Estimates of net community production from multiple approaches surrounding the spring ice-edge bloom in Baffin Bay

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Measurements of net community production (NCP) provide an upper constraint on the strength of the oceanic biological pump, the dominant mechanism for removing CO2 from the ocean surface and sequestering it at depth. In this investigation, our objectives were to describe the spatial and temporal variability of NCP associated with the spring ice-edge bloom in Baffin Bay and to identify the key environmental drivers controlling its variability. Using data collected between June 9 and July 10, 2016, we estimated NCP based on (1) underway measurements of surface water oxygen to argon ratios (O2:Ar), (2) underway measurements of the partial pressure of CO2, and (3) seasonal nitrate drawdown from discrete samples. These multiple approaches displayed high NCP (up to 5.7 mol C m$^{-2}$) in eastern Baffin Bay, associated with modified Atlantic waters, and low NCP (<1 mol C m$^{-2}$) in the presence of Arctic outflow waters in western Baffin Bay. Arctic outflow waters were characterized by low surface salinities and nitrate concentrations, suggesting that high freshwater content may have limited the nutrient availability of these waters. Different integration depths and timescales associated with each NCP approach were exploited to understand the temporal progression and succession of the bloom, revealing that the bloom was initiated under ice up to 15 days prior to ice retreat and that a large portion of NCP in eastern Baffin Bay (potentially up to 70%) was driven by primary production occurring below the surface-mixed layer.

Keywords: Net community production, Biological pump, Arctic, Ice-edge, Spring bloom

Introduction

Like other Arctic Seas, the waters of Baffin Bay are currently transforming under the influence of anthropogenic climate change. Sea-ice cover in Baffin Bay continues to decline in thickness and longevity, with the ice-free season growing longer (Tang et al., 2004; Landy et al., 2017; Bliss et al., 2019). This change results in warmer sea surface temperatures, especially in eastern Baffin Bay (Timmermans and Ladd, 2018), and earlier commencement of spring phytoplankton blooms (Kahru et al., 2011; Bélanger et al., 2013). Additionally, freshwater inputs to Baffin Bay continue to increase from various sources including increasingly fresh surface waters exported from the central Arctic Ocean (Haine et al., 2015; Carmack et al., 2016) and increasing glacial meltwater inputs from the Greenland ice sheet (Dahl-Jensen et al., 2011; Bamber et al., 2018). Combined, these changes will no doubt affect the primary production regime in Baffin Bay and have consequent effects for marine ecosystems and the marine carbon cycle in this region.

Observations indicate that the Arctic Ocean represents a net sink of atmospheric CO2 (both natural and anthropogenic), absorbing approximately 166–180 Tg C yr$^{-1}$ (MacGilchrist et al., 2014; Yasunaka et al., 2018). Primary production and the biological pump are key processes driving this net CO2 uptake through the production and subsequent export of organic carbon from the surface into the deep ocean, where it can be sequestered from interaction with the atmosphere. Autotrophic organisms, such as algae and phytoplankton, utilize light and dissolved inorganic carbon to produce organic carbon, a portion of which may then be consumed and respired by heterotrophic organisms converting organic carbon back into inorganic carbon. Net community production (NCP) is defined as the difference between gross primary production and ecosystem respiration. As such, NCP represents an upper constraint on the amount of organic carbon exported to depth (the biological pump), assuming that the storage of organic carbon in the surface-mixed layer is modest over large spatial and temporal scales (Falkowski et al., 2003).

Primary production is a highly seasonal process in the Arctic Ocean and its adjacent seas due to the requirements...
for sufficient light and nutrients. The retreat and melting of sea ice in the spring drive strong primary production because of the enhanced light availability and stratification from sea-ice meltwater in the marginal ice zone (Barber et al., 2015; Leu et al., 2015). Enhanced stratification surrounding the ice edge limits the depth of vertical mixing, allowing phytoplankton to remain near the surface with sufficient light for photosynthesis. Ice-edge blooms are a common occurrence throughout the Arctic Ocean (Perrette et al., 2011) and are generally intense and short-lived features that quickly deplete the nutrient stock within the shallow surface-mixed layer (Niebauer, 1991; Sakshaug, 2004; Tremblay et al., 2008). However, much still remains unknown about the conditions required to trigger the phytoplankton spring bloom in the presence of sea-ice cover. In the case of ice-edge blooms, increased light availability and stratification are occurring simultaneously, creating an ideal setting for bloom initiation. However, many instances of under-ice blooms have now been observed throughout the Arctic Ocean (Fortier et al., 2002; Mundy et al., 2009; Arrigo et al., 2012; Assmy et al., 2017), demonstrating that cracks, leads, melt ponds, or thin ice without snow cover can allow for sufficient light transmission to initiate under-ice pelagic blooms. Nearly 30% of the ice-covered Arctic Ocean is now estimated to permit under-ice blooms in July (Horvat et al., 2017), which raises the question of whether enhanced stratification is a necessity to initiate a bloom or whether sufficient light availability is enough.

