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Permafrost carbon: Stock and decomposability of a globally significant carbon pool

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[1] The magnitude of future CO_2-induced climate warming is difficult to predict because of uncertainties in the role of ecosystems and oceans as CO_2 sources and sinks. Siberia has extensive areas (1×10^6 km^2) of deep (up to 90 m) deposits of organic-rich frozen loess (wind-blown silt) that accumulated during the Pleistocene but have not been considered in most global carbon (C) inventories. Similar deposits occur less extensively in Alaska. Recent warming at high latitudes causes this permafrost (permanently frozen ground) to thaw, raising questions about the fate of C in thawing permafrost. Here we show that Siberian loess permafrost contains a large organic C pool (~450 GT—more than half the quantity in the current atmosphere) that decomposes quickly when thawed, and could act as a positive feedback to climate warming.
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1. Introduction

[2] It is difficult to predict the magnitude of future CO_2-induced climate warming because of uncertainties in the role of ecosystems and oceans as CO_2 sources and sinks. Siberia has extensive areas (1×10^6 km^2) of deep (up to 90 m) deposits of organic-rich frozen loess (wind-blown silt) that accumulated during the Pleistocene (Figure 1) [*Romanovsky*, 1993; *Schirrmeister et al.*, 2002; *Sher et al.*, 2005; *Tomirdiaro*, 1980; *Vasil'chuk and Vasil'chuk*, 1998, 1997]. The existence, spatial distribution, and carbon (C) concentrations of these sediments are well documented in the Russian literature but there has been only one preliminary estimate of their C stock [*Zimov et al.*, 1997], so they have never been considered in global C inventories. Similar deposits occur less extensively in Alaska [*Pewe and Journaux*, 1983]. Estimates of high-latitude soil C stocks are based largely on studies of the top 1–2 m of soils, predominantly in North America [*Batjes*, 1996; *Gorham*, 1991; *Melillo et al.*, 1995; *Prentice et al.*, 2001; *Smith et al.*, 2004]. Values for these C stocks have then been extrapolated to a pan-arctic basis [*Batjes*, 1996; *Smith et al.*, 2004]. Here we provide the first quantitative estimate of C stocks in Siberian loess permafrost and show that the pool is large enough (~450 GT C—more than half the quantity in the current atmosphere) to warrant consideration in global C inventories.

[3] High-latitude soil C stocks are of particular interest because recent warming thaws permafrost (permanently frozen ground) [*Osterkamp and Romanovsky*, 1999; *Romanovsky et al.*, 2001], raising questions about the fate of permafrost C [*Zimov et al.*, 2006]. Here we show that the C contained in Siberian loess permafrost decomposes quickly when thawed. We describe observations that provide insights into the processes by which C accumulates in or disappears from Siberian loess permafrost and discuss the implications of changes in permafrost C for the global C cycle.

2. Methods

[4] Near Cherskii on the Kolyma River in Northeast Siberia, we studied the process of C incorporation into permafrost by comparing six grasslands that spanned a wide range of depositional environments, soil moisture, and thaw depths (70–150 cm) to understand the past formation of loess permafrost when grassland ecosystems were widespread. We sampled three soil profiles per community in September 1998 (the time of maximum thaw) from the soil surface to about 20 cm beneath the maximum thaw depth. Communities along a sediment deposition gradient included an annually flooded *Calamagrostis* riverbank (meadow 1); two additional *Calamagrostis* meadows further from the river that receive less sediment input (meadows 2 and 3); a moist herb meadow (meadow 4); a dry herb meadow (dry meadow); and a steppe at the base of a steep south-facing slope (dry steppe) [*Zimov et al.*, 1999].

[5] We collected frozen loess permafrost (termed yedoma in Siberia) that formed during the Pleistocene from deep soil profiles recently exposed by river or coastal erosion. Sites were selected along a current climate gradient from a high-arctic coastal site to boreal forest. Sites were: Chukochii Mys (high arctic 1); Zelenyi Mys (forest tundra 1) Duvannyi Yar (forest tundra 2); and Stanchikovskii Yar (boreal forest). At the last site, the middle of the profile had a buried organic horizon from an interstadial. At Chukochii Mys and Duvannyi Yar, we also sampled sediments beneath former thermokarst lakes where the Pleistocene ice wedges had melted during the Holocene (causing the total soil thickness to decrease due to loss of ice wedge volume from the loess, forming the lake) (former lakes 1 and 2). The unfrozen loess sediments beneath the lakes refroze later in the Holocene after the lakes drained.

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Figure 1. The distribution of yedoma loess sediments in North Siberia (shaded). The study area along the Kolyma River is highlighted by the black rectangle. Modified from *Romanovsky* [1993].

