East Siberian Arctic inland waters emit mostly contemporary carbon

Inland waters (rivers, lakes and ponds) are important conduits for the emission of terrestrial carbon in Arctic permafrost landscapes. These emissions are driven by turnover of contemporary terrestrial carbon and additional pre-aged (Holocene and late-Pleistocene) carbon released from thawing permafrost soils, but the magnitude of these source contributions to total inland water carbon fluxes remains unknown. Here we present unique simultaneous radiocarbon age measurements of inland water CO₂, CH₄ and dissolved and particulate organic carbon in northeast Siberia during summer. We show that >80% of total inland water carbon was contemporary in age, but pre-aged carbon contributed >50% at sites strongly affected by permafrost thaw. CO₂ and CH₄ were younger than dissolved and particulate organic carbon, suggesting emissions were primarily fuelled by contemporary carbon decomposition. Our findings reveal that inland water carbon emissions from permafrost landscapes may be more sensitive to changes in contemporary carbon turnover than the release of pre-aged carbon from thawing permafrost.
Inland waters are highly abundant and important pathways for the export of terrestrial carbon (C) in northern high-latitude permafrost landscapes where ~50% (~1300 Pg C) of the global soil organic C pool is stored1–2. This C is increasingly vulnerable to destabilization and release due to permafrost thaw driven by rising Arctic air temperatures3–4. Inland waters in stable permafrost landscapes primarily receive terrestrial C from contemporary biological turnover within seasonally thawed topsoils3–7. As these landscapes warm, it is likely more contemporary C will be released to inland waters from sustained and enhanced biological turnover and event-based vegetation die-off (Arctic greening). As permafrost soils thaw, additional old C that had remained frozen since the Holocene is also released to inland waters7. Twenty-five percent of permafrost soil C is located in regions with substantial deposits of even older, ancient Pleistocene-aged sediments called Yedoma that is particularly vulnerable to thaw due to its high ice content, the majority of which is located in northeast Siberia1. These contemporary terrestrial C and old and ancient (pre-aged) C sources are released to inland waters in four main forms: dissolved and particulate organic C (DOC < 0.2–0.7 μm in size <POC), dissolved inorganic C (including CO2), and CH4 (ref. 2). Shifts in the ratio of these C forms from one to another, or increases in total inland water C concentrations, can drive the magnitude of potential climate feedbacks2,10,11.

Terrestrial C can be readily decomposed to CO2 and CH4 during transport and storage in inland waters and this is thought to be an important driver of permafrost landscape C emissions. Radiocarbon (14C) based studies in Yedoma regions have shown that pre-aged C is highly vulnerable to microbial decomposition to CO2 when released into streams12–14. At these sites, fluxes from targeted thermokarst lakes (formed from abrupt thaw processes) were an order of magnitude greater than respired organic matter (Fig. 2a) as detrital inputs of recent and ancient OM, which increase the level of correlation (R2 = 0.54–0.94; Supplementary Fig. 2). Including the thermokarst and Yedoma samples increased the level of correlation (R2 = 0.74–0.98; Supplementary Fig. 3), but CO2 and CH4 were significantly older (29,355 ± 2967 to 1234 ± 38 yBP; Fig. 2b).

Results
Inland water carbon isotope compositions. The 14C ages across the different sampling locations revealed an age gradient ranging from modern (post-1950 CE) to ancient (29,355 ± 2967 yBP; mean ± 1σ; Fig. 2a): pond sites were the youngest, followed by increasing 14C ages in the fluvial, small lake, thermokarst lake and Yedoma meltwater sites (Fig. 2b). Pond, fluvial and small lake 14C ages were statistically alike, ranging from 1679 ± 38 yBP to modern; thermokarst lake and Yedoma meltwater 14C ages were significantly older (29,355 ± 2967 to 1234 ± 38 yBP; Fig. 2b).

