Landscape-Scale Variability of Organic Carbon Burial by SW Greenland Lakes

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ABSTRACT

Lakes are a key feature of arctic landscapes and can be an important component of regional organic carbon (OC) budgets, but C burial rates are not well estimated. ^{210}Pb-dated sediment cores and carbon and organic matter (as loss-on-ignition) content were used to estimate OC burial for 16 lakes in SW Greenland. Burial rates were corrected for sediment focusing using the ^{210}Pb flux method. The study lakes span a range of water chemistries (conductivity range 25–3400 µS cm^{-1}), areas (< 4–100 ha) and maximum depths ($\sim$ 10–50 m). The regional average focusing-corrected OC accumulation rate was $\sim$ 2 g C m^{-2} y^{-1} prior to 1950 and 3.6 g C m^{-2} y^{-1} after 1950. Among-lake variability in post-1950 OC AR was correlated with in-lake dissolved organic carbon concentration, conductivity, altitude and location along the fjord. Twelve lakes showed an increase in mean OC AR over the analyzed time period, $\sim$ 1880–2000; as the study area was cooling until recently, this increase is probably attributable to other global change processes, for example, altered inputs of N or P. There are $\sim$ 20,000 lakes in the study area ranging from $\sim$ 1 ha to more than 130 km^2, although over 83% of lakes are less than 10 ha. Extrapolating the mean post-1950 OC AR (3.6 g C m^{-2} y^{-1}) to all lakes larger than 1000 ha and applying a lower rate of $\sim$ 2 g C m^{-2} y^{-1} to large lakes (> 1000 ha) suggests a regional annual lake OC burial rate of $\sim$ 10.14 $\times$ 10^9 g C y^{-1} post 1950. Given the low C content of soils in this area, lakes represent a substantial regional C store.

Key words: nitrogen; arctic; diatoms; tundra; biogenic silica; DOC; soil carbon.

INTRODUCTION

Lakes play a significant role in the terrestrial carbon (C) cycle (Cole and others 2007). As a result of the import of terrestrial C from their catchments, even lakes that are strongly heterotrophic (net ecosystem respiration > primary production) can bury considerable amounts of carbon at 10^3-y timescales (Kortelainen and others 2004). External subsidy of

lake C budgets by inputs of terrestrial C is particularly large in boreal systems and arctic lakes (Kling and others 1991; Algesten and others 2004). Arctic soils represent a major global carbon pool (Tarnocai and others 2009), and its fate in a warming world is a major unknown: Temperature increases may deepen the annual active thaw layer and increase transfer of this large soil C pool to aquatic systems (Schuur and others 2008; McGuire and others 2009). The release of C (as DOC) from terrestrial pools into surface and subsurface runoff increases input to streams and lakes, where it can be processed and released as CO_2 (Kling and others 1991; Sobek and others 2003). It is unclear whether the recent productivity increases widely postulated to occur in arctic lakes is a response to higher mean annual temperatures (Michelutti and others 2005) or enhanced deposition of reactive nitrogen (Hastings and others 2009; Wolfe and others 2013) and whether these production increases are translated into increased C burial. Similarly, it is unknown whether the increased C burial associated with greater aquatic primary production will balance the C lost (as CO_2) as soil carbon is transferred from the terrestrial to aquatic systems (Schuur and others 2008).

Sediment organic C content (measured directly as percent C or inferred from organic matter content) is widely used as a surrogate for lake productivity in paleolimnological studies (Meyers and Lallier-Verges 1999). As well as reflecting in-lake production and external (allochthonous) organic matter loading, the organic content is a function of factors controlling preservation, notably the presence and the absence of oxygen (Meyers and Lallier-Verges 1999), dilution by the inorganic matrix (which can be affected by catchment stability, shoreline erosion) (Dearing 1991), and spatial variability of sediment distribution in the basin (Rippey and others 2008; Engstrom and Rose 2013). Dilution effects can be corrected by use of dry mass accumulation rate data, if dating control is adequate.

Lakes are spatially and temporally variable with respect to productivity (for example, littoral–profundal differences, seasonality), but productivity measurements are often focused in pelagic areas, with the result that the contribution of the littoral zone to whole-lake production can be underestimated (Vadeboncoeur and others 2002). In arctic lakes with high transparency, benthic primary producers commonly contribute substantially to lake primary production (Vadeboncoeur and others 2003). Previous studies of temperate lakes have shown that sediment accumulation can be overes-

timated by the use of single, centrally located cores due to resuspension and focusing of sediment into the central areas of lakes (Rippey and others 2008). Thus, calculation of patterns of whole-lake deposition and accumulation of important elements and metals, such as Hg, P or C (Molot and Dillon 1996; Fitzgerald and others 2005; Engstrom and Rose 2013), requires cores from multiple depositional areas of the lake to adequately capture the spatial variability or application of a correction factor to account for spatial variability. This method, however, is very time-consuming. Recently, the ^{210}Pb flux method has been used to correct for sediment focusing (Lamborg and others 2013), and this approach allows for direct comparison of burial rates among lakes at the regional scale.

Here, we calculated the focusing-corrected organic C accumulation rates for 16 lakes located along Kangerlussuaq (Søndre Strømfjord in Danish), from the outer coast to the present-day ice margin (Figure 1). We consider the effect of the regional climate gradient and limnological variability on burial rates and the extent to which C burial rates have changed over the last ~ 100 years. The contribution of lake OC burial to regional C sequestration in this area is considered.

