Heterogenous CO$_2$ and CH$_4$ content of glacial meltwater of the Greenland Ice Sheet and implications for subglacial carbon processes

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Abstract. Accelerated melting of the Greenland Ice Sheet (GrIS) has increased freshwater delivery to the Arctic Ocean and amplified the need to understand the impact of GrIS meltwater on Arctic greenhouse gas (GHG) budgets. We measured carbon dioxide (CO$_2$) and methane (CH$_4$) concentrations and $\delta^{13}$C values and use geochemical models to evaluate subglacial CH$_4$ and CO$_2$ sources and sinks in water discharging from three subglacial outlets of the GrIS in southwest (Isunnguata and Russell Glaciers) and southern Greenland (Kiatutt Sermiat). CH$_4$ concentrations vary by orders of magnitude between sites and are saturated with respect to atmospheric concentrations at Kiatutt Sermiat, but are supersaturated at southwest sites, even though oxidation reduces concentrations by up to 50% during periods of low discharge. CO$_2$ concentrations range from supersaturated at Isunnguata to undersaturated at Kiatutt Sermiat. CO$_2$ is consumed by mineral weathering throughout the melt season at all sites, however differences in the magnitude of subglacial CO$_2$ sources result in meltwaters that are either sources or sinks of atmospheric CO$_2$. The predominant source of CO$_2$ at Isunnguata is organic matter (OM) remineralization, but Russell and Kiatutt Sermiat sites have multiple or heterogeneous subglacial CO$_2$ sources that maintain atmospheric CO$_2$ concentrations at Russell but not at Kiatutt Sermiat where CO$_2$ is undersaturated. These results highlight the variability in GHG dynamics under the GrIS. Constraining this variability will improve our understanding of the impact of GrIS melt on Arctic GHG budgets, as well as the role of continental ice sheets in GHG variations over glacial-interglacial timescales.

1 Introduction

Glaciers play an important role in global chemical cycles due to the production of fine-grained sediments that participate in carbonate and silicate mineral weathering reactions (Table 1), which are the principal sink of atmospheric CO$_2$ over geologic timescales (Berner et al., 1983; Walker et al., 1981). Variations in the weathering intensity of comminuted sediments may contribute to glacial-interglacial atmospheric CO$_2$ variations as sediments are alternately covered by ice and exposed following ice retreat. However, the importance of CO$_2$ consumption by mineral weathering is poorly understood, including effects from the advance and retreat of continental ice sheets (Ludwig et al., 1999). Recent evaluations of carbon budgets in proglacial environments indicate that mineral weathering results in net sequestration of atmospheric CO$_2$, suggesting
that proglacial systems are underrecognized as Arctic CO$_2$ sinks (St. Pierre et al., 2019), however alternate processes could lead to the production of greenhouse gases (GHG) in glacial systems. For instance, CH$_4$ production in anaerobic subglacial environments driven by the remineralization of organic matter (OM) contained in soils and forests covered during glacial margin fluctuations has been suggested as a potential carbon feedback to drive warming (Sharp et al., 1999; Wadham et al., 2008). Because the global warming potential of CH$_4$ is 25 times greater than CO$_2$, even limited subglacial methanogenesis has the potential to strongly impact the GHG effect of glacial meltwater. Combined inorganic and organic subglacial processes may therefore produce glacial meltwater that is a source or sink of GHG. While the net impact of these processes on modern carbon fluxes is poorly constrained, determining these impacts will improve modern carbon budgets as well as depictions of how fluxes may have evolved during the advance and retreat of continental ice sheets.

In subglacial environments where remineralization is limited by low OM availability, the major element solute load of glacial meltwater is typically dominated by products of mineral weathering reactions (Tranter, 2005). The extent of mineral weathering in subglacial environments depends in part on the availability of acids to drive reactions, namely sulfuric and carbonic acids (Table 1). Sulfuric acid is derived from the oxidation of reduced sulfur species, which largely occur as iron-sulfide minerals including pyrite (Tranter, 2005). Sulfide oxidation may occur abiotically, however the kinetics of microbially mediated sulfide oxidation is several orders of magnitude faster and may lead to local depletion of oxygen provided sufficient supply of sulfide minerals (Sharp et al., 1999). In contrast, carbonic acid may be derived from multiple external or in situ sources of CO$_2$ to the system. The dominant external source is supraglacial meltwater that flows to the subglacial system through moulins following equilibration with atmospheric CO$_2$ (Fig. 1). Unlike proglacial environments where free exchange between water and the atmosphere may resupply CO$_2$ consumed by weathering, subglacial environments may be partially or fully isolated from the atmosphere, limiting further atmospheric CO$_2$ invasion and thus the extent of mineral weathering with carboxylic acid. However, additional atmospheric CO$_2$ may be delivered in open portions of the subglacial environment though exchange in fractures or moulins along subglacial flow paths or in partially air-filled conduits, allowing a much greater magnitude of carbonic acid mineral weathering (Graly et al., 2017). CO$_2$ may also be derived from in situ sources, such as gaseous CO$_2$ contained in ice bubbles of basal ice, or fluid inclusions in rocks that release volatiles (including CO$_2$) following mechanical grinding (Macdonald et al., 2018). When OM is available, its remineralization also generates CO$_2$ (and potentially CH$_4$) along with nutrients, but low OM availability in many subglacial systems limits remineralization as a CO$_2$ source (Fig. 1).

The role of subglacial carbon processes may play an increasingly important role in modern Arctic carbon budgets as disproportionate warming increases glacial meltwater and sediment fluxes to the ocean, particularly from the Greenland Ice Sheet (GrIS). The Greenland Ice Sheet (GrIS) is the last remaining ice sheet in the Northern hemisphere following collapse of all others since the Last Glacial Maximum (~25 ka). It has been losing mass at increasing rates that averaged $286\pm20$ Gt/yr between 2010-2018, representing a six-fold increase since the 1980s (Mouginot et al., 2019). While mineral weathering significantly modifies the composition of GrIS subglacial discharge (e.g. Hindshaw et al., 2014; Deuerling et al., 2018; Urra
et al., 2019) and should consume CO$_2$ similar to other glacial and proglacial environments, the recent identification of microbially driven reactions (including methanogenesis) in subglacial environments of the GrIS indicates that organic processes may also play a role (Dieser et al., 2014; Lamarche-Gagnon et al., 2019; Musilova et al., 2017). The relative importance of subglacial GHG sinks (CO$_2$ consumption through mineral weathering) and sources (such as OM remineralization) determine the GHG composition of subglacial discharge, which may then serve as a source or a sink of atmospheric GHGs. Constraining the relative impacts and variability of these processes underneath the GrIS will provide important information regarding the current and future impact of GrIS loss on Arctic carbon budgets, as well the role of continental ice sheets on carbon cycle feedbacks.

To evaluate the net impact of carbon processes on the GHG composition of subglacial discharge of the GrIS, we compare water chemistry, dissolved CO$_2$ and CH$_4$ concentrations, and gas stable isotopic compositions between three subglacial discharge sites draining land-terminating glaciers of the GrIS over the melt seasons of 2017 and 2018 (Fig. 2). We employ mass balance models utilizing the concentrations of major cations and anions to determine the magnitude of the impact on CO$_2$ concentrations from mineral weathering reactions (Table 1). These results are combined with measured gas concentrations to determine the relative importance of mineral weathering compared to OM remineralization on the CH$_4$ and CO$_2$ content of subglacial discharge. We also assess the temporal and spatial variability of these processes under the GrIS to improve our understanding of carbon cycling in Greenland subglacial environments and the implications of GrIS mass loss on Arctic carbon budgets.