14CO2 values were consistently younger than the other 14C forms and 14CH4 was generally younger than PO14C, based on linear regressions testing for age divergence between the different 14C forms. These regressions were sensitive to extreme values due to limited sample numbers, so thermokarst lake and Yedoma 14C values were excluded initially. All 14C forms were significantly correlated with one another (p < 0.05; R2 = 0.54–0.94; Supplementary Fig. 2). Including the thermokarst and Yedoma samples increased the level of correlation (p < 0.001; R2 = 0.74–0.98; Supplementary Fig. 3), but CO2 and CH4 were still younger than DOC and POC (except thermokarst lake CH4 and POC, which were approximately the same; Fig. 2a).

The 613C–CO2 values reflect mixing of respiration of recent and decomposed organic matter with some contribution from methanogenesis as well24,25. The less negative 613C–CO2 values (−22.9 to −12.1‰) in comparison with CH4 were consistent with CO2 samples collected from Canadian Arctic inland waters7. The least negative 613C–CO2 values indicated there could be some exchange with atmospheric CO2 which has a less negative 613C–CO2 value (−9‰) than respired organic matter (−30.0 to −16.2‰)19,22,26, but all sites were supersaturated with CO2 compared with the atmosphere, limiting ingress of atmospheric CO2 into the water. The 613C–CH4 values indicated production from microbial methanogenesis (−63.8 to −46.7‰), although the least negative value was close to thermogenic CH4 (−45 to −20‰)24 but this could also reflect enrichment by partial methane oxidation19,24. The isotopic separation factor (εC) of 613C–CO2 and 613C–CH4 (εC = 613C–CO2 − 613C–CH4) can indicate the pathway of CH4 production. The εC values in this study were 32.7–45.6 (42.1 ± 4.8 [median ± 1σ]; Supplementary Data 1), which indicates CH4 formation from bacterial methyl-type fermentation (decomposition of methylated organic matter compounds; Fig. 2a) as opposed to the reduction of carbonate, thermogenic sources, or exchange with the atmosphere24. Geological contributions to CO2 or CH4 isotopic signatures are very unlikely at the study site (see Methods). The 613C–DOC values (−29.9 to −26.2‰) were consistent with freshwater DOC derived from the C3 photosynthetic pathway7,25,27; 613C–POC values were more negative...
The DOM was primarily derived from terrestrial material, and both microbial activity and photo-oxidation played a role in its decomposition. Relatively high humic content (Humification Index [HIX] values = 0.6–0.7), indicates a predominantly terrestrial origin, as do substantial contributions of parallel factor analysis (PARAFAC) components C1, C3, and C4 to the total DOM fluorescence signal, which are associated with high aromaticity and intermediate/high-molecular weight terrestrial DOM (Supplementary Fig. 5). Higher microbial contributions to the DOM pool were associated with older DOC, indicated by increasing biological index (BIX) values and decreasing humic content with older ages (Supplementary Fig. 6). This suggests increased microbial activity that could include the decomposition of organic C to CO2 and CH4, and/or the incorporation of organic C into microbial biomass. Photo-oxidation likely contributed to the decomposition of younger DOM which had higher molecular weight and aromaticity—characteristic of photosensitive molecules—than older DOM. This trend in molecular weight is shown by an increase in SR (slope ratio) and E2/E3 ratio with older ages, both inversely proportional to molecular weight, alongside a decrease in percent C1 (proportional contribution of PARAFAC component C1 to the total significant relationship, waters with higher overall C concentrations tended to be younger (Fig. 4).

Dissolved organic matter characteristics. We used absorbance and fluorescence spectroscopy to determine the source and decomposition pathways of inland water dissolved organic matter (DOM) which includes DOC, the C form with the highest concentrations. DOM can be produced by terrestrial and microbial production or decomposed by microbial degradation and photo-oxidation.