MATERIALS AND METHODS

Study Area

The area around Kangerlussuaq (Søndre Strømfjord in Danish, SW Greenland, 66–67°N, 51–53°W, Figure 1) contains $\sim 20,000$ lakes; the majority of the smaller lakes are glacially scoured basins, and there also are a smaller number of large (~ 20 km long) glacial fjord systems. The lakes are predominantly oligotrophic (< 10 µg TP L^{-1}) and dilute (conductivity < 100 µS cm^{-1}), although conductivity is higher in some lakes near the head of the fjord (around 51°W) (Table 1), where higher evaporation rates influence lake hydrological budgets (Anderson and others 2001); mean lake conductivity varies from 25 to 4000 µS cm^{-1}. The area is best known for the occurrence of a number of athalassic (non-marine) oligosaline lakes close to the end of the fjord, some of which are meromictic (permanently stratified). The lakes are primarily Na–Mg HCO_3 systems and pH varies from ~ 6 to 8.5. Dissolved organic carbon (DOC) concentration is enhanced in lakes in the central parts of the fjord system (maximum concentration can reach > 100 mg C L^{-1}), whereas lakes at the outer coast and along the ice margin have much lower DOC concentration (Anderson and Stedmon 2007). In
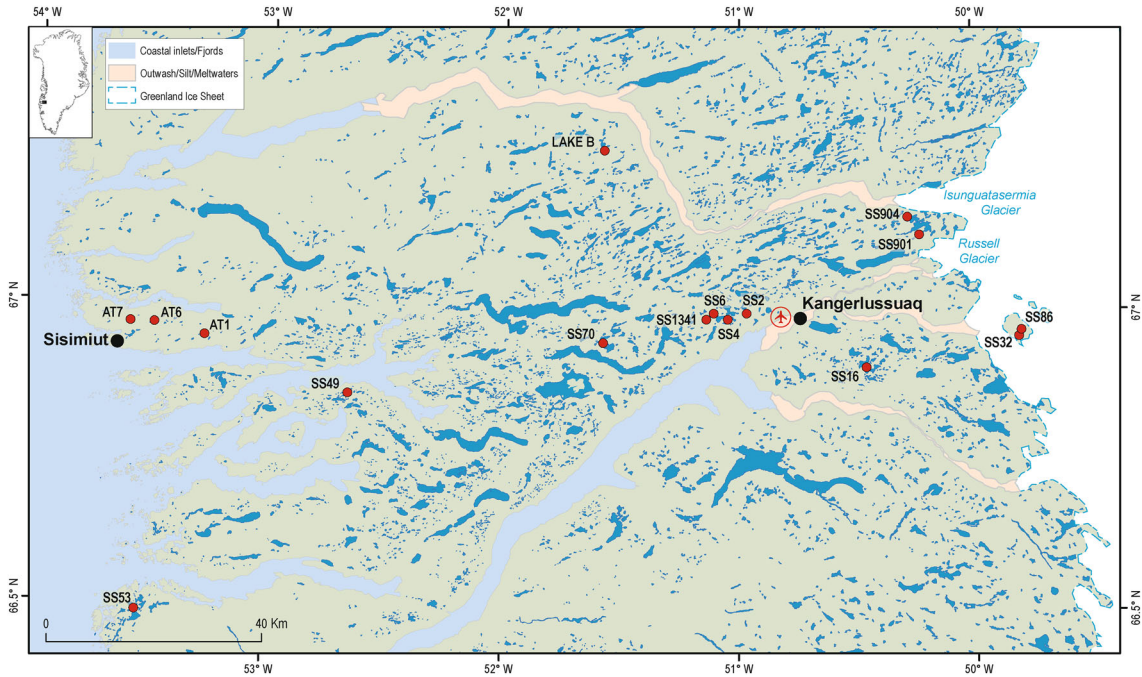
Figure 1. Location of study sites along Kangerlussuaq (Søndre Strømfjord) (SW Greenland). The allocation of the study lakes to the three local areas are indicated by different symbols

Table 1. Summary Environmental Data of the Study Sites

Lake	Latitude (°N)	Longitude (°W)	Altitude (m)	Area (ha)	Max depth (m)	DOC (mg L^{-1})	Conductivity (μS cm^{-1})	DIC (mg L^{-1})	Z_{eu} (m*)
SS32	66.97	49.80	470	16.3	21	7.6	48	4.9	9.4[#]
SS86	66.96	49.81	470	5	13	30.7	334	38.9	nk
SS901	67.13	50.23	375	12	14.5	7.6	102	9.4	16.8
SS904	67.16	50.28	425	12.4	11.4	5.8	78	4.5	20.0
SS16	66.91	50.46	477	3.3	12.2	10.2	60	6.4	6.4[$]
SS2	67.00	50.97	185	36.8	11.2	40.6	251	30.6	30.6
SS4	66.99	51.05	170	73	23	90	2538	180.0	12.0
SS6	67.00	51.11	175	21.5	12.8	106	3244	194.3	10.4
SS1341	66.99	51.14	350	7	16	16.3	338	30.2	19.5
Lake B	67.27	51.58	671	98.8	31	nk	nk	nk	nk
SS70	66.95	51.58	235	76.6	47	28	2510	52.5	nk
SS49	66.86	52.66	320	20.7	10.5	4.8	27	1.8	nk
SS53	66.49	53.53	50	7.3	8.3	–	36	2.0	nk
AT1	66.97	53.24	475	11	16.5	1.2	31	2.5	17.0
AT6	66.97	53.49	150	11.1	23	1.8	47	1.9	11.4
AT7	66.97	53.59	450	6.5	20	1.3	40	2.3	19.5

Calculated from Secchi depth.
[#]*Saros and others (2015).*
[$]*Saros and others (2016).*
nk: not known.
Sites are listed by longitude to reflect the sampling gradient from ice sheet margin to coast (Figure 1).

general, lakes have a poor macrophyte flora, although *Chara* spp. (a macro-alga) beds and filamentous green and cyanobacterial mats are substantial in the oligosaline lakes. Many lakes are fishless (see Jeppesen and others 2017).

The regional geology is mainly granodioritic gneiss with occasional ultrabasic intrusions. These geological anomalies create local variability in water chemistry, such as the high sulfate concentrations measured in lake SS70 (and neighboring

lakes) (see Anderson and others 2001; Figure 1). Climate is cool continental low Arctic, and the mean annual air temperature at Kangerlussuaq, the airport at the head of the fjord (Figure 1), is − 6°C. Precipitation inland is low ($\sim$ 150 mm y^{-1}) but higher at the coast ($\sim$ 500 mm y^{-1}). In the inland area, vegetation is dominated by dwarf shrub tundra (*Betula nana, Ledum groenlandicum, Rhododendron, Salix* spp.), but can be locally sparse as a result of moisture stress, particularly on south-facing slopes. Close to the ice sheet margin, *Ledum-S. glauca* heath and *Carex* spp. steppe are more abundant, while toward the coast, *Empetrum, Cassiope*, bryophytes and lichens are increasingly common. Soils are thin (30–50 cm) and relatively organic rich, but are generally restricted to north-facing slopes, valley bottoms and damper areas in the central, inland areas; there are occasional peat-like deposits associated with the richer vegetation growth found at lake outflows. At the ice margin and the coast, soils are much thinner. Surface runoff is low and generally limited to late May–early June, when the active layer is thawing. Aeolian activity occurs in the area, because of deflation from sand sheets adjacent to the ice sheet, and there is extensive dust deposition in inland areas (east of 52°W; Dijkmans and Törnqvist (1991)).

The lakes used in this study (Table 1) are representative of the range of lake types in the area in terms of their water chemistry, biological structure, and size and depth (Anderson and others 2001; Anderson and others 2009), with the exception of the large fjord-like lakes, which have not been sampled. Many of the lakes, presumably because of the extended period of ice cover (October until early June) and strong thermal stratification during summer, have laminated sediments, although none have proved to be annual (Anderson and others 2000). Sediment core studies indicate that the oligosaline lakes precipitated calcium carbonate in the past (McGowan and others 2003); dissolved inorganic carbon values are given in Table 1.

To allow comparison of C burial rates by location, lakes were allocated to one of three groups (coastal, inland and ice margin) based on their geographic location and water chemistry (see Anderson and others 2001; Whiteford and others 2016 for a discussion of limnological gradients). Jeppesen and others (2017) provide a comparison of water chemistry for lakes in west Greenland and those in the Zackenberg area of the northeast.