[6] Subsamples from both modern grasslands and yedoma permafrost were dried and analyzed for C using the Walkley-Black method [*Allison*, 1965]. We transported soils to the U.S. and reanalyzed a set of 153 samples (from this and other studies) on a Carlo Erba CHN analyzer to develop a regression relationship between the Walkley-Black method, which measures oxidizable compounds to total organic carbon, as measured on the CHN analyzer. To be conservative, we used the Walkley-Black C values, which averaged 13% less than values from the CHN analyzer, because these values were available for all our samples and were methodologically comparable to values obtained from the Russian literature. Samples included all organic C in the profile, including roots and buried organic matter; no carbonates were detected by acidification and subsequent remeasurement with the CHN analyzer. We present data on a gravimetric dry-mass rather than volumetric basis to avoid biases associated with regional or temporal variations in soil ice content.

[7] Immediately after yedoma sample collection in July 1998, we thawed and initiated respiration measurements of selected subsamples in 1 L (8 L for Duvannyi Yar samples) cylinders. One set of samples was measured during one summer at a stable temperature (+4°C) to simulate temperature conditions that thawed soils would experience within the soil profile; a second set was incubated over 6 seasons at ambient temperatures (0–24°C in summer) to simulate conditions that near-surface soils would experience after permafrost thaw. Incubations naturally warmed and thawed during the 6 growing seasons (May–Oct, 1998–2003) and were frozen during the rest of the year, just as would occur in the field. Soils were also warmed for 45 days in Nov–Dec 1999, resulting in 7 warm periods during 6 calendar years. At about weekly intervals during summer, we placed a lid on the chamber and measured midday respiration rate with a LI-COR 6200 infrared gas analyzer [*Zimov et al.*, 1999]. Water was added as necessary to maintain soil moisture. We surface-sterilized another set of frozen yedoma samples from Zelenyi Mys with 96% alcohol, removed the alcohol [*Gilichinskii et al.*, 1989], and incubated them at 0°C (n = 3) and 15°C (n = 3), and measured respiration rate daily to weekly for one summer.

[8] We measured in situ soil respiration of recently thawed unvegetated loess at Duvannyi Yar (two dates) and Zelenyi Mys (one date) in July, using 0.2 m^2 plastic chambers and a LI-COR 6200 infrared gas analyzer [*Zimov et al.*, 1999]. We estimated the mass of thawed soil beneath each chamber (n = 59) from thaw depth (47 ± 6.7 cm) and a representative bulk density of 1.65 g cm^{-3} and assumed that all thawed soil contributed to the observed surface flux. Bulk density of these mineral soils varied primarily due to ice crystal content, rather than C content since the latter was relatively low. Air temperatures ranged from 8–14°C during these measurements.

[9] A separate incubation was conducted at the University of Florida with yedoma sediments that were collected and transported frozen. These incubations were conducted at a constant temperature (15°C) in contrast to the earlier incubations, and the CO_2 respired by microbial activity was purified and split for ^{13}C analysis using a GasBench attachment for a Finnigan Delta XL Plus mass spectrometer. The other split was reduced to graphite in a hydrogen atmosphere with an iron catalyst and sent to the Keck Carbon Cycle Accelerator Mass Spectrometer facility at the University of California, Irvine for ^{14}C analysis [*Vogel et al.*, 1987]. Carbon dioxide concentrations were measured with a LI-COR 6252, and percent C measured with a Costech elemental analyzer. Respiration rate is reported as the average rate observed over the first thirty days, and the isotope values reflect the integrated CO_2 released during that time period.

[10] Our calculations of total organic C in permafrost include roots and buried organic matter, in contrast to our earlier calculations [*Zimov et al.*, 1997], which were based on literature values for soils from which roots had been removed. These sediments currently occupy 1 × 10^6 km^2 (estimated by digitizing yedoma areas in Figure 1), have an average thickness of 25 m (10–60 m) (estimated from the literature), an ice-wedge content of about 50% (estimated from the literature) [*Romanovsky*, 1993; *Schirrmeister et al.*, 2002; *Sher et al.*, 2005; *Tomirdiaro*, 1980; *Vasil'chuk and Vasil'chuk*, 1998, 1997], a bulk density of 1.65 g cm^{-3} (average from our measurements), and a C content of 2.56%, (n = 71 from Siberia and Alaska). At the beginning of the Holocene, the total C content of these sediments was about 528 GT. We reduced this estimate to 450 GT for current yedoma because 50% of the area is occupied by former thaw lakes with 30% less soil C.

