sup>), soil (7 to 30 kg m<sup>-2</sup>), and upper 3 m of permafrost (20 to 60 kg m<sup>-2</sup>) (10, 11, 25) = 50 to 100 g C m<sup>-2</sup> of lake area. We do not know the contribution of in-lake production plus dissolved organic C inputs to CH<sub>4</sub> fluxes, so we used values measured in Alaskan oligotrophic tundra lakes lacking major erosional inputs: 0.7 g CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup> (22). If 5 to 10% of this C input were converted to CH<sub>4</sub> in anaerobic sediments (26), this would yield an annual flux of ~16 (9 to 23) g CH<sub>4</sub> m<sup>-2</sup> year<sup>-1</sup>.
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19. We measured CH<sub>4</sub> flux in conical 8-m<sup>2</sup> plastic traps anchored beneath the ice surface, which funneled CH<sub>4</sub> bubbles into collection flasks. Samples were collected every 0.5 to 10 days, depending on flux. Fluxes from seven other chambers in two lakes were similar to those shown but gave incomplete seasonal data due to winter ice damage. Water depth was 10 m. CH<sub>4</sub> content of recently produced bubbles was >80% CH<sub>4</sub>, as determined ( $\pm 1\%$  accuracy) by gas chromatography (GC). Each sample was analyzed on both a TSVET-530 and a Shimadzu 14A GC with a thermal conductivity detector and a flame ionization detector, respectively. We used data from the thermal conductivity detector if CH<sub>4</sub> concentration was >1%; otherwise, we used data from the flame ionization detector. By running all samples through both detectors, we cross-calibrated the two analytical methods. CH<sub>4</sub> concentration of samples that remained in collection flasks >24 hours declined due to diffusion into the water column. In these cases, we estimated CH<sub>4</sub> content as  $0.8 \times$  gas volume.
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28. Samples were collected directly into glass serum vials from ebullition flux (15). Before analysis, water was added to vials to maintain high internal pressures, the sample was shaken vigorously to get dissolved CH<sub>4</sub> out of

solution, each sample was transferred to a 6-liter stainless steel container with a syringe, and analytical grade-zero air was added. CH<sub>4</sub> was separated from each sample and combusted. A subsample of the resulting CO<sub>2</sub> was analyzed for <sup>13</sup>C/<sup>12</sup>C by mass spectrometry, and the remaining CO<sub>2</sub> was catalytically reduced to graphite (29), and its <sup>14</sup>C/<sup>13</sup>C values measured by atomic mass spectrometry at the Lawrence Livermore National Laboratory.

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

Paul Wessel

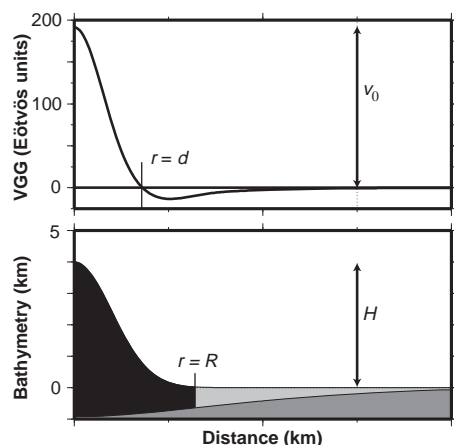
Satellite altimetry was used to identify and characterize Pacific intraplate seamounts. The gravimetric amplitudes of seamounts appear to be related to the age difference between the sea floor and seamounts; by inverting this relation, pseudo ages can be obtained for undated seamounts. These pseudo ages 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

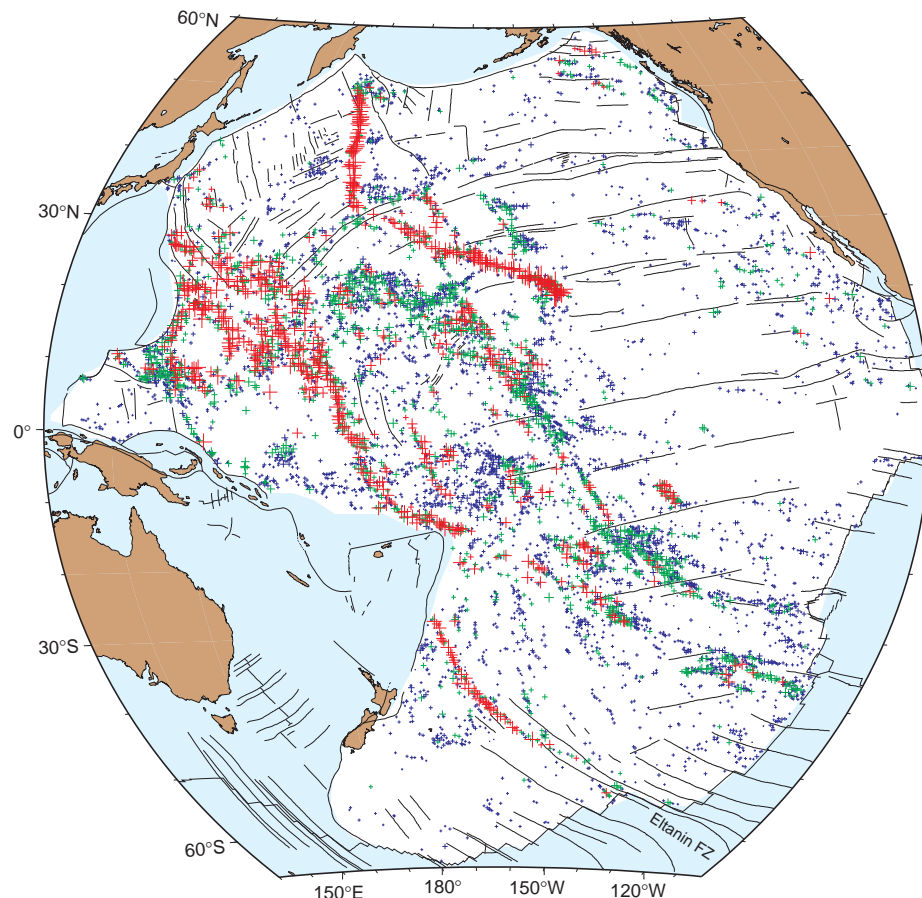
to be attracted to seamounts, leading to a sea surface (which approximates the geoid) whose shape reflects these underlying fea-

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.