Determination of lake area and density: Lake outlines were digitized from Danish Ordnance Survey (KMI) 1:250,000 scale maps of this region, and their area was estimated using ArcView.

Field Sampling

Short sediment cores (20–30 cm in length) were taken from the deepest part of each lake using a modified Hon-Kajak gravity corer (8.4 cm internal diameter) (Renberg 1991) or a freeze core. Kajak cores were sectioned in the field at 0.5-cm intervals. Freeze cores were returned frozen to the laboratory, where they were sectioned at $\sim$ 0.5-cm intervals (Renberg 1981).

Laboratory Methods

Percentage dry weight was determined for all samples (drying overnight at 105°C), and bulk density was estimated using standard methods (Dean 1974). The carbon (C) content of the sediments of selected lakes (SS70, SS53, SS1371, SS49) was measured using a Leco CHN analyzer and is reported as percent of sediment dry mass. For the Kajak cores from the remaining sites (Table 1), loss-on-ignition (LOI), a measure of organic matter content, was determined by weight loss following heating of dried samples at 550°C for 4 h (Dean 1974). Where both approaches were used, for example, at lake SS32, the correlation between the two methods is high ($r = 0.85$, $p < 0.001$), indicating that the approaches are comparable (Bindler and others 2001b). Loss-on-ignition data were converted to carbon using a standard correction factor of 0.47 (Dean 1974). Organic carbon accumulation rates (OC AR) were determined by multiplying the dry mass accumulation rate, derived from the ^{210}Pb chronologies (below), by the proportional OC content. As surficial sediments include sediment that is undergoing post-depositional mineralization, the uppermost 10 years were removed prior to estimating mean burial rates.

Data and Statistical Analyses

To explore possible factors controlling carbon burial rates, correlations (Pearson's product moment correlation, r) between post-1950 mean OC AR_{fc} and a number of limnological and morphometric variables were estimated (Table 1). The relationship between site location (as longitude) and OC AR_{fc} was estimated by fitting a quadratic regression curve, where the fit is reported as the coefficient of determination of the regression model (R^2). Correlations, regression and Loess smoothers were done using SigmaPlot v12.

To account for the effect of sediment focusing, the OC AR was corrected by the ratio of the estimated atmospheric ^{210}Pb flux at the site to the measured mean ^{210}Pb flux in the sediment core (see Lamborg and others 2013 for a detailed discussion of this approach) and is referred to as OC AR$_{fc}$. Assuming a general relationship of ~ 100 Beq m^{-2}/1000 mm y^{-1} of precipitation and an annual total precipitation gradient of 150–500 mm y^{-1}, estimated atmospheric ^{210}Pb fluxes in the Kangerlussuaq area are ~ 15 to 50 Beq m^{-2} y^{-1}, and the mean estimated from the cores analyzed in this study was 53 Beq m^{-2} y^{-1} (Table 2).

Biogenic silica (BSi) concentration was determined on freeze-dried sediments by extraction using a wet alkaline digestion employing 1% Na$_2$CO$_3$ with a correction applied for simultaneous mineral dissolution (Conley and Schelske 2001). BSi is used here as an approximate measure of primary production in lakes.

^{210}Pb Dating and Core Chronologies

Sediment cores were analyzed for ^{210}Pb, ^{226}Ra, ^{137}Cs and ^{241}Am by direct gamma assay using Ortec HPGe GWL series well-type coaxial low background intrinsic germanium detectors (Appleby and others 1986). ^{210}Pb was determined via its gamma emissions at 46.5 keV, and ^{226}Ra by the 295 keV and 352 keV γ-rays emitted by its daughter isotope ^{214}Pb following 3-week storage in sealed containers to allow radioactive equilibration. ^{137}Cs and ^{241}Am were measured by their emissions at 662 keV and 59.5 keV, respectively. The absolute efficiencies of the detectors were determined using calibrated sources and sediment samples of known activity. Corrections were made for the effect of self-absorption of low-energy γ-rays within the sample (Appleby and others 1986). ^{210}Pb results from some sites were reported elsewhere as part of paleolimnological studies addressing atmospheric pollution (Pb, Hg, POPs) and climate change in SW Greenland (Bindler and others 2001a; Bindler and others 2001b).

Radiometric dates were calculated using the CRS and CIC ^{210}Pb dating models (Appleby and Oldfield 1978), and the 1963 depth was determined from the ^{137}Cs and ^{241}Am stratigraphies where possible. Where there were discrepancies between these three different dating methods, a best chronology was calculated (following an assessment of all the data) using the procedures described in Appleby (2001).

Results

^{210}Pb and ^{137}Cs Profiles and Radiometric Inventories

Table 2 summarizes radiometric parameters from the lakes. ^{210}Pb/^{226}Ra equilibrium is reached at depths of between 3 and 8 cm at all but two sites,

Table 2. ^{210}Pb and ^{137}Cs Inventories for Cores From the Study Lakes

| Lake | Unsupported ^{210}Pb | | | | | | Weapons ^{137}Cs | |
| | Maximum Activity | | Inventory | | Flux | | Inventory | |
	Bq kg^{-1}	$\pm$	Bq m^{-2}	$\pm$	Bq m^{-2}y^{-1}	$\pm$	Bq kg^{-1}	$\pm$
SS32	160	9	871	42	27	2	443	14
SS86	161	17	164	13	5	1	51	2
SS901	326	17	1424	63	44	2	290	9
SS904	592	27	1619	47	50	2	290	8
SS16	470	35	878	39	27	2	149	7
SS1341	466	27	1137	51	36	2	165	6
SS2	219	32	481	77	15	3	408	20
SS4	235	16	782	39	24	2	254	9
SS6	114	10	538	35	17	2	302	9
SS70	254	28	881	75	27	3	504	21
SS49	809	66	1132	51	35	2	378	11
SS53	783	35	1969	71	61	3	539	17
AT1	2336	90	6995	172	218	5	832	23
AT6	1382	85	5221	170	162	5	715	24
AT7	2090	102	2522	78	78	2	387	13
Mean of all sites			1722		53.4		387	

SS16 and SS49. The greater depths at SS16 and SS49 (12–15 cm) are attributable to the low dry bulk density of the profundal sediments in these two lakes. Equilibrium depths measured in terms of cumulative dry mass at these two sites are, in fact, among the lowest among the sites. ^{210}Pb fluxes range from 5 to 218 Bq m^{-2} y^{-1}, with a mean value of 53 Bq m^{-2} y^{-1}. There is a trend of lower fluxes near the ice sheet and higher fluxes toward the coast, reflecting the regional precipitation gradient. ^{137}Cs inventories followed a similar pattern. Bulk sedimentation rates appear to have been relatively constant, apart from occasional brief episodes of more rapid accumulation at some sites (for example, SS2, SS70, SS16) due to disturbance events (for example, in-wash). All but two of the sites had a relatively well-resolved subsurface peak in ^{137}Cs concentration (at depths ranging from 1.25 to 5.75 cm), which records the 1963 fallout maximum from the atmospheric testing of nuclear weapons. This feature was used to correct the ^{210}Pb dates at those sites (SS4, SS70, SS49, SS86) where there was a significant discrepancy between the ^{137}Cs date and the ^{210}Pb date. At sites where there were differences between the ^{210}Pb models, the CRS model was generally considered to be more reliable.