3. Results and Discussion

[11] Grass-dominated tundra-steppe vegetation predominated in Pleistocene landscapes north of 45°N that were free of glacial ice during cold dry glacial periods [*French*, 1996; *Velichko*, 1973]. A similar climate exists today in northeast Siberia, where grasslands occur in areas receiving regular sediment deposition [*Zimov et al.*, 1995]. Yedoma permafrost forms syngenetically through loess deposition. In other words, when wind-blown or alluvial materials are deposited on the soil surface, the bottom of the thawed soil profile, which includes roots and other organic matter, becomes incorporated into permafrost. The C concentration in frozen loess soil beneath the active layer showed relatively little variation among modern grasslands (1.27–2.13%) and was similar between permafrost (1.60 ± 0.07%, n = 30) and the

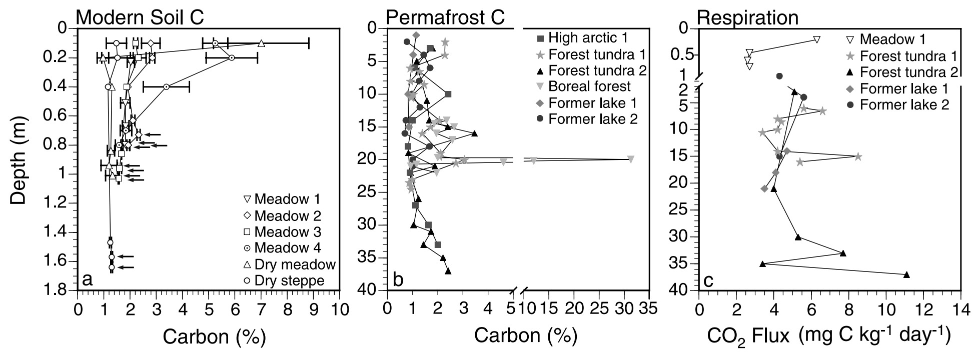

Figure 2. Vertical profiles of organic C content and respiration rates from Northern Siberian soils. (a) Modern grassland soils receiving contemporary sediment inputs. Data are means (±SE) of 3 profiles per site. Arrows denote samples within a profile that are currently in permafrost; all other samples are within the seasonal active layer. (b) Pleistocene-aged permafrost sediments beneath five profiles that never thawed since their formation and beneath former thermokarst lakes formed by melting of ice wedges during the Holocene (causing the soil thickness to decrease). These latter sediments later refroze after the lakes drained. At Stanchikovskii Yar (boreal forest), the middle of the profile had a high C content because it was a buried organic soil from an interstadial. There can be many such layers in yedoma exposures [*Schirrmeister et al.*, 2002; *Tomirdiaro*, 1980; *Vasil'chuk*, 2004]. (c) Soil carbon dioxide flux from laboratory incubations of thawed permafrost sediments and modern grassland soils. Respiration rate is reported as the average rate observed over the first summer of incubation (average soil temperature = 4.0°C).

base of the overlying thawed soil (1.66 ± 0.10%, n = 18; Figure 2a), indicating that soils do not change their C content as they become incorporated into permafrost in response to surface mineral deposition.

[12] The average C content of Late Pleistocene permafrost loess was 2.38% ± 0.38 (n = 57) across a climatic transect from arctic tundra to boreal forest along the Kolyma River in northeast Siberia (Figure 2b). The average C content of these sediments was intermediate between measurements from Alaskan yedoma (3.31 ± 0.32%, n = 14) and previous measurements from northeast Siberia (2.21 ± 0.28%; n = 9) [*Zhigotsky*, 1982], but lower than in yedoma