2 Methods

2.1 Study locations

Our three subglacial discharge locations are located in southwest (Fig. 2a, b) and southern (Fig. 2a, c) Greenland. The Isunnguata Glacier (IS; 67°09'27.1" N, 50°03'25.0" W) and Russell Glacier (RU; 67°05'22.1" N, 50°14'18.7" W) drain to the Watson River, which is one of the largest proglacial rivers in Greenland. Watson River discharge is monitored by PROMICE (Programme for Monitoring of the Greenland Ice Sheet; van As et al., 2018) and total discharge was 4.3 and 3.6 km$^3$ of water in 2017 and 2018, respectively (van As et al., 2018). The total catchment size for the Isunnguata is 15,900 km$^2$, though our water samples were collected from a smaller sub-catchment with a drainage area of approximately 40 km$^2$ (Lindbäck et al., 2015; Rennermalm et al., 2013). The total drainage area for the Russell glacier is estimated at 300 km$^2$. This estimate comes from subtracting the Leverett drainage area estimated at approximately 600 km$^2$ (Hawkings et al., 2016) from the approximately 900 km$^2$ catchment that includes the Russell and Leverett drainages (Lindback et al., 2015). While discharge from the third site in southern Greenland, Kiattut Sermiat (KS; 61°12'13.5" N, 45°19'49.1" W), is not monitored, a previous study using dye tracing techniques estimated approximately 0.22 km$^3$ of discharge in 2013, and its catchment size was estimated at 36 km$^2$ (Hawkings et al., 2016). Underlying lithologies differ between sites. Watson River sites are located near the boundary between the Archean Craton to the south and the southern Nagssugtoqidian Orogen to the north (Henriksen et
al., 2009). The Archean block is composed of granites and granulite facies orthogneisses that were intruded by mafic dykes during Paleoproterozoic rifting. These rocks were deformed and modified during subsequent continent-to-continent collision in the Paleoproterozoic to create the amphibolite facies gneisses of the southern Nagssugtoqidian Orogen (van Gool et al., 2002). Kiattut Sermiat lies within the Paleoproterozoic Ketilidian fold belt (Henriksen et al., 2009). Lithologies in this region include the Julianehåb Granite and associated basic intrusions and the sedimentary and volcanic rocks of the Mesoproterozoic Gardar Province that include a suite of alkaline igneous rocks and basaltic dykes with interbedded sandstones (Kalsbeek and Taylor, 1985; Upton et al., 2003).

Previous studies have characterized chemical weathering reactions in subglacial discharge to the Watson River (Deuerling et al., 2018; Hasholt et al., 2018; Yde et al., 2014), Kiattut Sermiat (Hawkins et al., 2016), and comparatively between these sites (Urra et al., 2019). There has been extensive work regarding ice sheet dynamics and hydrology in the Watson River catchment (Van As et al., 2017, 2018; Lindbäck et al., 2015) as well as Kiattut Sermiat (Warren and Glasser, 1992; Winsor et al., 2014). Previous studies near these study locations have identified CH$_4$ and CO$_2$ supersaturation in subglacial discharge of the Isunnguata site (Christiansen and Jørgensen, 2018; Ryu and Jacobson, 2012), while methanogenic microbial communities have been observed at Russell Glacier (Dieser et al., 2014) and CH$_4$ supersaturation at the Leverett Glacier (Lamarche-Gagnon et al., 2019), which also flows into the Watson River (Fig. 2c). Subglacial permafrost has been identified near the Isunnguata site (Ruskeeniemi et al., 2018) and attributed to Holocene fluctuations in the ice sheet margin. While a similar Holocene ice retreat and re-advance may have occurred in southern Greenland (Larsen et al., 2016), it is unknown whether this retreat led to the formation of organic deposits.

### 2.2 Sample collection

We collected water samples from subglacial discharge sites in spring and fall of 2017, and the summer of 2018 to observe seasonal variations in water composition. Samples were collected as close as possible to the subglacial discharge site, which was less than 10 m for the Isunnguata site, approximately 100 m for the Russell glacier site, and approximately 1.1 km for the Kiattut Sermiat site (Fig. 2). We collected water samples by pumping water through a 0.5-cm flexible PVC tube that was placed in flowing water as far as possible from shore (approximately 1-2 m). A YSI Pro-Plus sensor that was calibrated daily was installed in an overflow cup filled from the bottom to measure specific conductivity (Sp.C), temperature, pH, dissolved oxygen, and oxidation-reduction potential (ORP). These parameters were monitored until stable, between about 10 and 30 minutes, after which samples were collected and preserved in the field according to the solute to be measured after being filtered through a 0.45 µm trace-metal grade Geotech high capacity disposable canister filter. Samples for cations and anions were collected in HDPE bottles; cation samples were preserved with Optima-grade ultrapure nitric acid (pH<2) while no preservative was added to anion samples. Samples for ammonium (NH$_4$) were filtered into 15 mL polypropylene containers and frozen until analysis. Dissolved inorganic carbon (DIC) samples were filtered through 0.2 µm filters directly to the bottom of 20 ml Qorpac glass vials and allowed to overflow until sealed tightly with no headspace.
Gas samples were collected in duplicate via headspace extractions according to methods outlined in Repo et al. (2007) and Pain et al. (2019). Unfiltered water was pumped into the bottom of 500 mL bottles until they overflowed. Bottles were immediately capped with rubber stoppers fitted with two 3-way inlet valves. 60 mL of water was extracted from one inlet and replaced with 60 mL of atmospheric air (for spring and fall 2017 sampling trips) or ultrapure N₂ gas in a gas bag (summer 2018 sampling trip). Bottles were shaken for 2 minutes to equilibrate headspace gas with water, and headspace gas was extracted and immediately injected into 60 ml glass serum bottles that had been evacuated immediately prior to sample introduction. Samples were stored at room temperature until analysis, which occurred within one week of collection. Measured headspace concentrations were converted to dissolved concentrations using methods outlined in Pain et al. (2019). When atmospheric air was used for headspace extractions, atmosphere samples were collected in tandem and analyzed to correct each sample for calculated dissolved CO₂ and CH₄ concentrations and isotopic compositions. This correction altered CH₄ concentrations by up to 22% for one sample from the Russell glacier, though less than 5% for all other samples, and resulted in a correction of δ¹³C-CH₄ of up to 1.3‰. For CO₂, the correction altered concentrations by up to 15% for one sample collected at Kiattut Sermiat, though less than 10% for all other samples, and resulted in a correction of δ¹³C-CO₂ of up to 0.4‰.

For fall 2017 and summer 2018 sampling trips, alkalinity was measured in the field laboratory within 3 days of collection by titration with 0.01 N HCl using the Gran method. Because alkalinity measurements were not available for the spring 2017 sampling trip, we estimate alkalinity with PHREEQC modeling and the phreeqc.dat database (Parkhurst, 1997) using major cations and anions, pH, temperature, and DIC concentrations as model inputs.

2.3 Laboratory analysis

Gas samples were analyzed for CO₂ and CH₄ concentrations, and δ¹³C-CO₂ and δ¹³C-CH₄ on a Picarro G2201-i cavity ring-down spectrometer. Carbon isotopic compositions are reported in reference to Vienna Pee Dee Belemnite (VPDB). Check standards of known CO₂ and CH₄ concentrations and isotopic compositions were measured during each sample run and were accurate within 10%. Anion and cation concentrations were measured on an automated Dionex ICS-2100 and ICS-1600 Ion Chromatograph, respectively. Error on replicate analyses was less than 5%. DIC concentrations were measured on a UIC (Coulometrics) 5011 CO₂ coulometer coupled with an AutoMate Preparation Device. Samples were acidified and the evolved CO₂ was carried through a silver nitrate scrubber to the coulometer where total C was measured. Accuracy was calculated to be ±0.1 mg/L based on measurement of check standards. Dissolved ammonium (NH₄) concentrations were analyzed on a Seal AutoAnalyzer III. Error on check standards was less than 10%.

2.4 Methane modeling

To assess CH₄ sources and sinks, we calculate εₑ, or the carbon isotopic fractionation factor between CO₂ and CH₄ as defined in Whiticar (1999):

\[ \varepsilon_e = \delta^{13}C_{CO_2} - \delta^{13}C_{CH_4} \]
Values of $\varepsilon_c$ reflect methanogenesis pathways (acetoclastic or CO$_2$ reduction) as well as the extent of oxidation. Values of $\varepsilon_c$ between approximately 40 and 55‰ are produced for CH$_4$ produced via acetoclastic methanogenesis, while CO$_2$ reduction produces values between approximately 55 and 90‰. Lower values ($\varepsilon_c$ between 5 and 30) result when CH$_4$ oxidation predominates. Modern atmospheric input without additional alteration of CO$_2$ or CH$_4$ isotopic systematics results in a $\varepsilon_c$ value of approximately 40 (Whiticar, 1999).