The problems of dating arctic lake sediments are well known and are associated with the low total inventory and very low mean sedimentation rates (Wolfe and others 2004). The ^{210}Pb profiles measured in this study include sites with a consistent decline in activity with depth (for example, at SS16), and clearly resolved ^{241}Am and ^{137}Cs profiles, to very short, and less well-resolved profiles that are more typical of high arctic locations (Hermanson 1990). Low maximum ^{210}Pb concentration means that it is often possible to derive reliable chronologies only for the last $90 - 120$ years (not ~ 150 years as often cited). Moreover, the low activity means that the errors, particularly for older sediments, can be very high. At SS86, for example, the age estimate for the lowest sample with a ^{210}Pb date (5.25 cm) is 122 ± 21 years.

Lake C Accumulation Rates and Mean Regional C Burial Estimate

Carbon accumulation rates derived from individual lakes and corrected for sediment focusing range from 0.2 to 6.5 g C m^{-2} y^{-1} prior to 1950 increasing to 1.9–15 g C m^{-2} y^{-1} post-1950 across the three regions (Figure 2). The pre- and post-1950 regional median C burial rates of all lakes are significantly different ($p < 0.001$; Mann–Whitney)

(Figure 3A). Most (12 out of 16) lakes show an increase in C accumulation rates during the twentieth century (increase ratios range from 1.2 to 1.86) (Figure 3B).

Relationships between post-1950 mean OC AR$_{fc}$ and a number of limnological and morphometric variables are shown in Figure 4. Post-1950 mean C AR$_{fc}$ is correlated with lake conductivity ($r = 0.56$, $p < 0.05$), but the relationship to DOC concentration is not significant ($r = 0.45$, $p > 0.05$). For lake water nutrients, the only significant correlation was with TN ($r = 0.6$; $p < 0.05$). There is a weak negative relationship with maximum water depth ($r = -0.42$, ns), but a stronger, negative correlation with altitude ($r = -0.52$, $p < 0.05$) (Figure 4). Site location (as longitude) explains a significant component of the variability of OC AR$_{fc}$ along the ice margin-coast transect as determined by a quadratic relationship ($R^2 = 0.67$, $p < 0.005$) (Figure 4).

There are $\sim 20,000$ lakes in the area covered by Figure 1; the mean lake area is ~ 16 ha, and the largest lake is 139 km^2. The distribution of lakes is strongly skewed to lakes smaller than 10 ha (Table 3). Using the mean post-1950 OC AR$_{fc}$ of all lakes (3.6 g C m^{-2} y^{-1}, Table 3), the total regional annual C burial rate for lakes smaller than 1000 ha was estimated to be 8.5×10^9 g C y^{-1} (Table 3). Including those lakes with areas greater than 1000 ha (and applying the estimated regional rate), the total annual burial rate is 11.5×10^9 g C y^{-1}. However, using a lower OC AR for these large lakes of 2 g C m^{-2} y^{-1}, the estimated C burial rate was 10.14×10^9 g C y^{-1} (see Discussion).

DISCUSSION

Lakes are a key feature of arctic landscapes and represent an important component of regional C cycling (Kling and others 1991; Anderson and others 2009). Although arctic lakes are unproductive compared with lakes at lower latitudes, they may be an important C sink at the landscape scale (Anderson and others 2009), if only because of their density on the landscape, similar to the pattern seen in boreal regions (Prairie 2008). There are two problems in estimating C burial rates in arctic lakes at the landscape scale. First, given the high total number of lakes and their relative inaccessibility, relatively few lakes have been sampled. As a result, the validity of upscaling in these exercises is unclear. Second, this issue is made more problematical by the spatial heterogeneity of sediment deposition within the lakes themselves (Fitzgerald and others 2005; Rippey and others 2008; En-
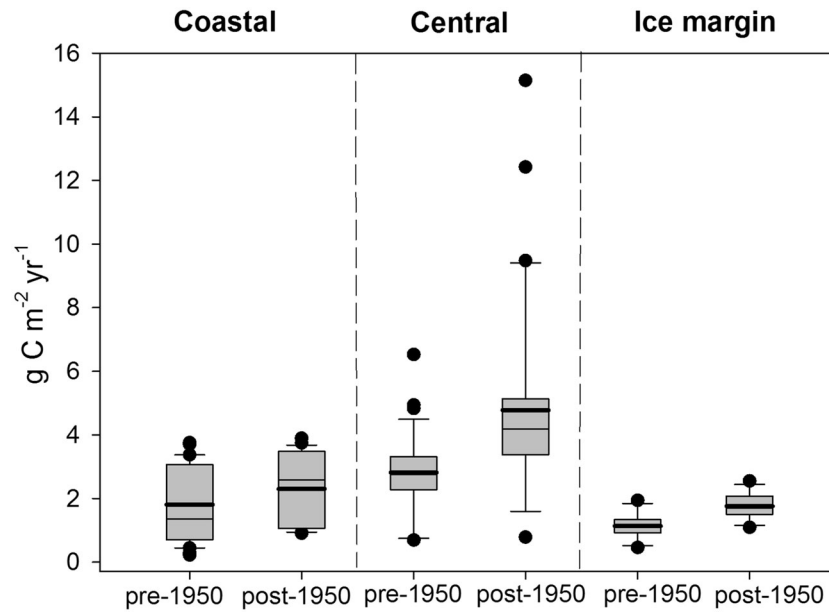
Figure 2. Box plots of pre- and post-1950 mean OC AR$_{fc}$ (g C m^{-2} y^{-1}) for the three local areas along the fjord. The horizontal thick line within the box is the mean value, and the thinner the median value
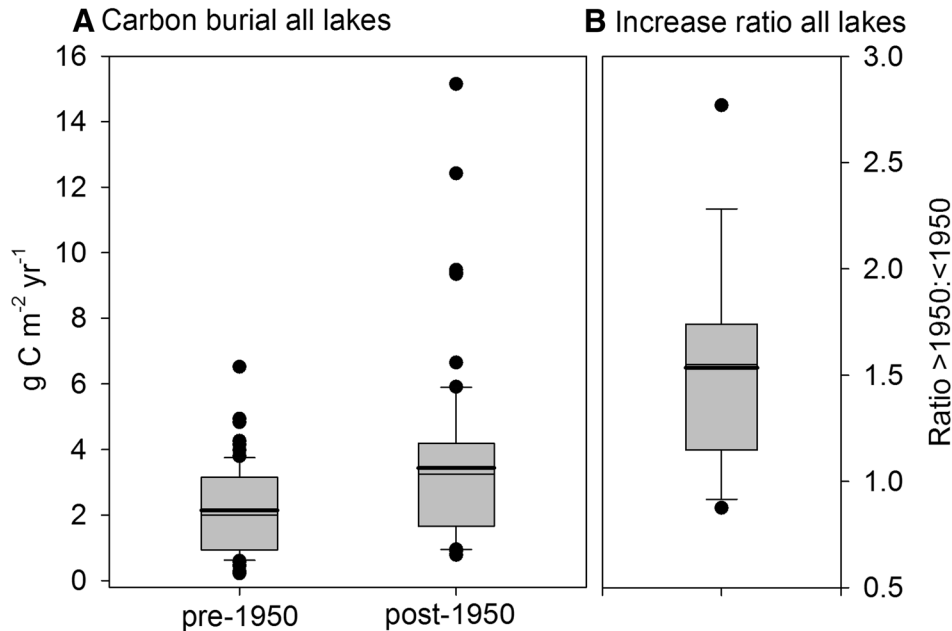

