North Siberian Lakes: A Methane Source Fueled by Pleistocene Carbon


The sizes of major sources and sinks of atmospheric methane (CH₄), an important greenhouse gas, are poorly known. CH₄ from north Siberian lakes contributes ~1.5 teragrams CH₄ year⁻¹ to observed winter increases in atmospheric CH₄ concentration at high northern latitudes. CH₄ 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 CH₄ occurs at 65° to 70°N. Concentrations are highest in March to April and lowest in summer (1). Photochemical oxidation of CH₄ contributes to the low summer levels (2) but does not explain why the seasonal amplitude of atmospheric CH₄ is twice as high in the Northern as in the Southern Hemisphere, given large summer effluces from North American bogs and tundra (3, 4) and modest CH₄ fluxes from Siberian wetlands (5). Between August and April, 5.8 Tg (1 Tg = 10¹² g) of CH₄ 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 CH₄. Here we provide evidence for a large winter CH₄ 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 thermokast (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⁻² year⁻¹, including ~16 g CH₄ m⁻² year⁻¹; we estimate that half of this CH₄ was derived from Pleistocene C (13). Siberian lake sediments produce CH₄ bubbles in lakes throughout the year (14), particularly near shores with active erosion. During winter, the bubbles form koshkas, which are flat bubbles of CH₄ in lake ice separated by ice films that periodically sublimate and release CH₄ to the atmosphere. In areas where CH₄ 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 g CH₄ g⁻¹ sediment at 15°C (mean ± SE, n = 3) over 12 months, equivalent to 5% of the C originally present in the soil; 26 ± 2 g CH₄ g⁻¹ were emitted at 3.5°C, and 19 ± 2 g CH₄ 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 CH₄ emitted from two representative thaw lakes near Cherskii, Republic of Sakha (Yakutia), Russia (69°N, 161°E). CH₄ collected from these lakes in winter (April) had an average ¹³C age of 27,200 years (Table 1). This age indicates that Pleistocene sediments deposited 20,000 to 40,000 ¹³C years ago (11) contributed 60% to 100% of CH₄ fluxes from these lakes. In contrast, CH₄ emitted in the summer (July) had an average ¹³C age of 9,200 years, indicating that Pleistocene C fueled 23 to 46% of summer methanogenesis and thus that more CH₄ 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, CH₄ from Alaskan lakes was only 200 years old (15) because Alaska lacks extensive Pleistocene sediments.

The δ¹³C value of CH₄ collected from Siberian lakes was ~–71 to –73 (Table 1). This value is less than that produced in summer by Alaskan tundra lakes (δ¹³C = –61 ± 2) (15) or North American wet tundra (δ¹³C = –66 to –63) (15, 16). These values imply that the Siberian winter-collected CH₄ 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 CH₄ was variable, but most samples from the Siberian lakes were low (δD = –370), indicative of a biotic source for CH₄, low oxidation rates in the water column, and CH₄ production by fermentation (17, 18).

We measured CH₄ ebullition fluxes from two thaw lakes using large funnels suspended beneath the ice (19). CH₄ 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>
<thead>
<tr>
<th>Lake no.</th>
<th>Lake depth (m)</th>
<th>δ¹³C</th>
<th>CH₄ age (14C</th>
<th>Modern C (%)</th>
<th>δD</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Summer</td>
<td></td>
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<tr>
<td>13</td>
<td>5</td>
<td>–69.5</td>
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<td>8,330 ± 240</td>
<td>35.3 ± 1.0</td>
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<td>–72.1</td>
<td>8,370 ± 180</td>
<td>35.1 ± 0.8</td>
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<tr>
<td>10</td>
<td>101</td>
<td>–</td>
<td>8,330 ± 100</td>
<td>35.3 ± 0.4</td>
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<tr>
<td>Average</td>
<td>–70.8 ± 0.7</td>
<td>–</td>
<td>9,200 ± 800</td>
<td>32 ± 3</td>
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<tr>
<td></td>
<td>Winter</td>
<td></td>
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<td></td>
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<tr>
<td>13</td>
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<td>27,200 ± 4700</td>
<td>5 ± 3</td>
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</tbody>
</table>

†To whom correspondence should be addressed. E-mail fschapin@socrates.berkeley.edu.