Contemporary versus pre-aged carbon sources. A five-source isotope mass balance model was used to estimate relative contributions to the measured 14C values from C sources measured at the study site. These C sources were defined as contemporary (modern atmospheric C, topsoil and late-Holocene basal peat: 2013 CE to 1180 yBP) and pre-aged C (old Holocene C and ancient Yedoma C: 1765 yBP to >27,920 yBP) (Fig. 2a; see Methods). The majority of inland water C observed in this study was contemporary, but pre-aged C contributed to all inland water C forms where permafrost thaw occurred (Fig. 2a). Pond, fluvial and small lake C was primarily derived from modern and topsoil C sources (>56% ± 33% [mean ± σ]; Fig. 3; Supplementary Table 2). When basal peat sources were included, these three contemporary sources accounted for >83 ± 17% of the C in these waters (Table 1). In the thermokarst lake and Yedoma meltwater, the majority of C was pre-aged (>50% ± 35%) (Fig. 3; Table 1).

Carbon concentrations. DOC had the highest concentrations of the four C forms measured (15.8 ± 12.3 mg C L−1 [mean ± σ]; Fig. 4; Supplementary Fig. 4). POC and CO2 concentrations were statistically alike (2.6 ± 3.0 mg C L−1 versus 2.8 ± 3.2 mg C L−1), lower than DOC but higher than CH4 (0.1 ± 0.3 mg C L−1). Pond C concentrations were consistently higher than fluvial, small lake or thermokarst lake sites, with the latter three containing statistically alike concentrations (Fig. 4). Yedoma meltwater had the highest organic C concentrations (DOC = 42.5 mg C L−1, POC = 8.5 mg C L−1; Table 1). Sites with higher CO2 concentrations had younger 14CO2 ages (p < 0.05, R2 = 0.71; the best fit was exponential; Fig. 4c). While this was the only statistically
TABLE 1 Carbon concentrations/fluxes, upscaled emissions and sources during the study period (23 July–17 August 2016).

| Sample location          | DOC (mg L⁻¹) | POC (mg L⁻¹) | CO₂ (mg C m⁻² d⁻¹) | CH₄ (mg C m⁻² d⁻¹) | Coverage of study area | Upscaled emissions to the atmosphere (mg C) |
|--------------------------|--------------|--------------|--------------------|--------------------|------------------------|---------------------------------------------|
| Pond                     | 25.5 ± 11.8  | 3.6 ± 1.6    | 3.64 ± 2.5         | 3.04 ± 1.5         | 12.1 ± 1.9             | 2784 ± 349                                  |
| Fluvial                   | 2.4 ± 12.1   | 1.9 ± 1.4    | 3.54 ± 11.8        | 3.04 ± 1.5         | 1.2%                   | 897.4 ± 115.9                               |
| Small lakes              | 8.2 ± 12     | 1.4 ± 0.8    | 10 ± 10            | 10 ± 10            | 0.1%                   | 83.0 ± 17                                   |
| Thermokarst lake         | 6.5 ± 12     | 2.4 ± 0.8    | 3.54 ± 12          | 2.4 ± 0.8          | 1.5%                   | 12.1 ± 1.9                                  |
| Yedoma meltwater         | 4.2 ± 12     | 6.5 ± nd     | -2784 ± 239        | -7284 ± 239        | 89.4%                  | 897.4 ± 115.9                               |
| Tundra                   | nd           | nd           | nd                 | nd                 | nd                     | nd                                          |

All values are mean ± 1 standard deviation. CO₂ and CH₄ fluxes were measured using a closed cuvette and chamber system, respectively. The coverage of study area was calculated as the area of the site in km². The upscaled emissions were calculated as the area-weighted mean of the carbon sources in the site. DOC concentrations were measured using a high-performance liquid chromatography system, and POC concentrations were measured using an ICP-OES system. CO₂ and CH₄ concentrations were measured using a gas chromatograph system.