Figure 3. **A** Box plot of pre- and post-1950 OC AR$_{fc}$ (g C m^{-2} y^{-1}) for all lakes. **B** Box plot of the increase ratio (OC AR post-1950/pre-1950) for all lakes

gstrom and Rose 2013). The relatively few studies of C burial in the Arctic are generally based on single cores (Anderson and others 2009). Single, deepwater cores may over-estimate whole-lake sediment (and C) burial rates (Rippey and others 2008). Realistically, this problem can either be re-solved by use of a multiple core approach or the use of the ^{210}Pb focusing correction method (Anderson and others 2013). These corrected values can then be compared across lakes.

In-lake Variability of Organic Matter (OC) Burial

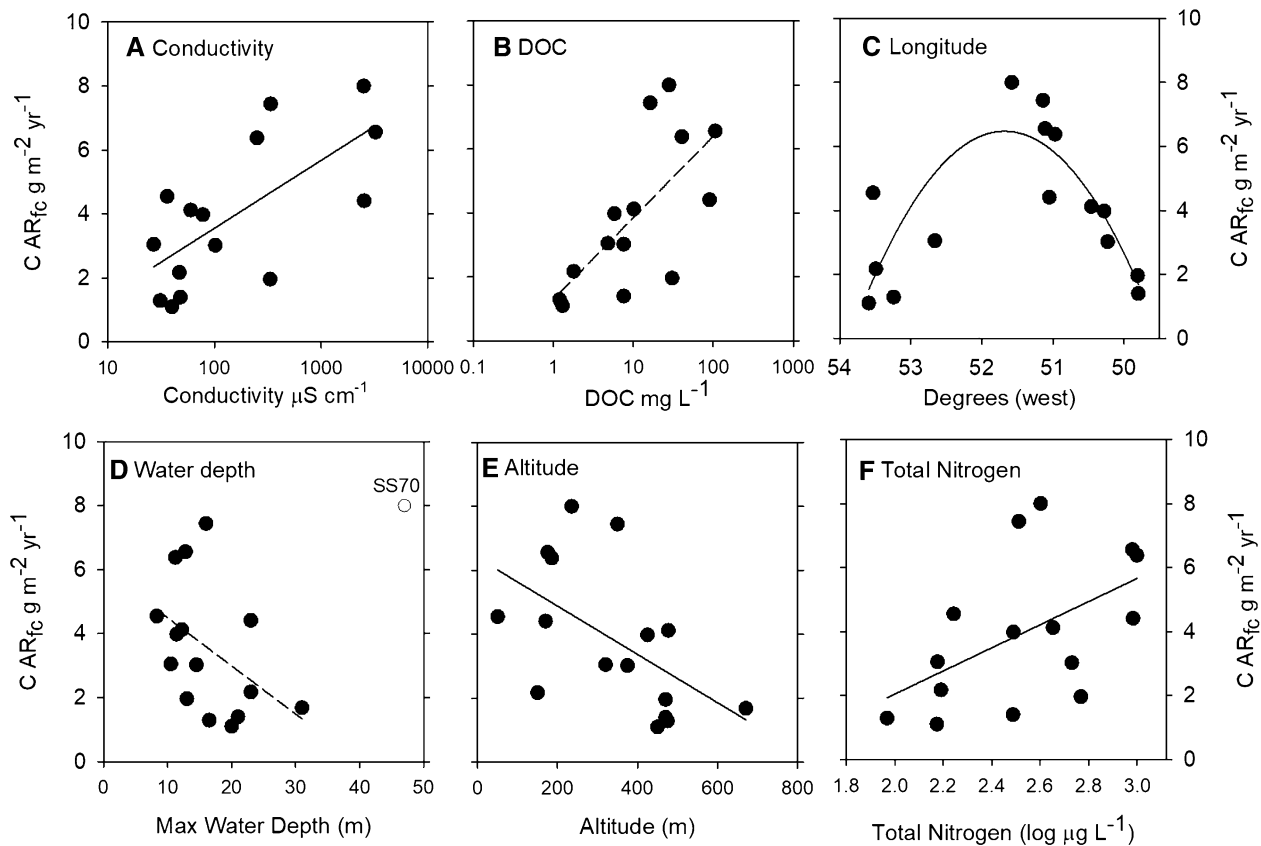Sediment focusing and changes in lake sedimen-tation patterns through time can affect estimates of

Figure 4. Relationships between the post-1950 mean OC AR_{fc} (g C m^{-2} y^{-1}) at individual lakes and selected water chemistry and morphometric/landscape variables. **A** Conductivity, $r = 0.56$, $p < 0.05$; **B** dissolved organic carbon (DOC), $r = 0.45$, ns; **C** longitude, $R^2 = 0.67$, $p < 0.001$; **D** maximum water depth, $r = -0.42$, ns (note the outlier, SS70); **E** altitude, $r = -0.52$, $p < 0.05$; **F** total nitrogen, $r = 0.60$, $p < 0.05$. Lines of best fit are indicated for guidance only; dashed lines indicate a nonsignificant relationship

Table 3. Total Number of Lakes in the Study Area (Figure 1) for Given Lake Area Ranges, Cumulative Lake Areas and Estimated Regional Annual Organic C Burial Rate

Lake area (ha)	Number of lakes	Total lake area (ha)	Total regional annual C burial (g C yr^{-1} × 10^9) for each lake area interval	Cumulative annual C burial (g C yr^{-1} × 10^9) lakes up to 1000 ha
<10	16,379	36,437	1.31	1.31
10–50	2565	54,158	1.95	3.26
50–100	375	25,980	0.94	4.2
100–1000	470	118,931	4.28	8.48
>1000	24	83,001	1.66 (2.98)	10.14
Total	19,813	318,506		10.14 (11.46[#])

A mean C accumulation rate of 3.6 g C m^{-2} y^{-1} was used for the time period post-1950 and 2 g C m^{-2} y^{-1} for lakes > 1000 ha.
[#]*The annual burial rate based on the application of a mean C accumulation rate of 3.6 g C m^{-2} y^{-1} to lakes > 1000 ha is given in parentheses and is discussed in the text.*

lake C budgets (Cole and others 2007). Basing estimates on multiple cores within each basin can control for these temporal changes (Kortelainen and others 2004; Ferland and others 2012), but this approach is time-consuming. However, the mean ^{210}Pb flux in a sediment core can be used to correct for sediment focusing, although it presupposes that the regional atmospheric flux is well constrained

(Lamborg and others 2013). ^{210}Pb fluxes are generally low in the Arctic (Hermanson 1990) and have not been measured for the Kangerlussuaq region specifically but are well constrained for the Greenland ice sheet. The approach used here, estimating the ^{210}Pb flux from precipitation, is, however, better than propagating uncorrected C burial rates.