We calculated CH$_4$ oxidation using the isotopic method outlined in Mahieu et al. (2008) and Preuss et al. (2013). The fraction of oxidized methane ($f_{ox}$) in an open system is given by:

$$f_{ox} = \frac{\delta_E - \delta_P}{1000(\alpha_{ox} - \alpha_{trans})}$$

(6)

where $\delta_E$ is the measured $\delta^{13}$C-CH$_4$ value for each water sample, $\delta_P$ is $\delta^{13}$C-CH$_4$ of produced methane, $\alpha_{ox}$ is the oxidation fractionation factor, and $\alpha_{trans}$ is a fractionation factor resulting from diffusive transportation of CH$_4$. While the exact value of $\delta_P$ is unknown, diagenetic alteration of $\delta^{13}$C-CH$_4$ values through oxidation or transport only enrich $\delta^{13}$C-CH$_4$ signatures, therefore the value of $\delta_P$ is taken as the most depleted $\delta^{13}$C-CH$_4$ signature assuming it is the least impacted by diagenetic alteration. Literature-reported values for $\alpha_{ox}$ range between 1.003 and 1.049. We calculate the fraction of oxidized methane with the largest fraction factor ($\alpha_{ox} = 1.049$; Mahieu et al., 2008), which yields the minimum amount of CH$_4$ oxidation required to explain the observed variations in $\delta^{13}$CH$_4$, and thus is a conservative estimate for CH$_4$ oxidation. Literature-reported values for $\alpha_{trans}$ vary from 1 for advection-dominated systems to 1.0178 for diffusion-dominated porous media (Visscher et al., 2004; Mahieu et al., 2008; Preuss et al., 2013). We assume that transport is advection dominated and thus assume $\alpha_{trans} = 1$.

### 2.5 Mineral weathering and carbonate modeling

We used major cation and anion concentrations and alkalinity to partition solutes into the four mineral weathering reactions in Table 1 after correcting solute concentrations for marine aerosol deposition using measured chloride concentrations and standard seawater element ratios. The mass balance model followed the methods of Deuerling et al. (2019). After apportioning solutes to mineral weathering reactions, we used the stoichiometries of reactions to calculate the impact of each reaction on dissolved CO$_2$ concentrations (Table 1). The mineral weathering model apports solutes to reactions in Table 1 based on the ratios of Ca/Na and Mg/Na in silicate minerals in stream bedload samples, which were taken to be 0.54 and 0.38, respectively, for Isunnguata and Russell Glacier samples (Deuerling et al., 2019; Hindshaw et al., 2014; Wimpenny et al., 2010, 2011) and 0.39 and 0.27, respectively, for Kiattut Sermiat samples (Da Prat and Martin, 2019). Because mineral weathering reactions may both add and remove CO$_2$, we discuss both the net impact of mineral weathering on CO$_2$ concentrations (Net CO$_2$-MW), which may have a positive or negative value:

$$[\text{Net CO}_2\text{-MW}] = [\text{CO}_2\text{-CarbCA}] + [\text{CO}_2\text{-CarbSA}] + [\text{CO}_2\text{-SilCA}]$$

(7)

as well as the total impact of mineral weathering on CO$_2$ concentrations (Total CO$_2$-MW),
[Total CO$_2$-MW] = \|[CO$_2$-CarbCA]\| + \|[CO$_2$-CarbSA]\| + \|[CO$_2$-SilCA]\| \ (8)

where changes in the concentrations of CO$_2$ are defined by their absolute values. To discuss the relative importance of individual reactions, we define proportional contributions of each reaction as follows:

\[
\%\text{CO}_2\text{-CarbCA} = \frac{\|[\text{CO}_2\text{-CarbCA}]\|}{[\text{Total CO}_2\text{-MW}]} \times 100 \ \ (9a)
\]

\[
\%\text{CO}_2\text{-CarbSA} = \frac{\|[\text{CO}_2\text{-CarbSA}]\|}{[\text{Total CO}_2\text{-MW}]} \times 100 \ \ (9b)
\]

\[
\%\text{CO}_2\text{-SilCA} = \frac{\|[\text{CO}_2\text{-SilCA}]\|}{[\text{Total CO}_2\text{-MW}]} \times 100 \ \ (9c)
\]

We combine measured CO$_2$ concentrations with Net CO$_2$-MW in order to determine the magnitude of CO$_2$ production or consumption in the subglacial environment due to processes besides mineral weathering. This analysis assumes that the concentration of CO$_2$ measured at the subglacial outlet is equivalent to the net change in CO$_2$ due to mineral weathering plus the sum of all other subglacial CO$_2$ sources and sinks. We refer to the sum of all other subglacial CO$_2$ sources and sinks as CO$_2$-total, which represents the amount of CO$_2$ that must have been supplied to the subglacial environment to balance the mineral weathering CO$_2$ sink:

\[
\text{CO}_2\text{measured} = \text{Net CO}_2\text{-MW} + \text{CO}_2\text{-total} \ \ (10)
\]

The sources of CO$_2$ to CO$_2$-total may be evaluated through the use of Keeling plots, which are constructed as the inverse of CO$_2$ concentrations ([CO$_2$]$^1$) versus stable isotopic composition ($\delta^{13}$C-CO$_2$). If variations in the concentration and isotopic composition of CO$_2$ arise from the mixing of two CO$_2$ reservoirs with constant isotopic compositions and concentrations (Keeling, 1958), a linear relationship is expected between [CO$_2$]$^1$ and $\delta^{13}$C-CO$_2$: The y-intercept of a regression between these variables represents the isotopic composition of the high-CO$_2$ end member. Because measured CO$_2$ concentrations include both subglacial CO$_2$ sources and sinks, which may include considerable consumption through mineral weathering reactions, the magnitude of the total subglacial CO$_2$ source is taken as CO$_2$-total. We therefore construct Keeling plots between [CO$_2$-total]$^1$ and measured $\delta^{13}$C-CO$_2$ values because while mineral weathering impacts the concentration of CO$_2$, its isotopic composition is not appreciably altered (Myrttinen et al., 2012) compared to the range of isotopic compositions of potential CO$_2$ end members, namely OM remineralization, atmospheric CO$_2$, and lithogenic CO$_2$ sources due to mechanical grinding (Fig. 1).

### 2.6 Discharge relationships

We evaluate the relationship between subglacial CH$_4$ and CO$_2$ dynamics and discharge using discharge records provided by PROMICE (Programme for Monitoring of the Greenland Ice Sheet; van As et al., 2018). Discharge records are collected at the outlet of the Watson River, which represents the combined discharge of Isunnguata and Russell glaciers as well as other outlet glaciers including the Leverett Glacier and major tributaries including the Orkendalen River (Fig. 2).
Watson River discharge estimates are therefore greater than the true discharge of our individual sampling locations, however we assume that discharge at the Isunnguata and Russell subglacial outlet sites is roughly proportional to Watson River discharge and exhibits similar temporal variability (Rennermalm et al., 2012). Because diurnal fluctuations in river discharge can be large, and differing water travel times from subglacial outlet sites to the Watson River mouth induces a lag of up to 8 hours between maximum daily discharge at subglacial discharge sites and the Watson River outlet, we compare subglacial CH₄ and CO₂ concentrations to average daily discharge, calculated as the average of hourly discharge estimates over the days on which subglacial discharge water samples were collected. Because no discharge information is available for Kiattut Sermiat, we assess discharge relationships at Watson River (Isunnguata and Russell) sites only.