Discussion

We demonstrate that the inland waters in this permafrost landscape were primarily fueled by contemporary C. We also present direct evidence that pre-aged C can be simultaneously released in all four main inland water C forms where permafrost thaw is occurring. CO₂ and CH₄ tended to be younger than DOC and POC, indicating that contemporary C fuels summer inland water emissions in the study region. This is due to the relatively high abundance of contemporary organic C relative to pre-aged C, and contemporary DOC being more readily photo-oxidized than pre-aged C. The flipside of this finding is that pre-aged inland water C emissions described in previous studies may be caused by the absence of contemporary C as much as by the vulnerability of pre-aged C to microbial decomposition. Given the predominance of labile contemporary C in the inland waters observed in this study, these systems show a high connectivity with contemporary terrestrial C, and are thus likely more sensitive to shifts in contemporary C cycling than the cycling of thawed pre-aged C.

Evidence for contemporary C fueling inland waters in this permafrost landscape is demonstrated by the pond, fluvial, and small lake sites having the largest upscaled inland water C emissions (equivalent to 2.3 ± 1.5% of the tundra C sink), of which >80% was contemporary C (Table 1). In general, sites with higher C concentrations tended to be younger (Fig. 4). However, the high proportion of pre-aged C contributing to the thermokarst lake and Yedoma meltwater demonstrates that permafrost thaw will release old and ancient C to all inland water C forms. Across this gradient of inland water ¹⁴C ages (Fig. 2b), the generally close relationship of CH₄ with DOC and POC with DOC suggests that C release across these different forms was consistently from similar sources. The divergent relationship between DO¹⁴C/PO¹⁴C and CO₂ and CH₄ and PO¹⁴C and CH₄ was driven by the younger age of the gaseous C (Fig. 2a). This implies that the C fueling inland water CO₂ and CH₄ emissions was generally younger than mobilized DOC and POC⁷,32, and that the most labile organic matter pools were also the youngest.
Permafrost thaw can result in high concentrations of low molecular weight DOC and high POC concentrations being released to inland waters\(^2,12\). Incubation-based \(^{14}\)C studies have suggested that this pre-aged C may be rapidly decomposed to CO\(_2\) (refs. 12–14). Our simultaneous in situ \(^{14}\)CO\(_2\) and \(^{14}\)CH\(_4\) observations were generally younger than DOC and POC suggesting that contemporary C is also being rapidly decomposed in thaw affected inland waters, potentially at a faster rate or in greater proportion than pre-aged C. The utilization of pre-aged C by microbes previously demonstrated in streams draining thawing ancient Yedoma sediments\(^12,13,14\) could therefore just be due to the limited availability of contemporary organic C in those systems. This effect is amplified by these studies targeting streams heavily impacted by thermokarst activity, and the incubation approaches used which restrict contemporary inputs entirely. Our findings show that the large amount and high lability of contemporary C mean that it can dominate C cycling compared with thawed pre-aged C in permafrost systems at the landscape-scale.

The majority of terrestrial C released into the study inland waters was in the form of DOC (Fig. 4)\(^3\). Our DOM spectroscopy results indicated that the decomposition pathways for DOC was dependent on its source: younger DOC was more vulnerable to photo-oxidation whilst older DOC was more vulnerable to microbial decomposition. Pre-aged permafrost DOC is known to be readily decomposed by microbes\(^12,13,14\), and this is supported by the older CO\(_2\) and CH\(_4\) ages observed in our thermokarst lake and Yedoma meltwater (Fig. 2a). Photo-oxidation can also actively decompose DOC in Arctic inland waters\(^34\), but is less effective at decomposing ancient Yedoma C\(^35\). Our samples were collected in late summer when daylight hours can extend for virtually the whole day, so photo-oxidation was likely an