^{210}Pb inventories commonly are lower in the Arctic than at lower latitudes, which reflects the generally low precipitation rates in polar regions and longer periods of ice cover (Wolfe and others 2004). As in temperate areas, in SW Greenland the pattern of ^{210}Pb distribution reflects the regional precipitation gradient, with higher inventories at the coast (for example, AT1, AT6: $\sim$ 5000 Bq m^{-2}) compared with the drier inland areas (for example, SS2: < 500 Bq m^{-2}) (Table 2). The contrasting inventories of the two adjacent lakes on the nunatak (SS32 and SS86) may reflect the different catchment/lake ratios and mean volumes (Table 2).

Assuming that the atmospheric flux at the head of the fjord is $\sim$ 15 Beq m^{-2} y^{-1}, focusing by a factor of two is suggested for lakes SS4, SS32 and SS16. Using the ^{210}Pb focusing factor at SS16 gives a mean basin OC AR similar to that derived from a multicore study at this site (Anderson unpublished), indicating that an uncorrected OC AR derived from a single central core is double the basin mean. Overall, however, the ^{210}Pb data suggest that focusing is not a major problem (the average correction factor was $\sim$ 2, range < 1 to > 5) when the fluxes are compared with that expected from atmospheric inputs.

Sediment focusing may be less important in this climate setting relative to boreal and temperate lakes, for several reasons. First, arctic lakes may be less prone to wind-driven resuspension and sediment redistribution because of the long periods of ice cover. In addition, in much of the Arctic (including SW Greenland), thermal stratification occurs very rapidly after ice melt, as ice off coincides with the timing of maximal solar radiation input. Thus, water column mixing is limited. An unknown factor, however, is ice scouring by drifting ice immediately prior to ice melt (Hermanson 1990). Given ice thicknesses of 1.5 m, scouring would limit long-term sedimentation in water depths below 2 m in larger lakes; it may be less of a problem in small lakes because of the reduced fetch. The possible importance of ice drift in larger lakes is indicated by observations of ice push ramparts.

Comparative OC Burial Rates

Focusing-corrected organic C burial rates in SW Greenland lakes during the last 100 years range from $\sim$ 1 to greater than 10 g C m^{-2} y^{-1}. Anderson and others (2009) estimated the mean Holocene burial rate (using ^{14}C chronologies and not focusing corrected) for the Kangerlussuaq area as 5.5 g C m^{-2} y^{-1}, higher than that calculated in the present study for the period 1950–2000 and corrected for focusing using ^{210}Pb fluxes (Figure 2), suggesting that the Holocene rates are slight overestimates. Burial rates in high arctic lakes are probably closer to the low end measured here (1–2 g C m^{-2} y^{-1}). Lundin and others (2015) reported rates of 5.3 to 24.6 g C m^{-2} y^{-1} (n = 6) for arctic Sweden although these rates are not focusing corrected. Stallard's (1998) estimate of the global average (4.5 g C m^{-2} y^{-1}) is similar to that for Finland (Kortelainen and others 2004) and for the present study of SW Greenland lakes (Table 3). The rates calculated for Greenland lakes are, however, low compared to those estimated for the boreal zone (15 $\pm$ 9.4 g C m^{-2} y^{-1}) (Heathcote and others 2015) and culturally impacted lakes ($\sim$ 60 g C m^{-2} y^{-1}) (Anderson and others 2013), but are close to the minimum reported as the global average by Tranvik and others (2009), although the latter rate was not focusing corrected and was derived from a variety of methods. The observed range in the present study ($\sim$ 1–10 g C m^{-2} y^{-1}) is similar to the whole-basin OC AR estimated for lakes in the Toolik area of Alaska (6 to 12 g C m^{-2} y^{-1}) (Anderson and others unpublished). At Kassjön, a small boreal forest lake in Sweden with some agricultural activity in the catchment, Rippey and others (Rippey and others 2008) estimated a mean OC AR of $\sim$ 10 g C m^{-2} y^{-1} over the period 1900–1988, based on 43 cores.

Landscape and Water Chemistry Controls on OC Burial

As well as lake morphometry (that is, sediment focusing) and mineralization processes, there are also landscape-scale factors that influence carbon burial at individual sites (Buffam and others 2011) and mean that extrapolating from individual lakes to regional means should be done with care. The weak, positive relationship between OC AR$_{fc}$ and DOC (r = 0.45; Figure 4) is not unexpected given the regional range in DOC concentration (Table 1), and likewise the relationship with conductivity because DOC and conductivity are themselves highly correlated (see Anderson and Stedmon

2007). There is also a weak relationship between OC AR_{fc} and TN, but this reflects the high organic N content of DOM in the central lakes.

The relationship between OC AR_{fc} and DOC suggests that the sediment OM is partly derived from the in-lake DOC pool, which is assumed to be refractory terrestrial DOC (Anderson and Stedmon 2007). However, the role of seasonal DOM inputs from macrophytes and its degradation may also be an important mechanism (Osburn and others 2017). The slight regional decline in OC AR (Figure 5) over the last ~ 10 years is unlikely to be an artifact of post-deposition mineralization as the opposite effect would be expected, that is, increasing OC AR in the most recent sediments (Sobek and others 2009; Sobek and others 2014).

The inverse relationship between OC AR and water depth observed for the Kangerlussuaq area has been observed elsewhere and reflects the greater time the particles spend sinking through the water column and are exposed to bacterial respiration, but large, deep lakes also tend to be less productive (Meyers and Lallier-Verges 1999).
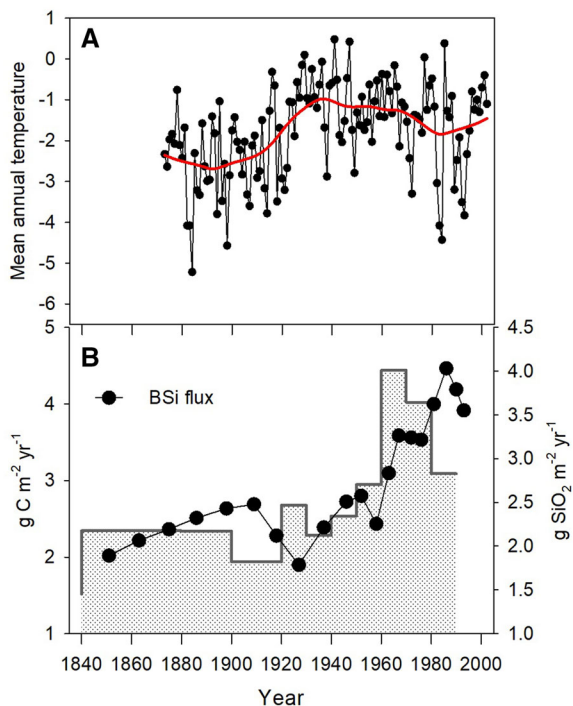