3 Results

3.1 Temporal variability in water chemistry and gas concentrations

Chemical parameters differ between subglacial discharge sites as well as throughout the melt season. Specific conductivity (Sp.C; Fig. 3a) is typically highest at Kiattut Sermiat (26±8 μS/cm), followed by Russell (22±5 μS/cm) and Isunnguata sites (13±9 μS/cm; Fig. 3a). All sites show variability throughout the melt season, with lowest values occurring in the summer for Isunnguata and Russell, while Sp.C drops continuously throughout the melt season for KS. Sites differ in pH, and values at Kiattut Sermiat (8.2±0.4) are higher than both Russell (7.2±0.2) and Isunnguata (6.6±0.6; Fig. 3b), and while values vary throughout the melt season, no consistent trend is identified between sites. The saturation of dissolved oxygen (D.O.) with respect to atmospheric concentrations is similar between sites, though Isunnguata (98±8%) values fall below Russell (115±16%) and Kiattut Sermiat (117±11%) during all sampling times and exhibit undersaturation in the mid-summer, while Russell and Kiattut Sermiat are consistently supersaturated (Fig. 3c). Alkalinity is similar at Russell (93±31 μeq/L) and Kiattut Sermiat (93±26 μeq/L) which are higher than at Isunnguata (39±25 μeq/L), but all reach minimum values in summer (Fig. 3d). CH₄ concentrations differ by orders of magnitude between sites (Fig. 3e) and are consistently supersaturated with respect to atmospheric concentrations at Isunnguata (648±411 ppm or 1575±997 nM) and Russell (58±33 ppm or 110±78 nM), while close to atmospheric equilibrium at Kiattut Sermiat (4±2 ppm or 9±5 nM). δ¹³C-CH₄ values (Fig. 3f) are similar between Isunnguata (-54.7±7.5‰), Russell (-52±7.3‰), and Kiattut Sermiat (-57.6±14.2‰). Measured CO₂ concentrations (Fig. 3g) are consistently supersaturated with respect to atmospheric concentrations for Isunnguata (685±230 ppm or 58±18 μM), near atmospheric equilibrium for Russell (442±31 ppm or 29±4 μM) and undersaturated for Kiattut Sermiat (263±33 ppm or 19±2 μM). δ¹³C-CO₂ values (Fig. 3h) are lower in spring and fall for Isunnguata (-16.6±4.0‰) compared to Russell (-13.7±2.3‰) and Kiattut Sermiat (-16.1±1.6‰) sites, though similar seasonal variation occurs for all sites with relatively more depleted values in the spring and fall compared to summer.
3.2 Methane oxidation and relationship with discharge

Values of εc are similar throughout the melt season for Isunnguata (38±10‰) and Russell (38±9‰) and are relatively higher in the summer sampling period, while Kiattut Sermiat εc values are higher on average (42±13‰; Fig. 4a) with lowest values in the summer. Estimates of fox are similar between Isunnguata (17±15%), Russell (23±15%), and Kiattut Sermiat sites (25±22%; Fig. 4b). However, fox values are higher in the spring and fall sampling times compared to summer for Isunnguata and Russell and approach 50% in the spring, while Kiattut Sermiat values decrease throughout the melt season.

CH4 concentrations are unrelated to Watson River average daily discharge for Isunnguata, but significantly negatively correlated for Russell (Fig. 5a). The fraction of CH4 oxidized is moderately negatively correlated with discharge for both Isunnguata and Russell, though the relationship is not significant for either site (Fig. 5b). Discharge is positively correlated with εc for both Isunnguata and Russell (Fig. 5c). While the relationship is only significant for Isunnguata, the slopes and intercepts of regression lines for Isunnguata and Russell are similar.

3.3 Mineral weathering and CO2 models

Mineral weathering leads to net sequestration of CO2 at all three sites (Fig. 6a). The magnitude of Net ΔCO2 differs between sites with average values lowest at Isunnguata (-39±37 µM) followed by Russell (-65±32 µM) and Kiattut Sermiat (-98±17 µM). Individual mineral weathering reactions produce differing contributions between sites and over the melt season, with notable differences between Watson River sites (Isunnguata and Russell) and Kiattut Sermiat (Fig. 6b). For instance, the proportional contribution of CarbSA is similar between Isunnguata (17±11%) and Russell (15±6%), though is lower at Kiattut Sermiat (8±1%; Fig. 6b). Kiattut Sermiat has a relatively greater contribution from CarbCA (62±2%) compared to Isunnguata (41±10%) and Russell (38±6%), while SilCA is lower at Kiattut Sermiat (28±1%) compared to Isunnguata (41±17%), and Russell (47±11%). Kiattut Sermiat additionally exhibits low seasonal variability in the proportional contributions of individual mineral weathering reactions compared to Isunnguata and Russell sites.

CO2-total represents CO2 concentrations in the subglacial environment prior to addition and/or consumption of CO2 through mineral weathering (Eq. 10; Fig. 7). Because the Net CO2-MW is always negative (more consumption then production), the value of CO2-total is always greater than measured concentrations (CO2-measured). Regardless of differences in CO2-measured between sites, the average CO2-total values are similar between sites and average 91±47 µM for Isunnguata, 94±33 µM for Russell, and 117±16 µM for Kiattut Sermiat.

Keeling plots between [CO2-total]−1 and δ13C-CO2 indicate no linear relationship for Russell or Kiattut Sermiat samples, however a strong linear correlation is observed for Isunnguata (r²=0.99; p<0.001) with the removal of one outlier, which also had the lowest CO2-total value (Fig 8a). Isunnguata samples also show a significant positive correlation between the magnitude of CO2-total and NH4 concentrations (r²=0.73; p<0.05), unlike Russell or Kiattut Sermiat samples, though several Russell samples fall close to the regression line between CO2-total and NH4 of Isunnguata samples (Fig. 8b).
Because Isunnguata samples exhibit a linear relationship between [CO\textsubscript{2}-total] and \(\delta^{13}\text{C-CO}_2\) consistent with two endmember mixing, we utilize the \(\delta^{13}\text{C-CO}_2\) of samples and end members defined by the Keeling plot in an isotopic mixing model to calculate the relative contributions of CO\textsubscript{2} sources to CO\textsubscript{2}-total. We take the y-intercept (-27.5‰; Fig. 8a) of the Keeling plot regression as the \(\delta^{13}\text{C-CO}_2\) for the high-CO\textsubscript{2} end member (assumed to be generated by OM remineralization; CO\textsubscript{2-OM}) and assume atmospheric CO\textsubscript{2} as the low-CO\textsubscript{2} end member (CO\textsubscript{2-atm}; \(\delta^{13}\text{C-CO}_2 = -8\%o\)). The relative (%CO\textsubscript{2-atm} and %CO\textsubscript{2-OM}) contributions of these CO\textsubscript{2} sources to Isunnguata CO\textsubscript{2-total} vary throughout the melt season, with CO\textsubscript{2-OM} being the dominant source in the early and late melt season while CO\textsubscript{2-atm} is the dominant source in the mid-melt season (Fig. 9a). The absolute magnitude of CO\textsubscript{2-OM} is approximately 5-10 times greater in the early (89 \(\mu\)M) and late (117 \(\mu\)M) melt seasons compared to mid-melt seasons (13-23 \(\mu\)M), while the magnitude of CO\textsubscript{2-atm} varies relatively little throughout the melt season at around 50 \(\mu\)M (Fig. 9b). Both %CO\textsubscript{2-atm} and %CO\textsubscript{2-OM} are significantly (p<0.05) correlated with Watson River discharge, and the correlation is positive for %CO\textsubscript{2-atm} and negative for %CO\textsubscript{2-OM} (Fig. 9c). The magnitude of CO\textsubscript{2-OM} exhibits a power-law relationship with discharge and is highest at lowest discharge, while the magnitude of CO\textsubscript{2-atm} is invariable as a function of discharge (Fig. 9d).

4 Discussion

We observe orders of magnitude variability in dissolved CH\textsubscript{4} and CO\textsubscript{2} concentrations in subglacial discharge of the GrIS, indicating significant differences in the magnitudes of the sources and sinks of these gases across time and space. Supersaturation of both CO\textsubscript{2} and CH\textsubscript{4} with respect to atmospheric concentrations indicates that Isunnguata discharge is a source of both gases to the atmosphere, neighboring Russell Glacier discharges water that is a source of CH\textsubscript{4} but near equilibrium with respect to CO\textsubscript{2}, while Kiattut Sermiat in southern Greenland is a sink of atmospheric CO\textsubscript{2} but near equilibrium with respect to CH\textsubscript{4} (Fig. 3e, g). Because CH\textsubscript{4} dynamics may be largely microbially driven while CO\textsubscript{2} dynamics include microbial as well as abiotic mineral weathering processes, we first discuss CH\textsubscript{4} dynamics including a comparison of concentrations, isotopic compositions, and extent of oxidation between sites and over the melt season. We then discuss CO\textsubscript{2} concentrations, impacts of mineral weathering reactions (Table 1), and an assessment of subglacial CO\textsubscript{2} sources, including OM remineralization. These assessments will contribute to our understanding of the variability and controls of CH\textsubscript{4} and CO\textsubscript{2} concentrations in subglacial discharge from the GrIS and may improve predictions of the impact of future ice melt on Arctic carbon budgets.