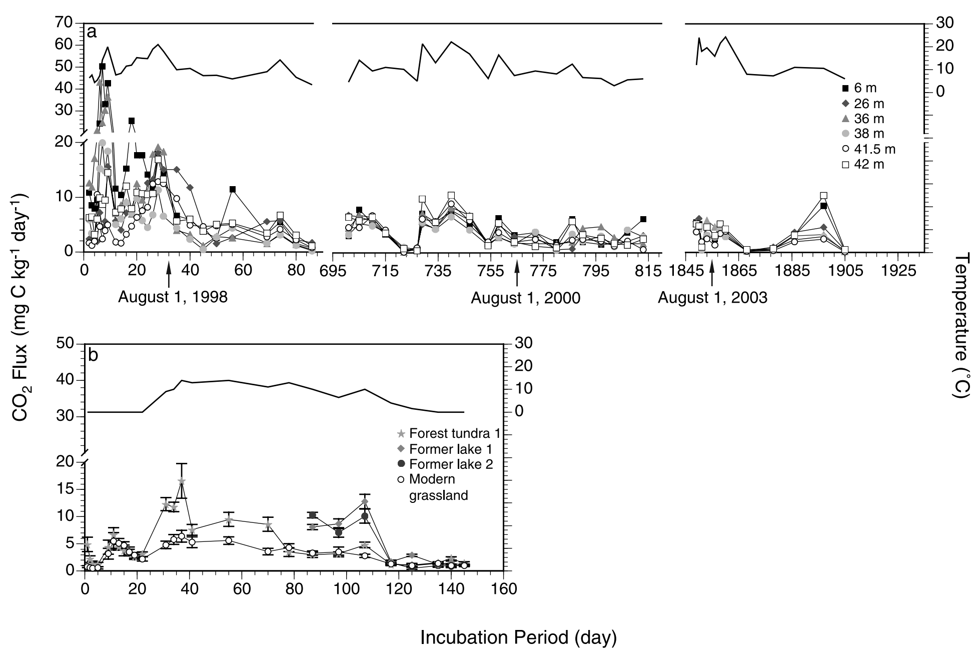

Figure 3. Time course of respiration from laboratory incubations of Northern Siberian soils. (a) Late Pleistocene permafrost sediments from 6 depths at Duvannyi Yar (forest tundra 2; n = 6) incubated at ambient air temperature (average = 11°C) for six years. Summer temperature is depicted as the blue line at the top referenced to the right axis. (b) Respiration of Pleistocene loess that had never thawed (Zelenyi Mys, forest tundra 1, n = 6); sediments beneath former thermokarst lakes (Chukochii Mys, and Duvannyi Yar; former lake 1 and 2, respectively; n = 3); and soils from modern *Calamagrostis* meadows (modern grassland; n = 12).

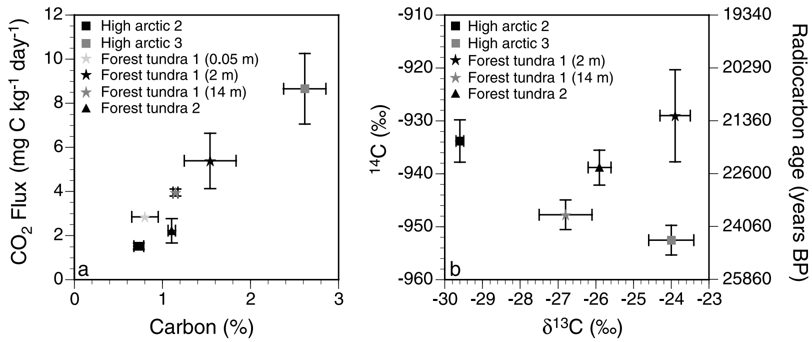

Figure 4. Laboratory incubation respiration rates and C isotopes of respired CO_2 from Northern Siberian soil incubations. (a) The relationship between respiration rate and bulk soil C content. Data are means (±SE) of 3 samples per site. Sites are a subset of those denoted in Figure 1, with the addition of a site at Sukharnaya (high arctic 2) and at Medvezhka (high arctic 3). Three depths from Zelenyi Mys (forest tundra 1) were incubated, including the current surface soil as a reference. (b) The $\delta^{13}C$ and $\Delta^{14}C$ values (left axis) of respired CO_2, with corresponding ^{14}C age (right axis). The surface soil had a positive mean radiocarbon value (modern) and is not shown.

from the Laptev seacoast (mean ~5%, minimum 2%, n = 55) [*Schirrmeister et al.*, 2002]. We also sampled soils beneath former thermokarst lakes, i.e., lakes that had formed by thawing ice-rich loess and refroze after the lakes drained. These soils had profiles that were only half as deep, due to the loss of ice volume, and had 30% less C than did permanently frozen loess, due to C loss primarily as Pleistocene-aged methane and carbon dioxide were emitted from anaerobic sediments beneath lakes when they thawed in the early Holocene [*Zimov et al.*, 1997, *Walter et al.*, 2006]. We calculate the C pool of yedoma permafrost to be ~450 GT, based on our measured C concentration and bulk density and published reports on the areal extent, thickness, and ice content of yedoma [*Romanovsky*, 1993; *Tomirdiaro*, 1980; *Vasil'chuk*, 2004; *Vasil'chuk and Vasil'chuk*, 1997]. Inclusion of this yedoma C pool stored in deep loess permafrost approximately doubles previous inventory estimates of soil C storage in northern latitudes, which typically have been estimated only for the surface 1–2 m [*Batjes*, 1996; *Gorham*, 1991; *Melillo et al.*, 1995; *Prentice et al.*, 2001; *Smith et al.*, 2004].