![Diagram](https://example.com/diagram.png)
Decomposition of pre-aged organic C has most notably been demonstrated in CH$_4$ ebullition from thermokarst lake sediments that was up to 42,900 yBP in age$^{15}$. In the present study, dissolved CO$_2$ emissions were larger and younger than dissolved CH$_4$, consistent with previously observed Arctic inland water emissions ($3–6420$ mg C–CO$_2$ m$^{-2}$ d$^{-1}$ versus $1.0–182$ mg C–CH$_4$ m$^{-2}$ d$^{-1}$)$^{6,7,36–40}$. Both dissolved CO$_2$ and CH$_4$ emissions were younger and tended to be smaller than CH$_4$ ebullition ($0–255$ mg C–CH$_4$ m$^{-2}$ d$^{-1}$)$^{40}$ which was not measured in our study. CH$_4$ ebullition is usually generated at depth, as opposed to dissolved CH$_4$ which represents a spatiotemporally integrated signal of lake sediment and terrestrial inputs$^{19}$. The generally older age of CH$_4$ ebullition compared with dissolved CH$_4$ indicates that where contemporary organic C is limited, such as deep unfrozen lake sediments (taliks), pre-aged C will drive CH$_4$ production$^{15}$. For example, in Alaskan lakes, dissolved CO$_2$ and CH$_4$ collected in winter were older than in summer. This was because frozen lake surfaces prevented inputs of contemporary terrestrial C and stopped ebullition reaching the surface, which allow these bubbles containing old CH$_4$ and CO$_2$ to dissolve into the water column (refs. $^{6,19}$).

These age differences further indicate that microbial utilization of pre-aged C is due to the restricted supply of contemporary organic C, rather than preferential mineralization of pre-aged organic matter.

Inland waters can rapidly turnover contemporary organic C$^{32,41}$, and the few existing in situ $^{14}$C measurements of dissolved CO$_2$ and CH$_4$ across the Arctic have been contemporary in age$^{6,7}$. The export of pre-aged C by inland waters from thawing permafrost landscapes may instead occur as DOC and POC. Our comparison of inland water CO$_2$ and CH$_4$ emissions with the measured terrestrial sink does not account for lateral transport of organic C. With DOC and POC concentrations consistently c. 4 times higher and generally older than diffuse CO$_2$ and CH$_4$ (Figs. 2 and 4), lateral transport is a potentially important component of landscape C loss in the region. Mobilized DOC and POC may be transported laterally to lake, fluvial or marine systems where it can be decomposed to CO$_2$ and CH$_4$ or buried in sedimentary reservoirs$^{2,42}$. Rates of lateral transport may be greater at other times of the year than our study period depending on hydrological regimes, e.g., during the Spring snow-melt pulse, and as a result of abrupt thaw events$^2$. It is therefore important to track all C forms across different times of the year to fully capture the export of pre-aged permafrost C by inland waters.

Our study region had a slightly lower proportional coverage of inland water bodies compared with a selection of Arctic regions in a recent study: of 31 sites, 29 had higher proportional inland water coverage but only by 4.5 ± 3.9% (mean ± 1σ)$^{23}$. Inland waters types and their areal coverage in permafrost landscapes could increase or decrease under climate change depending on localized topography and drainage patterns$^{4,21,23,43}$. There is thus a need to capture C dynamics across a range of water types within the same landscapes, as shown by the variable C concentrations and isotopic compositions seen across our study sites.

Our observations demonstrate that inland waters in permafrost landscapes are sensitive to both contemporary and pre-aged C release, although contemporary C dominates in this landscape. This means that regardless of whether the changing Arctic climate will increase inputs of contemporary or pre-aged C to inland waters, their annual CO$_2$ and CH$_4$ emissions will likely go up, offsetting terrestrial C sinks. In the study region, inland water CO$_2$ and CH$_4$ emissions were a small offset of the terrestrial sink (2.3 ± 1.5% in total), so large increases in inland water coverage and supply of terrestrial C to these waters is needed in order for them to significantly offset the terrestrial sink.
Fig. 4 Concentrations of all carbon forms. a Dissolved organic carbon (DOC). b Particulate organic carbon (POC). c Carbon dioxide (CO2), and d Methane (CH4) for each sample location; the thick horizontal lines represent the median, limits of the boxes represent upper and lower quartiles, whiskers extend to 1.5 times the interquartile range, dots represent all data points, letters indicate statistical differences between the sample locations using ANOVA (see Methods). Inset: relationship between 14C content and carbon concentrations (absorbance at 350 nm [a350] for DOC; Supplementary Fig. 4)46, excluding Yedoma meltwater (see Methods) (*indicates P < 0.05).