4.1 Sources and sinks of CH\textsubscript{4}

Differences in CH\textsubscript{4} concentrations and relationships with discharge between sites imply heterogeneity in both the extent and controls of subglacial methanogenesis under the GrIS. CH\textsubscript{4} supersaturation occurs at the two subglacial discharge sites that flow to the Watson River (Isunnguata and Russell), and concentrations are similar to the ranges reported in discharge of the Leverett Glacier (up to 600 nM; Lamarche-Gagnon et al., 2019), adjacent to the Russell Glacier in this study. However,
CH$_4$ concentrations are near atmospheric equilibrium for the Kiattut Sermiat site (Fig 2e). Because methanogenesis is an anaerobic OM remineralization pathway, it is more likely to occur in subglacial environments isolated from atmospheric O$_2$ sources. Widespread observations of methanogenesis in glacial meltwater of southwest Greenland from this and other studies (Christiansen and Jørgensen, 2018; Dieser et al., 2014; Lamarche-Gagnon et al., 2019), and limited observations of CH$_4$ in subglacial discharge in southern Greenland, suggests heterogeneity in subglacial conditions that support methanogenesis. This heterogeneity may include variations in the distribution of subglacial organic substrates, such as subglacial permafrost reservoirs identified in previous studies (Ruskeeniemi et al., 2018), which were implicated as an OM source to drive methanogenesis under the Leverett Glacier (Lamarche-Gagnon et al., 2019). Organic deposits under the GrIS may have formed as a consequence of Holocene ice margin fluctuations following the encroachment of ice over previously established soils and terrestrial vegetation (Ruskeeniemi et al., 2018). Methanogenesis fueled by organic material overridden during ice sheet growth has been suggested as a potential climate feedback over glacial interglacial timescales (Wadham et al., 2008), and may contribute to variations in CH$_4$ concentrations between southwest and southern sites in this study.

Subglacial methanogenesis may additionally be controlled by hydrologic factors as the subglacial hydrological network develops throughout the melt season and channelization of meltwater conduits increases subglacial drainage efficiency (Andrews et al., 2015; Cowton et al., 2013). Drainage efficiency impacts both subglacial water residence time as well the transport of aerobic supraglacial meltwater to the ice bed. Both residence time and oxygen delivery may impact subglacial redox status and methanogenesis potential, and favor methanogenesis when oxygen supply rates are low compared to OM remineralization rates. This condition is most likely to be met in distributed subglacial systems that are hydrologically isolated with limited inputs from aerobic supraglacial meltwater. Such a hydrologic control on methanogenesis at the Russell Glacier is supported by the significant negative correlation between CH$_4$ concentrations and Watson River average daily discharge (Fig. 5a), suggesting that CH$_4$ production occurs predominantly during periods of low discharge and greater residence time. Alternatively, higher CH$_4$ concentrations during low discharge could result from the dilution of relatively small volumes of methanogenic subglacial meltwater with increasing volumes of aerobic supraglacial meltwater. The lack of relationship between CH$_4$ concentrations and discharge at the Isunnguata Glacier may indicate a greater influence of subglacial outburst events, in which hydrologically isolated methanogenic meltwater pockets are stochastically drained as the subglacial drainage network extends throughout the melt season (Fig. 5a). Subglacial outburst events were also implicated by heterogeneous CH$_4$ concentrations in subglacial discharge of the Leverett Glacier, and could contribute to heterogeneity in CH$_4$ concentrations at the Isunnguata Glacier (Lamarche-Gagnon et al., 2019).

While our results suggest heterogeneity in the extent and controls of methanogenesis between outlet glaciers, the microbial methanogenesis pathway as well as CH$_4$ oxidation dynamics are consistent between sites. Methanogenesis pathways may be evaluated by $\delta^{13}$C-CH$_4$ as well as $\varepsilon_c$ values because methanogenesis pathways impart different isotopic signatures to CH$_4$ and CO$_2$ (Whiticar and Schoell, 1986). Dieser et al. (2014) measured a microbial $\delta^{13}$C-CH$_4$ production signal at the Russell Glacier with values between -63‰ and -64‰, which was interpreted to reflect a possible combination of CH$_4$ produced through
both acetoclastic and CO₂ reduction pathways. The most depleted δ¹³C-CH₄ value measured at the Isunnguata in this study is close to that of Dieser et al. (2014) at -62.7‰ (Fig. 3f), and similar to values reported by Lamarche-Gagnon et al. (2019) for the Leverett Glacier, suggesting similar methanogenesis pathways across this region. While the exact contributions from each methanogenesis pathway cannot be inferred from isotopic information alone, the range of εc values at outlet glaciers are consistent with predominantly acetoclastic methanogenesis during the peak melt season (Fig. 4a). However, εc values fall below the expected range from acetoclastic methanogenesis during the early and late melt seasons, likely resulting from variations in the extent of subglacial CH₄ oxidation. Seasonal variation in CH₄ oxidation is supported by consistency between εc and fox values, which both indicate the greatest impact of oxidation (approaching 50%) in the early melt season compared to peak melt season (Fig. 4a, b), with additional evidence of elevated CH₄ oxidation in the late melt season at both Isunnguata and Russell glaciers.

The extent of CH₄ oxidation may be controlled by multiple factors including oxygen availability, subglacial residence times, and subglacial hydrology, similar to methanogenesis. A hydrologic control of CH₄ oxidation is supported by relationships between fox and εc with Watson River daily discharge at both Isunnguata and Russell Glaciers: fox is negatively related with discharge for both sites (Fig. 5b) while εc is positively correlated with discharge (Fig. 5c) although the correlation is only significant at Isunnguata. These correlations suggest that CH₄ oxidation is greatest during periods of low flow, which may be associated with greater residence times to allow subglacial CH₄ oxidation. The delivery of oxygen to the subsurface by supraglacial melting does not appear to be a limiting factor in subglacial CH₄ oxidation, which should increase fox as more oxygenated supraglacial water is delivered to the subglacial system. Instead, the observed greater CH₄ oxidation during periods of low discharge may reflect CH₄ oxidation following mixing between draining methanogenic subglacial meltwater pockets with aerobic subglacial meltwater. Longer transit times during periods of low flow may allow more subglacial methane oxidation to occur than during peak discharge, when the development of channelized flow paths reduces meltwater residence time in the subglacial environment.

Our results indicate a high degree of heterogeneity in subglacial methanogenesis under the GrIS, as well as a significant impact of CH₄ oxidation, which serves to reduce atmospheric CH₄ fluxes. Given the observed heterogeneity in this study, further investigation of the spatial variability in outlet glacier CH₄ concentrations is needed to determine the impact of GrIS loss on Arctic and global CH₄ budgets, while a better understanding of the controls of these differences will improve models of how CH₄ fluxes from subglacial discharge will change with continued warming.

### 4.2 Sources and sinks of CO₂

Dissolved CO₂ concentrations in subglacial discharge are consistently supersaturated with respect to atmospheric concentrations at Isunnguata Glacier, near atmospheric equilibrium at Russell Glacier, and undersaturated at Kiattut Sermiat Glacier, indicating that glacial meltwater from the GrIS can serve as either a source or sink of CO₂ to the atmosphere. Differences in dissolved CO₂ concentrations also imply variability in carbon processes under the GrIS (Fig. 3g). Because our
approach to assess the magnitude of subglacial CO$_2$ sources (including subglacial OM remineralization) depends in part on modelling results of CO$_2$ consumption by mineral weathering (Eq. 10), we first discuss the impacts of mineral weathering reactions, followed by a discussion of subglacial CO$_2$ sources, including OM remineralization.