*Not determined. †Subsamples analyzed separately.

S. A. Zimov, S. P. Davidov, S. F. Prosiannikov, North-East Scientific Station, Pacific Institute for Geography, Far-East Branch, Russian Academy of Sciences, Republic of Sakha, Yakutia, 678880 Cherskii, Russia.†S. F. Prosiannikov and I. P. Semiletov, Pacific Oceanographic Institute, Far-East Branch, Russian Academy of Sciences, Vladivostok, Russia.

S. F. Chapin III and M. C. Chapin, Department of Integrative Biology, University of California, Berkeley, CA 94720–3140, USA.

S. Trumbore and S. Tyler, Department of Earth System Sciences, University of California, Irvine, CA 92697–3100, USA.

†To whom correspondence should be addressed. E-mail fschapin@socrates.berkeley.edu.

800

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times of low atmospheric pressure, as in northern 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 over-turned the collection funnels; the flux (2 cm³ fluxes were so high that they frequently overflowed the collection flasks, and the CH₄ release 0.52 g of dissolved CH₄ m⁻² year⁻¹). This large CH₄ release over 1 month would act as a positive feedback to climate warming.

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 the net CH₄ release from Siberian peatlands (20 to 60 kg m⁻² year⁻¹) (10, 11, 25) to 50 to 100 g C m⁻² year⁻¹ of lake area. We do not know the contribution of in-lake production plus dissolved organic C inputs to CH₄ fluxes, so we used values measured in Alaskan oligotrophic tundra lakes lacking major erosional inputs: 0.7 g CH₄ m⁻² year⁻¹ (26). If these fluxes are typical of north Siberian lakes, we measured CH₄ flux in conical 8-m² plastic traps anchored beneath the ice surface, which funneled CH₄ 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, CH₄ content of recently produced bubbles was ~80% CH₄, as determined (±1% accuracy) by gas chromatography (GC). Each sample was analyzed on both a TSVET-530 and a Shimadzu 1A GC with a thermal conductivity detector and a flame ionization detector, respectively. We used data from the thermal conductivity detector if CH₄ concentration was >1%; otherwise, we used data from the flame ionization detector. By running all samples through both detectors, we calibrated the two analytical methods. CH₄ concentration of samples that remained in collection flasks >24 hours declined due to diffusion into the water column. In these cases, we estimated CH₄ content as 0.85 g CH₄ volume.

The atmosphere north of 60°N has about 380 Tg CH₄ [8% of the total global atmospheric CH₄ pool of 4800 Tg (18)] and increases during winter by 1.5% (25 ppb) (17), amounting to 5.8 Tg.

**REFERENCES AND NOTES**

6. The atmosphere north of 60°N has about 380 Tg CH₄ [8% of the total global atmospheric CH₄ pool of 4800 Tg (18)] and increases during winter by 1.5% (25 ppb) (17), amounting to 5.8 Tg.
8. We estimate that 400,000 Tg C accumulated in Pleistocene loess from soil volume (1 10 km²), 40 m depth (10, 11) and C density (1 g cm⁻³ of soil × 0.08 to 0.04% organic C × 48 sites throughout Siberia, 0 to 50 m depth) (12).
11. Y. K. Vasil’chuk, Izotopnoe kislorodnyi sostav podzem-

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**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 Kolymsk River Valley (69°N, 161°W) in northern Siberia (19).
Sizes and Ages of Seamounts Using Remote Sensing: Implications for Intraplate Volcanism

Paul Wessel

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 features (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

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**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.