Figure 5. Summary figure illustrating the cooling trend in SW Greenland (regional air temperature and two indicators of increasing lake production). **A** Regional climate trends—long-term MAT temperature record from Nuuk (www.dmi.dk); **B** BSi accumulation rate (g SiO_2 m^{-2} y^{-1}) at SS16, an independent measure of lake production and the regional OC AR_{fc} (g C m^{-2} y^{-1}), presented as period means (range 20–10 years)

Anderson and others (2014) observed a significant relationship between OC AR_{fc} and water depth in European lakes, although the lake depth gradient in that dataset was greater than that observed in this study.

The inverse relationship between OC AR_{fc} and altitude implies some climatic control, but catchments at higher altitude tend to have reduced vegetation and little, if any, soil cover, thereby reducing biogeochemical cycling and hence nutrient transfer from catchments to the lakes (Wookey and others 2009). Further indication of a regional climate control is suggested by the higher OC AR_{fc} in the central lakes (Figures 2, 4E) which are significantly warmer than the coast and ice margin lakes, but which also tend to have higher nutrient concentrations (Whiteford and others 2016). However, this inference has to be offset against the fact that the post-1950 increases in OC AR_{fc} (Figure 5) occurred while the area was cooling (Box 2002; Mernild and others 2014). Moreover, although there are significant differences in regional lake temperatures, there are also strong regional limnological gradients (Anderson and others 2001; Whiteford and others 2016).

Temporal Variability of OC AR

Long-term carbon burial in lakes is a reflection of three main processes: autochthonous production, inputs of allochthonous C from the catchments and loss rates (mineralization) in both the water column and sediments (Cole and others 2007). The Greenland data suggest that mean OC AR_{fc} increased from less than 2 g C m^{-2} y^{-1} prior to 1950 to more than 3.6 g C m^{-2} y^{-1} post-1950 (Figures 2, 3). This trend therefore suggests a recent production increase, as has been observed elsewhere in the Arctic (Michelutti and others 2005; Catalan and others 2013). If so, processes other than warming are responsible for the regional production increase in these lakes, as mean growing season temperatures decreased over this period in SW Greenland (Box 2002) (Figure 5). A more parsimonious interpretation may be that the difference merely reflects incomplete mineralization after incorporation in the sediments (Sobek and others 2009) (see above). It may take more than 50 years before the C content of aquatic sediments stabilizes, but most mineralization occurs within 10 years of deposition (Galman and others 2008). Sobek and others (2014) observed sediment CH_4 in a number of lakes in the study area; the CH_4 production rates were very low and are indicative of slow decomposition over decades. Countering the

conclusion that C burial rates are primarily the function of mineralization processes is the range of limnology, water depths and morphometry covered by the sampled lakes (four of the lakes have anoxic hypolimnia): It is difficult to expect the post-deposition mineralization rates to be similar across these lakes. Moreover, some lakes show declines in OC AR over the most recent time periods, suggesting that the profiles are not merely artifacts of mineralization processes.

There are very limited estimates of primary production in the study lakes, but given the low nutrient concentrations and short growing season, primary productivity will be low (Lizotte 2008). In the smaller (< 20 ha) and shallower (< 20 m) lakes, two factors may contribute to enhanced preservation of organic C (relative to temperate systems): anoxia and lower mean annual water temperatures (Sand-Jensen and others 1999). Mineralization rates are lower in anoxic conditions, and although anoxia is a dominant feature of the oligosaline and meromictic lakes at the head of the fjord, it also occurs in freshwater, oligotrophic lakes as a result of morphometric controls on thermal stratification and under ice in the winter (Whiteford and others 2016). For example, the OC AR at the surface of SS16 (> 10 g C m^{-2} y^{-1}) may reflect locally enhanced preservation due to the recent development of hypolimnetic anoxia. How long the lake has had seasonal anoxia is not known, but the presence of clear laminations in the surface sediment of deepwater cores (Anderson unpublished field observations) suggests it is a relatively recent phenomenon, perhaps reflecting altered stratification patterns, despite the late twentieth century cooling in this area (Box 2002). A more general process that is applicable to all lakes, however, may be the low hypolimnetic temperatures that occur in high-latitude lakes (Sand-Jensen and others 1999), but this conclusion is not supported by the work of Sobek and others (2009). Moreover, the decline in the regional mean OC AR during the late twentieth century (Figure 5) is the opposite trend expected if mineralization was a dominant factor. It is unlikely, therefore, that the increase in OC AR is simply a product of a recent reduction in diagenesis.

There is substantial evidence for recent ecological change and production increases in arctic lakes (Catalan and others 2013), including SW Greenland (Perren and others 2009). An independent measure of lake production is provided by BSi accumulation rates which are not subject to the same diagenetic controls as organic matter. Diatoms and chrysophytes contribute substantially to whole-lake (that is, both benthic and planktonic) production in oligotrophic arctic lakes. At SS16, the BSi AR increases steadily from ∼ 1940 to 1990, paralleling the regional OC AR increase (Figure 5), suggesting that there is a production increase in these lakes. The question remains, however, what is driving this increase?

Regional warming has been implicated in many parts of the Arctic (Michelutti and others 2005), but SW Greenland exhibits no warming trend from 1950 as elsewhere in the Arctic (Figure 5) (Box 2002). Given the limited hydrologic runoff in this area, changing OC AR is unlikely to be the result of transfer of terrestrial carbon, although this cannot be excluded at the coast where there is greater rainfall (see Leng and others (2012) for a discussion). One explanation for the decline in OC AR may be the decreased DOC concentration which has been observed in a number of lakes in the Kangerlussuaq over the last ∼ 12 years (Saros and others 2015). A direct causal link between the DOC decline and declining OC AR assumes that the C burial rate is largely controlled by flocculation of the refractory DOC pool and that the decline in the DOC pool does not only reflect photochemical processing or mineralization in the water column (Cory and others 2014). Whereas the role of flocculation of allochthonous DOC has been demonstrated for boreal lakes with a high load of terrestrial C (von Wachenfeldt and Tranvik 2008), the processes related to DOC settling/deposition in the Kangerlussuaq lakes are less clear.