### 4.2.1 Subglacial CO$_2$ sink: mineral weathering reactions

Although mineral weathering reactions may either increase or decrease dissolved CO$_2$ concentrations (Table 1), the net impact of mineral weathering at our study sites is to consume CO$_2$ (Fig. 6a). Net consumption occurs because the CO$_2$ source from Carb$_{SA}$ is ubiquitously low compared to sinks from either Carb$_{CA}$ or Sil$_{CA}$ (Fig. 6b). The range in Net CO$_2$-MW is similar between subglacial discharge sites (between 10-150 µM; Fig. 6a), but average values increase from Kiattut Sermiat to Russell to Isunnguata, likely reflecting the relative weatherability of alkaline igneous rocks, granulite facies gneisses, and amphibolite facies gneisses. Kiattut Sermiat is characterized by a relatively high proportion of Carb$_{CA}$ compared to Watson River sites, which may arise from the presence of trace carbonates in abundant readily weatherable basaltic intrusions as has been implicated in other studies (Urra et al., 2019). The relatively greater influence of carbonate dissolution compared to silicate dissolution on Total CO$_2$-MW at Kiattut Sermiat may also relate to more rapid dissolution kinetics of carbonates, which allow carbonate dissolution to have a large influence on major cation and anion load even when carbonates are only present in trace amounts (Tranter, 2005). At Isunnguata and Russell glaciers, Sil$_{CA}$ has a greater influence than Carb$_{CA}$ on Total CO$_2$-MW, which could result from either a lower abundance of trace carbonates to participate in weathering reactions, or relatively longer subglacial residence times that would allow a greater accumulation of silicate weathering products.

Despite the high impact of Carb$_{CA}$ on Total CO$_2$-MW at Kiattut Sermiat compared to Isunnguata and Russell sites, Carb$_{SA}$ is notably lower at Kiattut Sermiat than other sites and suggests a limited role for sulfuric acid weathering that may relate to subglacial sulfide oxidation dynamics. Lower abundances of sulfide minerals in the subglacial environment may limit the production of sulfuric acid, and could result from differences in lithology between sites, the depletion of sulfide minerals due to prior weathering (Graly et al., 2014), or weathering occurring in anoxic environments that limit the oxidation of sulfide to sulfuric acid (Deuerling et al., 2019). The kinetics of sulfide oxidation may also significantly differ between sites depending on the relative contributions of abiotic compared to microbially mediated sulfide oxidation, as microbially mediated sulfide oxidation is several orders of magnitude faster than abiotic sulfide oxidation. Rapid microbially mediated sulfide oxidation has been implicated in the development of anaerobic conditions, which could also support subglacial methanogenesis (Sharp et al., 1999). Observations of higher CH$_4$ concentrations as well as higher contributions of Carb$_{SA}$ at Isunnguata and Russell compared to Kiattut Sermiat in this study may therefore be linked to subglacial microbial activity, which is known to vary based on factors such as the presence of organic and fine-grained rock flour to serve as growth substrates, insulation from fluctuations in temperature, and delivery of nutrients and organic matter from supraglacial sources (Sharp et al., 1999). If microbially driven, our results suggest possible linkages between microbial processes and subglacial mineral weathering regimes, with significant impacts to both CH$_4$ and CO$_2$ dynamics due to the role of Carb$_{SA}$ as a CO$_2$ source (Table 1).
4.2.2 Subglacial CO₂ sources

Mineral weathering leads to net CO₂ consumption in all subglacial discharge samples, and thus the measured CO₂ concentrations represent only a fraction of the total CO₂ that would have been present in the absence of mineral weathering reactions (CO₂\textsubscript{total}; Eq. 10). CO₂ sources could include dissolution of atmospheric gases in air-filled conduits or fractures in ice, or CO₂ contained in ice bubbles (Fig. 1; Anklin et al., 1995; Graly et al., 2017). CO₂ may also be produced through mechanical grinding and volatilization of fluid inclusions (Macdonald et al., 2018) or OM remineralization. While previous studies have indicated that additional atmospheric CO₂ input through fractures and air-filled conduits may supply sufficient CO₂ to drive the observed extent of mineral weathering in many subglacial environments, including several sites in Greenland (Graly et al., 2017), CH₄ concentrations elevated above atmospheric equilibrium at the two Watson River sites reflects OM remineralization that would also contribute CO₂. While the magnitude of this source and its relative importance compared to other subglacial CO₂ sources is currently unknown, differing sources of carbonic acid for mineral weathering reactions carry different implications for subglacial CO₂ budgets. For instance, carbonic acid weathering driven by invasion of atmospheric CO₂ would represent a sink of atmospheric CO₂, but carbonic acid weathering driven by OM remineralization would instead serve to consume CO₂ from in situ sources and limit its potential as an atmospheric source. Determining the sources of carbonic acid to subglacial weathering reactions is therefore critical to understand the controls of mineral weathering in subglacial environments as well as the role of that process in atmospheric CO₂ sequestration.

Comparisons between measured δ¹³C-CO₂ in subglacial discharge samples and likely δ¹³C-CO₂ values of CO₂ sources indicate that CO₂ sources differ between sites, with OM remineralization as the predominant CO₂ source at the Isunnguata but not at Russell or Kiattut Sermiat glaciers. Keeling plots of [CO₂\textsubscript{total}⁻¹] versus δ¹³C-CO₂ indicate that CO₂\textsubscript{total} at Isunnguata discharge may be represented by a two-end member mixing model, in contrast to Russell and Kiattut Sermiat glaciers (Fig. 8). Mixing model end members include a ¹³C-enriched, lower-CO₂ source and a ¹³C-depleted, higher-CO₂ source (Fig. 8a). The y-intercept of the regression between [CO₂\textsubscript{total}⁻¹] versus δ¹³C-CO₂ (representing the isotopic signature of the high-CO₂ endmember) is -27.4‰, which is close to what would be expected from OM remineralization. For instance, CO₂ from remineralized OM in Greenlandic heath soils ranged between approximately -27 to -25‰ (Ravn et al., 2020), and between -20 to -30‰ for thawed Alaskan permafrost soils (Mauritz et al., 2019), both of which may be similar to subglacial organic matter. An additional correlation is observed between CO₂\textsubscript{total} and NH₄ concentrations for Isunnguata samples, which would be expected from OM remineralization (Fig. 8b). The low-CO₂ end member could reflect atmospheric CO₂ input, which should result in a δ¹³C-CO₂ value of approximately -8‰. While the δ¹³C-CO₂ value of the lowest-CO₂\textsubscript{total} samples in the Isunnguata Keeling plot (e.g. highest [CO₂\textsubscript{total}]⁻¹ not including the outlier) are slightly depleted compared to atmospheric values at -12.1‰ (Fig. 8a), even the lowest CO₂ concentrations measured at Isunnguata are supersaturated with respect to atmospheric concentrations (Fig. 3g). Supersaturation suggests that OM remineralization contributes CO₂ even for low CO₂-concentration samples and isotopically depletes the subglacial CO₂ reservoir.
While both atmospheric CO\textsubscript{2} (CO\textsubscript{2-atm}) and CO\textsubscript{2} derived from OM remineralization (CO\textsubscript{2-OM}) provide carbonic acid to drive subglacial mineral weathering as well as CO\textsubscript{2} supersaturation in Isunnguata discharge, their absolute and relative contributions are controlled by different processes. Understanding the controls of CO\textsubscript{2} acquisition may improve understanding of subglacial carbon dynamics as well as the conditions necessary for subglacial environments to become CO\textsubscript{2} sources to the atmosphere. CO\textsubscript{2-OM} is the dominant source to CO\textsubscript{2-total} at Isunnguata in the early and late melt seasons (Fig. 9a) when discharge is low (Fig. 9c), while CO\textsubscript{2-atm} is the dominant CO\textsubscript{2} source in the peak melt season when discharge is high. This switch in dominant CO\textsubscript{2} sources occurs because the magnitude of CO\textsubscript{2-OM} has a strong negative association with discharge, approaching zero during maximum discharge times (Fig. 9d), while CO\textsubscript{2-atm} remains relatively constant over the melt season (Fig. 9b) and the range of discharges (Fig. 9d). The high contributions of CO\textsubscript{2-OM} during low discharge could reflect higher residence times that allow greater biogeochemical modification and accumulation of OM remineralization reaction products, including CO\textsubscript{2}. The chemostatic behavior for CO\textsubscript{2-atm} indicates that invasion of atmospheric CO\textsubscript{2} is independent of the extent of chemical weathering, which exhibits strong seasonal variation (Fig. 6a). Chemostatic behavior of CO\textsubscript{2-atm} may indicate that CO\textsubscript{2-OM} maintains CO\textsubscript{2} concentrations at or above atmospheric saturation concentrations, and no additional atmospheric CO\textsubscript{2} dissolution would be needed to maintain equilibrium. Chemostatic behavior of CO\textsubscript{2-atm} could additionally indicate that the Isunnguata subglacial drainage is largely closed to atmospheric exchange. Both mechanisms are supported by the consistent CO\textsubscript{2} supersaturation observed in subglacial discharge at Isunnguata (Fig. 3g), which suggests that limited atmospheric exchange prevents significant outgassing prior to discharge. These results first imply that CO\textsubscript{2} supersaturation due to OM remineralization is likely during low flow conditions in systems that are relatively closed to the atmosphere. Moreover, CO\textsubscript{2-OM} is the main driver for mineral weathering during low-flow conditions: while CO\textsubscript{2-MW} exceeds -150 µM late in the melt season (Fig. 6a), CO\textsubscript{2-atm} provides only about a third of this CO\textsubscript{2} (Fig. 8b), suggesting that the remainder was driven by CO\textsubscript{2-OM}. In this case, only a fraction of the mineral weathering products measured in subglacial outflow at Isunnguata are directly involved in the sequestration of atmospheric CO\textsubscript{2}, and the majority of mineral weathering serves to consume CO\textsubscript{2} from \textit{in situ} CO\textsubscript{2-OM} production.