Methods
Study approach. The study site is in the oligotrophic tundra of the Kytalyk Nature Reserve in the Indigirka River lowlands (70.83°N, 147.49°E; Fig. 1). Average active layer depths in the tundra around and beneath the ponds during the study period, July–August, were 26 ± 12 cm (mean ± sd, n = 370; Table S1). The landscape is dominated by a series of drained lake basins formed by thaw driven subsidence and drainage21, and the floodplain of the Berelekh River, a major tributary to the Indigirka. A single Yedoma ridge (up to 30-m high) lies within the study region, into which the thermokarst lake is eroding and from which the Yedoma meltwater samples were collected. This ridge is an remnant of organic-rich, ~2–6% C by mass, Pleistocene-aged permafrost sediments deposited as loess or fluvial silts with ice content 30–90% by volume22. The ridge covers ~12% of the landscape surface and is potentially distributed throughout the region as a result of thaw and erosion22 (Fig. 1); such deposits are common throughout northeast Siberia1.

The study region is underlain by continuous permafrost limiting potential geological inputs to aquatic CO2 or CH4, and neither 14CO2 nor δ13C-CO2 (which are in isotopic equilibrium with total dissolved inorganic C) correlated with geochemical indicators of rock weathering (e.g., water Ca2+ concentrations). Geological inputs, such as rock weathering47 or CH4 seeps48, were therefore unlikely to have influenced our observations.

Mean air temperatures during the study period (25 July–17 August 2016) were 10.3 ± 3.3 °C, 7.6 ± 4.7 °C during the growing season (June–September 2016), and −12.4 ± 16.7 °C in the year preceding the sampling campaign. These values are within normal temperature ranges for the site (monthly averages range from 9.7 °C in June to −34.2 °C in January)22. Precipitation at the study site occurs mainly as rain during the summer, with an annual mean of 232 mm22; during the sampling period, 30 mm of rain fell compared with 129 mm over the 2016 growing season (Supplementary Fig. 1). Vegetation in the tundra areas of the site consists primarily of Eriophorum vaginatum and Eriophorum angustifolium, Betula nana, Carex sp., and Sphagnum sp. The ponds tend to be surrounded by E. vaginatum, E. angustifolium, B. nana, Carex sp., and Sphagnum sp., while the fluvial systems are bordered primarily by Arctophila fulva, Arctagrostis latifolia, and Salix pulchra49.

Sampling and analyses. We collected concurrent 14C, δ13C, and C concentration samples for DOC, POC, and dissolved CO2 and CH4 from fluvial, pond, small lake and thermokarst lake sites, and Yedoma meltwater. Samples were collected from the edges of water bodies as could be accessed from the shoreline. Multiple C concentration samples were collected from the lake and stream locations to account for spatial variability. Only one location was sampled for 14C and δ13C in each of the study lakes due to the prohibitive cost of analyses. However, 14C in dissolved CO2 and CH4 have been shown to be relatively uniform in lakes of equivalent size in Alaska19. The stream was sampled in two locations and the five different ponds were sampled for 14C and δ13C, but again widespread sampling was limited due to cost. Ponds were only sampled once since their small size means that variations in C source are likely small (Table S1). DOC and POC, and dissolved CO2 and CH4 concentration samples were collected three times from each site during the study period (except Yedoma meltwater, where n = 1) at three-day intervals in daylight hours for consistency. DOM samples were also collected at the same frequency as the concentration samples but only at sites where 14C samples were collected (Fig. 1; except Yedoma meltwater, where n = 1). 14C and δ13C samples were collected at a single time point from a selection of these sites between 05 and 09 August 2016 (Fig. 1; Supplementary Data 1).50–54; additional δ13C samples were...
also collected for CH4 and DOC for quality control15–17 (see Supplementary Methods; Supplementary Fig. 8; Supplementary Data 1 and 2).