Much of the Arctic is nutrient limited, and production increases in oligotrophic arctic lakes require a nutrient source (Kling and others 1991). Internal recycling can contribute nutrients, and there is evidence for increased regional NO_3 deposition from the Greenland ice core (Hastings and others 2009). Although N deposition rates are low (< 1 kg ha y^{-1}) in the study area (Curtis and others 2018), sustained chronic input over decadal timescales can have marked effects on nutrient-poor ecosystems (Elser and others 2009). Bergstrom and others (2005) have shown a pronounced effect of N deposition in oligotrophic boreal and alpine lakes. An agreement between C burial rates and NO_3 deposition over the last 150 years has been observed in boreal lakes (Heathcote and others 2015).

Chronic reactive N deposition will lead to P limitation in nutrient-depleted systems (Elser and others 2009). The lakes used in this study are strongly P-limited in the spring following the N pulse associated with snow melt (Tye and others 2005; Curtis and others 2018), but then are co-

limited (N and P) during the summer (Whiteford unpublished). Long-term N deposition can alter lake and catchment stoichiometry and stimulate loss of alkalinity (Catalan and others 2013).

For the ice margin and central lakes, an alternative explanation is increased phosphorus inputs associated with regional dust deposition (Anderson and others 2018). Camarero and Catalan (2012) have shown that atmospheric dust inputs enriched in phosphorus can alleviate nutrient limitation in remote lakes. In SW Greenland, there is widespread dust deposition associated with deflation of fine particulates from glacio-fluvial outwash plains (Dijkmans and Törnqvist 1991) derived from the Greenland ice sheet which is a major local source of phosphorus (Hawkings and others 2016). It is possible that this alternative, natural atmospheric nutrient source, when coupled with long-range reactive N deposition associated with industrialization (Galloway and others 2008; Wolfe and others 2013) is sufficient to drive the observed aquatic production increase (Figure 5).

Regional Upscaling

The regional estimate of annual C burial in SW Greenland (Table 3) is based on total lake area, that is, assumes burial over 100% of any lake basin. This is unlikely to be the case in larger lakes, where stony, boulder-strewn littoral zones are common. However, small lakes dominate the regional lake distribution (Table 3), and in these systems littoral zones with substantial accumulation of organic sediments are common. The negative relationship between C burial and lake depth (Figure 4) is well known (Kortelainen and others 2004); C burial in large (> 1000 ha), deep lakes reflects loss rates in the water column and probably has greater affinity to processes operating in the oceans. For these reasons, it is inappropriate to extrapolate the mean estimated C burial rate from small lakes in the Kangerlussuaq area (that is, ~ 3.6 g C m^{-2} y^{-1}) to the large, deep fjord-like lakes (Figure 1; Table 3). Kortelainen and others (2004) indicated the importance of small lakes for C burial and the relatively low C accumulation rate in large lakes in Finland. In the present study area, there are only a few large lakes (> 1000 ha) (Figure 1), but they represent $\sim 26\%$ of total lake area. Because of the lack of data on either the C content of the sediments or bulk sediment accumulation rates in these large lake systems, we applied the lowest post-1950 mean rate observed in this study (~ 2 g C m^{-2} y^{-1}) to the largest lakes (> 1000 ha) to obtain a complete regional mean C burial rate estimate (Table 3). An alternative to extrapolating the mean of ^{210}Pb-focusing-corrected rates from lakes is to undertake detailed multiple coring (to estimate whole-basin fluxes) and side-scan sonar surveys to delimit sedimentation in lakes (Kortelainen and others 2004; Fitzgerald and others 2005), an approach that is difficult to apply to many lakes at the regional scale in remote areas such as SW Greenland.

The question remains, however, how applicable are the rates in this study to lakes elsewhere on Greenland? Importantly, the Kangerlussuaq climatic–limnological gradient probably encapsulates the full range of lake types and water chemistries occurring today on Greenland. The two main areas of limnological research on Greenland are Zackenberg (NE Greenland) and around Nuuk, some 400 km south of Kangerlussuaq, both sites of detailed ongoing annual monitoring of a small subset of lakes in each area. There are, however, no ^{210}Pb-dated sediment carbon burial rates available for these areas. In both Zackenberg and Nuuk, the lakes are dilute (conductivity < 50 µS cm^{-1}) and nutrient poor (Topp-Jørgensen and others 2017). The Nuuk lakes are similar to the coastal lakes in the present study (around Sisimiut) in terms of chemistry, that is, they are typical arctic lakes, and will probably have burial rates ~ 2 g C m^{-2} y^{-1}. Arguably, for the ice-free area of west Greenland (sensu lato), with many thousands of lakes, effort could be put into obtaining cores from more lakes (including the larger systems) to better constrain the regional rate. In the Kangerlussuaq area, sediment cores have been taken from ~ 120 lakes in this area (0.6% of the total) and only less than 0.1% of the total have been dated using ^{210}Pb. Nonetheless, we posit that our estimates from individual cores used for regional upscaling are relatively robust (Table 3).

REGIONAL SIGNIFICANCE OF AQUATIC C BURIAL

The Arctic represents a key component of the global terrestrial C budget, because tundra soils contain more than 50% of C stored in soils globally (Tarnocai and others 2009). This terrestrial carbon has considerable implications for the climate system if it is released into the atmosphere as the Arctic warms (Schuur and others 2008). The soils in the Kangerlussuaq area are, however, relatively thin, have OC content of ~ 8–16% and are absent on many south-facing slopes because moisture stress restricts vegetation development; vegetation

cover is limited to $\sim 50\%$ of the land area (Anderson and others 2009). Thus, in the Kangerlussuaq area, lakes are a significant component of the regional C cycle. Applying the mean post-1950 OC AR_{fc} (3.6 g C m^{-2} y^{-1}) to all lakes across the region, regardless of lake type and morphometry (Table 3), the amount of C sequestered annually by lakes in SW Greenland today is $\sim 11.5 \times 10^9$ g C y^{-1} (using 2 g C m^{-2} y^{-1} for the largest lakes, this estimate is $\sim 10 \times 10^9$ g C y^{-1}) (Table 3).

This study provides an estimate of contemporary regional OC burial by lakes in SW Greenland and illustrates the need to integrate lakes into estimates of regional OC AR (Cole and others 2007). Regional lake density in this study is relatively high (14% of the land area is water) and is typical of West Greenland from $\sim 60°N$ to $\sim 70°N$: In this area, lakes are undoubtedly an important part of the regional C cycle (Anderson and others 2017). Many boreal and arctic lakes have high CO_2 emission rates because of in-lake mineralization of terrestrial C (Kling and others 1991; Algesten and others 2004), a process that will change with regional warming. Most net heterotrophic systems studied to date are boreal lakes with organic-rich catchments: peatlands and/or organic forest soils, which are not characteristic of the Kangerlussuaq area. Anderson and Stedmon (2007) argued that terrestrial inputs of DOC are limited for the inland lakes in the Kangerlussuaq area, and preliminary $p CO_2$ data (C. Barry & NJ Anderson unpublished) suggest that lakes in the Kangerlussuaq area of SW Greenland are net autotrophic (despite the high DOC concentration) and so the lakes are essentially islands of C sequestration at the landscape scale as has been observed elsewhere (Tank and others 2009).

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