While δ\textsuperscript{13}C-CO\textsubscript{2} values of Russell and Kiattut Sermiat samples are within the range of Isunnguata samples, suggesting possible contributions of CO\textsubscript{2-atm} and CO\textsubscript{2-OM}, non-linear Keeling plots indicate variability in the CO\textsubscript{2} concentration and/or isotopic composition of end members, or significant contributions of at least one other major subglacial CO\textsubscript{2} source. We address both possibilities here. While atmospheric CO\textsubscript{2} should be relatively invariable, CO\textsubscript{2-OM} may vary both in concentration and isotopic composition, depending on variability in the quantity and composition of organic deposits, as well as remineralization rates. For instance, if remineralization largely occurs in hydrologically isolated subglacial meltwater pockets, some variability in the concentration and δ\textsuperscript{13}C-CO\textsubscript{2} of CO\textsubscript{2-OM} is likely. While no data yet exist to characterize the variability in subglacial OM reservoirs, variability in either concentration or isotopic composition of CO\textsubscript{2-OM} could plausibly result in non-linear Keeling plots here.
Additional subglacial CO₂ sources could include atmospheric CO₂ contained in ice bubbles, or lithogenic CO₂ liberated by mechanical grinding, though both of these sources would be expected to enrich rather than deplete the δ¹³C-CO₂ values of the samples relative to modern atmospheric δ¹³C-CO₂ values. Ice bubbles contain gaseous CO₂ at concentrations and isotopic compositions reflecting atmospheric conditions during ice formation. While heterogeneity may result from gas bubbles recording changes in atmospheric CO₂, variability in δ¹³C-CO₂ of gas bubble CO₂ should be only a few per mil, which is small compared to the variation observed in Russell and Kiattut Sermiat samples (Tipple et al., 2010; Fig. 8a). Gas bubble CO₂ should also be ¹³C-enriched compared to modern atmospheric CO₂ due to fossil fuel contributions, and thus would be unlikely to cause the variation in sample δ¹³C-CO₂ values that are more ¹³C-depleted than modern atmospheric δ¹³C-CO₂ values (Fig. 8a). Recent work has also highlighted the potential for subglacial mechanical grinding to result in CO₂ production through the volatilization of fluid inclusions (Macdonald et al., 2018). While volatilization of fluid inclusions through mechanical grinding was found to produce sufficient CO₂ to drive approximately 20% of mineral weathering in Svalbard subglacial environments, the expected isotopic composition of lithogenic CO₂ is more ¹³C-enriched than our measured δ¹³C-CO₂ values. Because mechanical grinding should produce CO₂ with an isotopic composition reflecting the lithogenic source, (Lüders et al., 2012), its contributions here are likely limited. For instance, estimates of δ¹³C for bulk hydrocarbons in fluid inclusions in the Ilímaussaq alkaline complex of South Greenland have values of -4.5±1.5‰ (Madsen, 2001), which is close to the δ¹³C-CO₂ of CO₂ in fluid inclusions in the Bamble granulite sector of South Norway (~ -6‰; Newton et al., 1980). There is an additional possibility of atmospheric exchange between the subglacial outlet site and our water sampling sites that could serve as an additional CO₂ source or sink. However, atmospheric CO₂ exchange after discharge would have the same impact on Keeling plots as atmospheric CO₂ exchange prior to discharge. Although Kiattut Sermiat CO₂ concentrations are undersaturated with respect to atmospheric concentrations and would promote invasion of atmospheric CO₂, measured δ¹³C-CO₂ values are more ¹³C-depleted than modern atmospheric CO₂ and are not consistent with atmospheric CO₂ as the sole or dominant source of CO₂ to glacial meltwater samples (Fig. 8a). While more information is needed to determine the sources of CO₂ to Russell and Kiattut Sermiat samples, δ¹³C-CO₂ values of samples from both sites imply mixing between a ¹³C-depleted CO₂ source, such as OM remineralization, and one or more ¹³C-enriched CO₂ sources, such as atmospheric or lithogenic CO₂.

5 Conclusions

Subglacial reactions impact the concentrations of CO₂ and CH₄ in subglacial discharge of GrIS, which act as either sources or sinks of GHG to the atmosphere. CH₄ concentrations of subglacial discharge are likely controlled by the availability of subglacial OM to drive methanogenesis as well as the extent of CH₄ oxidation. Regional differences in subglacial OM deposits may account for the occurrence of methanogenesis in southwest outlet glaciers (Isunnguata and Russell) in this and other studies, contrasting with southern Kiattut Sermiat where little CH₄ production occurs. During the early melt season, oxidation consumes nearly 50% of CH₄ produced at southwest sites, and relationships between discharge and CH₄ oxidation
(e$c$ and $f_{ox}$) suggest that CH$_4$ oxidation depends on longer subglacial residence time during periods of low discharge. While mineral weathering consumes CO$_2$ throughout the melt season at all three sites, additional CO$_2$ resupplied from atmospheric and subglacial sources increases the CO$_2$ concentrations of subglacial discharge. The magnitude of additional CO$_2$ sources (CO$_2$-total) is insufficient to maintain atmospheric equilibrium at Kiattut Sermiat, leading subglacial discharge to be a sink of atmospheric CO$_2$, while CO$_2$-total maintains close to atmospheric equilibrium concentrations at the Russell Glacier. At Isunnguata, however, OM remineralization produces more CO$_2$ than is consumed by mineral weathering and causes meltwater to be a source of CO$_2$ to the atmosphere. This finding implies that subglacial mineral weathering serves to partially or fully consume CO$_2$ produced from in situ sources under the GrIS but does not necessarily result in direct consumption of modern atmospheric CO$_2$. The important role of OM remineralization in subglacial environments of the GrIS determined by this and other studies also implies links between subglacial OM deposits and export of other biogeochemical solutes from the GrIS, including nutrients as well as redox-sensitive elements. While the export of nutrients from the GrIS has been the focus of numerous studies (Bhatia et al., 2013; Hawking et al., 2016; Lawson et al., 2014), little is currently known regarding the role of OM sources in governing these exports. Given the variability in GHG concentrations observed in this study, constraining the extent of heterogeneity in outlet glaciers of the GrIS as well as the biogeochemical, hydrologic, and geologic controls of this heterogeneity will be important for upscaling atmospheric fluxes as well as efforts to predict impacts of ice loss on carbon budgets due to current and future melting of the GrIS.