**Isotope mass balance model.** The Stable Isotope Mixing Models in R (simmer) package28 was used to estimate relative contributions of potential C sources to 14C samples from each group of sample sites (fluvial, pond etc.; Supplementary Table 2). The isotope mass balance model was only used for 14C because δ13C is fractionated during CO2 and CH4 production, and so a dual isotope mass balance, common in aquatic organic matter studies, would not work. This approach therefore provides the theoretical possible contributions of different sources following previous studies29, but here the sources are further constrained by using measured soil organic matter 14C values obtained from the study region22. While some water bodies, for example the ponds (see Introduction), are unlikely to receive contributions daily from pre-aged C sources, we applied the isotope mass balance model with the same end-members to all samples for consistency to demonstrate all possible source contributions that could produce the observed 14C values. We used simm to carry out a mass balance (Eq. 1), fitting the potential contributions of five different C sources (see below) using Markov chain Monte Carlo modeling. Ranges and uncertainties in the 14C signature of the different sources were propagated through the model, and model fitting carried out by the Just Another Gibbs Sampler code. Model convergence was only considered acceptable when the upper confidence interval of the final model outputs was 1.00 ± 0.10 (ref. 29). We excluded any food web specific functions from the mass balance, such as trophic enrichment factors, or preferential source contributions29.

\[
\text{C}_{14\text{group}} = C_{14\text{stored}} + C_{14\text{topsoil}} + C_{14\text{basalpeat}} + C_{14\text{dil}} + C_{14\text{ancient}}
\]

(1)

\[
\text{C}_{14\text{group}} \text{represents the } 14\text{C content of the different inland water C components within each group of sample sites (pmC; Supplementary Data 1), which is the sum of the mass balance between the proportional contributions of five different C sources; to achieve mass balance, the relative contributions of the five different sources is assumed to sum to 1. These five sources follow the model used by ref. 26 but is adapted here to incorporate soil/sedimentary 14C data obtained from the current study site22.}
\]

**Inland water CO2 and CH4 emissions.** Inland water CO2 and CH4 fluxes (F) were calculated using Eq. (2):

\[
F = \Delta C \kappa
\]

(2)

where \(\Delta C\) is the difference between the observed gas concentration in the water and the expected concentration if the dissolved gas was in equilibrium with the atmosphere, and \(\kappa\) is the gas transfer velocity from water to the atmosphere.

Atmospheric CO2 and CH4 concentrations were taken from the ambient air samples collected over the course of the study period and complemented with atmospheric concentration data from the eddy covariance tower. Lake \(k_{\text{water}}\) values were calculated using the relationship with wind speed at 10 m (\(U_{10}\) in m s\(^{-1}\)) and \(k_{\text{water}}\) (k normalized to a Schmidt value of 600) from ref. 66:

\[
k_{\text{water}} = 2.07 + 0.215 \times U_{10}^{0.7}
\]

(3)

\(U_{10}\) values were available from the eddy covariance site at 30 min intervals49. Pond \(k_{\text{water}}\) values were taken from ref. 61 (0.36 cm h\(^{-1}\)), a low \(k_{\text{water}}\) value in the context of inland water \(k_{\text{water}}\) values. Following ref. 50, for fluvial sites we used the boreal and Arctic stream \(k_{\text{water}}\) value from ref. 61 (13.1 cm h\(^{-1}\)), which is at the lower end of fluvial \(k\) values but at the high end of inland water \(k\) values in general. The \(k_{\text{water}}\) values can then be converted to:

\[
\kappa = \frac{k_{\text{water}}}{\text{Sc}}
\]

(4)

where Sc is the Schmidt number in freshwater computed from water temperature.

**Tundra CO2 and CH4 emissions.** Tundra fluxes at the study site were calculated using the eddy covariance method26. The eddy covariance tower at the study site (Fig. 5) was installed in 2013 to measure net ecosystem exchange, using an ultrasonic anemometer (Gill Instruments, Lymington, UK, type R3-50) for wind speed and temperature, and water vapor and CO2 concentrations were measured by an open-path infra-red gas analyzer (LI-COR LI-7500, Lincoln, NE, USA)49. An open-path infra-red gas analyzer (LI-COR LI-7700) has been used to measure atmospheric CH4 concentrations since 2014. All sensors measure every 10 Hz and are installed at a height of 4.7 m. Energy balance closure and good co-spectral shapes required for validation of the eddy covariance methods were demonstrated at the site previously42,49.