[13] Reactivity is just as important as pool size in governing the potential impact of yedoma on the current global C cycle. We used several independent approaches to document the respiratory loss of C when frozen yedoma thaws. Each of these experiments showed that this C was lost quite quickly when permafrost thawed:

[14] 1. In situ respiration of recently thawed unvegetated loess averaged 2.3–5.3 g C m^{-2} d^{-1} (3.0–6.0 mg C kg^{-1} of soil d^{-1}) in July, greater than the mid-summer rates currently observed in soils from productive ecosystems in the region [*Zimov et al.*, 1999].

[15] 2. In the laboratory, we measured respiration rates at natural soil temperatures of soils collected from the sites described above: (a) frozen loess with Pleistocene ice wedges, (b) refrozen loess from the sediments of a former thermokarst lake, and (c) modern grassland soils and associated permafrost. All sediments (collected at depths up to 50 m) had high respiration rates (3–11 mg C kg^{-1} of soil d^{-1}; Figure 2c). The respiration rates of Pleistocene sediments were generally higher than those of lower horizons of productive grassland soils. Soil respiration in these incubations remained high for 6 summer seasons of measurement (Figures 3a and 3b), indicating that C availability to microbes remained relatively high throughout the experiment, a result also observed with modern boreal and tundra soils of high C lability [*Neff and Hooper*, 2002; *Weintraub and Schimel*, 2003]. The average respiration in the sixth year was 3.18 mg C kg^{-1} d^{-1} (average temperature 14.8°C). The respiration rates in this experiment include bursts of respiration associated with initial thaw and annual freeze-thaw events (as would occur naturally) but are conservative because we did not measure winter respiration.

[16] 3. Frozen loess that we surface-sterilized to remove possible contamination by modern microbiota respired at rates of 2.3 mg C kg^{-1} d^{-1} at 0°C and 4.4 mg C kg^{-1} d^{-1} at 15°C, indicating rapid initiation of respiratory activity of live microbes present in Pleistocene-aged permafrost [*Gilichinskii et al.*, 1995].

[17] 4. The unvegetated loess at Zelenyi Mys that had thawed eight years previously had 39% less soil C (1.10 ± 0.07%, n = 5), averaged across the depth profile, than adjacent sediments that had remained frozen (1.71 ± 0.20%, n = 3; same bulk density), indicating a rapid and substantial loss of C after thaw of Pleistocene-aged loess. The thawed soil had a respiration rate that was 40% less than that of the permafrost soil (and proportional to remaining soil C content; data not shown), when incubated at the same temperature.

[18] Another set of incubations, described in more detail by *Dutta et al.* [2006], showed that yedoma respiration rate increased linearly with soil C content, and that the source of this respiration was ancient C (21000–25000 yr BP) that became available to microbes when the soil thawed (Figures 4a and 4b).

[19] Each of these experiments showed that Siberian permafrost C is quite decomposable when thawed, both from in situ and in laboratory incubations. The large quantities of readily decomposible C currently present in frozen loess have important implications for atmospheric CO_2 concentrations. The respiration rates we observed (>3 mg C kg^{-1} d^{-1}) are equivalent to 1.1 kg C (T soil)$^{-1}$ yr^{-1}, assuming the minimum in situ respiration that we observed in recently thawed yedoma. The average C content that we measured in yedoma

soils of Siberia and Alaska was 25.6 kg C T^{-1} (n = 71). Thus, the typical release time for C in yedoma could be a few decades after it thaws if conditions were similar to the incubations. Substantial permafrost thaw is plausible in central Siberia, where permafrost temperature (0 to −3°C) has warmed to near the melting point [*Romanovsky et al.*, 2001]. Thawing of permafrost can be self-sustaining due to changes in surface energy balance and the heat released by soil respiration that add heat and continue the thawing process [*Zimov et al.*, 1996]. The recent warming [*Briffa et al.*, 1995] and increased fire frequency [*Kasischke et al.*, 1999; *Vlassova*, 2002] that have occurred in Siberia in the past 30 years could trigger permafrost warming and thaw similar to recent trends in Alaska [*Romanovsky et al.*, 2002] and projections for the panarctic [*Lawrence and Slater*, 2005].

[20] Organic C stored in permafrost has unique properties that influence its role in the global C budget. The permafrost C pool accumulates slowly at a rate controlled by sedimentation, can be preserved for hundreds to millions of years when frozen, and, although deep in the soil profile, is highly decomposible and can be released quickly when thawed. These properties suggest that factors inducing high-latitude climate warming should be mitigated to minimize the risk of a potentially large CO_2 release that would cause a positive feedback to climate warming.

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