### Author Contribution

Jonathan B. Martin and Ellen E. Martin participated in conceptualization, data collection, data interpretation, reviewing and editing the manuscript, and acquired the funding for this project. Shaily Rahman participated in data collection and early interpretation and presentation of results. Andrea J. Pain participated in conceptualization, data collection, analysis, and interpretation, and took the lead on writing the manuscript with contributions from Jonathan B. Martin and Ellen E. Martin.

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element concentrations used for geochemical modeling (https://cn.dataone.org/cn/v2/resolve/urn:uuid:65d272f6-d280-4fcc-8aaa-4805f12ca6ae).

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Table 1. Mineral weathering reactions and impacts on dissolved CO$_2$ concentrations.

| Eq. | Mineral      | Acid     | Abbreviation$^1$ | Reaction                                                                 | Impact on CO$_2$ |
|-----|--------------|----------|------------------|--------------------------------------------------------------------------|------------------|
| (1) | Carbonate    | Carbonic | Carb$_{CA}$      | (Ca, Mg)CO$_3$ + H$_2$O + CO$_2$ $\rightarrow$ (Ca$^{2+}$, Mg$^{2+}$) + 2HCO$_3^-$ | CO$_2$ sink      |
| (2) | Sulfuric     | Carb$_{SA}$ |                | 2(Ca, Mg)CO$_3$ + H$_2$SO$_4$ $\rightarrow$ 2(Ca$^{2+}$, Mg$^{2+}$) + SO$_4^{2-}$ + H$_2$O + CO$_2$ | CO$_2$ source    |
| (3a) | Silicate     | Carbonic | Sil$_{CA}$      | (Ca, Mg)Al$_2$Si$_2$O$_8$ + 2CO$_2$ + 3H$_2$O $\rightarrow$ (Ca$^{2+}$, Mg$^{2+}$) + 2HCO$_3^-$ + Al$_2$Si$_2$O$_5$(OH)$_4$ | CO$_2$ sink      |
| (3b) |             |          |                 | (Na, K)AlSi$_3$O$_8$ + CO$_2$ + 5.5H$_2$O $\rightarrow$ (Na, K) + HCO$_3^-$ + 0.5Al$_2$Si$_2$O$_5$(OH)$_4$ + 2H$_4$SiO$_4$ | CO$_2$ sink      |
| (4a) | Sulfuric     | Sil$_{SA}$ |                | (Ca, Mg)Al$_2$Si$_2$O$_8$ + H$_2$SO$_4$ + H$_2$O $\rightarrow$ (Ca$^{2+}$, Mg$^{2+}$) + SO$_4^{2-}$ + Al$_2$Si$_2$O$_5$(OH)$_4$ | No impact        |
| (4b) |             |          |                 | 2(Na, K)AlSi$_3$O$_8$ + H$_2$SO$_4$ + 9H$_2$O $\rightarrow$ 2(Na$^+$, K$^+$) + SO$_4^{2-}$ + Al$_2$Si$_2$O$_5$(OH)$_4$ | No impact        |

$^1$Abbreviations are based first on the mineral class (carbonate = carb, silicate = sil) and then on the acid (carbonic acid = CA, sulfuric acid = SA)
Figure 1: Conceptual diagram of subglacial sources and sinks of CO$_2$ and CH$_4$. Arrow indicate the direction of fluxes. Boxes represent processes, and sources of gases to subglacial meltwaters are indicated by green text while sinks of gases to subglacial meltwater are indicated by red text. Gas bubbles, mechanical grinding, and OM remineralization are grouped because all are CO$_2$ and CH$_4$ sources.
Figure 2. © Google Earth satellite images of study locations in a) Greenland including b) Isunnguata (dark blue circle) and Russell (light blue square) subglacial discharge sites flowing to the Watson River in southwest Greenland (c) and the Kiattut Sermiat site (orange triangle) in southern Greenland. Discharge monitoring of the Watson River is collected by PROMICE (van As et al., 2018) at the location represented by the yellow star.
Figure 3. Chemical parameters at Isunnguata (IS), Russell (RU) and Kiattut Sermiat (KS) subglacial discharge sites versus day of year for a) specific conductivity (Sp.C), b) pH, c) dissolved oxygen (D.O.) percent saturation, d) alkalinity (Alk), e) measured CH$_4$ concentrations (left y-axis in ppm and right y-axis in nM), f) $\delta^{13}$CH$_4$ values, g) measured CO$_2$ concentrations (left y-axis in ppm and right y-axis in µM), and h) $\delta^{13}$C-CO$_2$ values. Atmospheric equilibrium concentrations are indicated by dashed lines and taken as 1.9 ppm for CH$_4$ and 410 ppm for CO$_2$. Error bars on CH$_4$ and CO$_2$ concentrations and stable isotopic compositions represents the standard deviation of replicates and are smaller than symbols for some data points.
Figure 4. CH$_4$ dynamics over the course of the 2017 and 2018 melt seasons including a) the carbon fractionation factor ($\varepsilon_C$) between dissolved CO$_2$ and CH$_4$ and b) the fraction of CH$_4$ oxidized ($f_{OX}$) for Isunnguata (IS), Russell (RU) and Kiattut Sermiat (KS) samples. Fields of $\varepsilon_C$ representing methanogenesis and oxidation values are based on Whiticar (1999). Values of $\varepsilon_C$ between approximately 40 and 55 are produced for methanogenesis via acetate fermentation, while CO$_2$ reduction produces values between approximately 50 and 90. Lower values result from a predominant isotopic signature of CH$_4$ oxidation. Atmospheric input without additional alteration of CO$_2$ or CH$_4$ isotopic systematics results in a $\varepsilon_C$ value of approximately 40.
Figure 5. Discharge relationships with CH$_4$ dynamics including a) CH$_4$ concentrations, b) $f_{ox}$, and c) $e_c$ for Isunnguata (IS) and Russell (RU) samples. Regressions are shown by dotted lines for Isunnguata and dashed lines for Russell samples. Average daily discharge estimates are for the outlet of the Watson River and provided by PROMICE (van As et al., 2018). Horizontal error bars represent the standard deviation of average daily discharge for days samples were collected and are smaller than symbols for some data points.
Figure 6. Mineral weathering model results in a) net impact of mineral weathering reactions on CO$_2$ (Net CO$_2$-MW; Eq. 7) for Isunnguata, Russell, and Kiattut Sermiat subglacial discharge sites (where negative values of Net CO$_2$-MW indicate net sequestration of CO$_2$ due to mineral weathering), and b) the proportional contribution of each mineral weathering reaction to the total change in CO$_2$ (% Total CO$_2$-MW Eq. 9a-9c).
Figure 7. Calculated CO$_2$-total values for a) Isunnguata, b) Russell, and c) Kiattut Sermiat subglacial discharge against day of the year.
Figure 8. Correlations between the magnitude of CO$_2$-total and a) $\delta^{13}$C-CO$_2$ and b) NH$_4$ for Isunnguata (IS), Russell (RU) and Kiattut Sermiat (KS) samples. Asterisk denotes the Isunnguata outlier not included in the regression between CO$_2$-total and $\delta^{13}$C-CO$_2$ in panel a).
Figure 9. Estimates of contributions of atmospheric CO$_2$ (CO$_2$-atm) compared to that derived from remineralized OM (CO$_2$-OM) for Isunnguata subglacial discharge site including a) relative contributions according to the day of the year, b) absolute magnitudes according to the day of the year, c) relative contributions according to Watson River average daily discharge, and d) absolute magnitudes according to Watson River average daily discharge. Average daily discharge estimates are for the outlet of the Watson River and provided by PROMICE (van As et al., 2018). Horizontal error bars represent the standard deviation of average daily discharge for days samples were collected and are smaller than symbols for some data points. Regressions between average daily discharge and proportional and absolute contributions of CO$_2$-atm and CO$_2$-OM sources are shown with dotted lines.