Tundra fluxes were calculated via EddyPro (version 6.2.2) to provide 30 min mean fluxes. Processing configuration included: 2D rotation of sonic anemometer coordinates, compensation of flux fluctuations due to temperature and pressure spectrometric corrections (for Li7700 – CH4 flux only) and spectral corrections. Fluxes of CO2 and CH4 associated with implausible values of gas concentrations, wind speed or sonic temperature caused by frost, precipitation, water condensation on the analyzer lenses or other conditions of sensors disturbance were removed.

Trends were additionally de-tanked following the approach of Dai et al.28. Daily fluxes (nmol CO2 m\(^{-2}\) s\(^{-1}\) and nmol CH4 m\(^{-2}\) s\(^{-1}\)) were summed and daily standard deviations propagated (by calculating the square root of the sum of squares of daily 1σ uncertainties) for the 23-day study period (25 July–17 August 2016), then converted to a daily C flux (mg C m\(^{-2}\) d\(^{-1}\)). We sampled plus the surrounding tundra, but we did not try and upscale beyond this. The footprint of the eddy covariance tower captures an area of tundra that reflects the majority of land surrounding the sampled inland water locations, and is considered to be representative of the tundra CO2 and CH4 cycling in the region48 (Fig. 1). In general, the average of the footprint region (0.2 km\(^2\)) of the small lakes (0.2 km\(^2\)) and the thermokarst lake (0.5 km\(^2\)) were calculated from Worldview satellite imagery on 15 July 2015 (Fig. 1). Pond coverage was obtained at the study site from satellite imagery and aerial photography from 2002 to 2013 site in a recent study using image classification (49.0 ± 0.2% of the study region)22. Daily fluxes of grouped sample locations (fluvial, pond etc.) were multiplied by 23 days and the relative areas of each inland water system to give a total flux to the atmosphere in Mg C for the period 25 July–17 August 2016. 1σ uncertainties in the flux estimates were propagated through the upscaling by providing maximum and minimum values in the total upcaled fluxes (\(\pm \sigma\) in Table 1). Relative contributions to fluxes from contemporary (atmospheric, topsoil and basal peat C) versus pre-aged (old and ancient C) sources were estimated by multiplying the fluxes by the proportional contributions of these sources to the total C load (Supplementary Table 2).

**Statistics.** Statistics were analyzed in R (version 3.5.1). Linear models were used to analyze correlations between groups, and the strength of relationships assessed using analysis of variance (ANOVA), with post hoc Tukey’s honestly significant difference tests (significance level: p < 0.05) carried out using the agricolae package for letters denoting group similarity in, e.g., Fig. 4. Correlations between variables were assessed using Pearson’s correlations in the dmtc package (p-values, e.g., p-values in Eq. 6). Yedoma meltwater values were not included in the correlation with C concentrations in Eq. 4 (insets) because they represent a Yedoma signal that has yet to be released into an inland water system (unlike the thermokarst lake samples), and because there was no Yedoma 14CH4 value, nor CO2 or CH4 concentration values, to allow a complete comparison of all correlations.

**Data availability** All data is available in the Supplementary information and Supplementary Data 1 and 2.

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J.F.D., T.M., R.A., J.v.H. and A.J.D. conceived the study and contributed to general logistics. J.F.D., O.H.M., L.B.M., H.L., R.v.L., R.P., S.K., J.v.H. and A.J.D. contributed to fieldwork and logistics. J.F.D., M.M.R., L.B.M., M.H.G., H.L., R.v.L., A.V.B., S.B., T.L., T.R. and J.E.V. contributed to lab work, sample analyses and data analyses. J.F.D. wrote the paper; all authors contributed to draft versions prior to submission.

Competing interests
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