In the post-bloom phase, after the surface bloom has depleted the nutrient stock of the surface-mixed layer, primary production may continue in deeper waters as subsurface chlorophyll maxima (SCMs). These features are quite common in the post-bloom Arctic Ocean (Tremblay et al., 2008; Martin et al., 2010; Ardyna et al., 2013; Brown et al., 2015) and will cause the nitracline to deepen as they continue to consume nutrients at greater depths. Ultimately, the SCM will settle at a depth between the nitracline and the bottom of the euphotic zone balancing the requirements for light and nutrients. Despite the known ubiquity of SCMs, much remains to be understood about their ecological significance and their contribution to NCP.

Recent investigations focused on northern Baffin Bay (in the historical location of the North Water Polynya) and have documented decreases in annual NCP and phytoplankton biomass, attributing this decrease to freshening and increased water column stratification (Bergeron and Tremblay, 2014; Blais et al., 2017). However, these results may not be widely applicable to all of Baffin Bay, as water mass characteristics and sea-ice conditions vary widely from east to west. In fact, Bélanger et al. (2013), using a satellite-based model to assess primary production trends from 1998 to 2010, showed positive trends in primary production rates in southeastern Baffin Bay, despite negative trends in northern Baffin Bay. The synthesis study of Codispoti et al. (2013) was the last investigation to specifically present estimates of NCP in southern Baffin Bay. However, the results of that synthesis study were calculated using data from relatively few sampling stations averaged over large spatial scales, and winter nutrient concentrations used as a baseline to estimate NCP over the West Greenland shelf were collected only between 1962 and 1987. Therefore, we predict that the NCP estimates of Codispoti et al. (2013) will differ greatly from those calculated with more recent measurements. Further, the unique set of methods we used in this investigation allows us to provide additional information on the temporal progression of the bloom across Baffin Bay and to distinguish NCP contributions from the surface-mixed layer and SCM.

In this study, we present springtime estimates of NCP across southern Baffin Bay that are associated with a receding ice cover and a transient ice-edge bloom. The main objectives of this study are to describe spatial and temporal variability of NCP in Baffin Bay and to provide insight into the environmental factors driving this variability. To infer information about the spatial and temporal progression of the bloom, we make use of three different approaches of estimating NCP that integrate over different timescales and depth. Our three approaches to estimating NCP rely on (1) sea surface measurements of the oxygen-to-argon ratio (O$_2$/Ar), (2) sea surface measurements of the partial pressure of CO$_2$ (pCO$_2$), and (3) discrete water column measurements of nitrate concentrations.

**Oceanographic setting**

The oceanographic setting of Baffin Bay plays a critical role in determining the amount of primary production that can occur, as sea-ice conditions and water mass circulation patterns dictate the availability of light and nutrients in the surface ocean. Within Baffin Bay, an overall cyclonic circulation is comprised of two main currents: the southward flowing Baffin Island Current (BIC) that exports cold Arctic outflow waters southward to the Labrador Sea and the northward flowing West Greenland Current (WGC) that mainly advects warm modified Atlantic waters northward along the West Greenland shelf (see Figure 1). These two current systems lead to an east–west divide in the water mass assemblies of Baffin Bay (Bâcle et al., 2002; Tang et al., 2004; Münchow et al., 2014; Randelhoff et al., 2019). Sea-ice retreat in Baffin Bay typically progresses from east to west, due to the temperature gradient between the WGC and BIC current systems, and features an ice-edge bloom that tracks the westward moving ice edge (Perrette et al., 2011). Randelhoff et al. (2019) recently provided a detailed summary of the spring sea-ice retreat and water masses present in Baffin Bay. Here we briefly review the major water masses of Baffin Bay (although Baffin Bay deep waters below 800 m are not discussed here) and refer the reader to Randelhoff et al. (2019) and references therein for more detail.

In eastern Baffin Bay, the water column is divided into three main water masses: (1) warm surface waters, characterized by $T > 0$ °C and $S < 33.5$, which are only present during the ice-free season when solar insolation heats the upper mixed layer; (2) winter Atlantic water (WW), with $T < -1$ °C and $33.5 < S < 34$ located at depths of approximately 40–100 m; and (3) Atlantic waters (AW), characterized by $T > 1$ °C and $S > 34$ and concentrated at depths between 150 and 500 m over the continental slope.
In western Baffin Bay, AW and WAW are still present but are located at greater depths in the water column. AW is generally located at depths greater than 250 m, and WAW is located between depths of 150 and 250 m. Above the WAW sits Arctic water, with $T < -1 \, ^\circ{C}$ and $S < 33.5$, which arrives in Baffin Bay from the channels of the Canadian Arctic Archipelago (CAA) and is transported southward by the BIC. Due to the presence of multiple shallow sills in the CAA, Arctic water entering the Baffin Bay is primarily of Pacific origin, as deeper Atlantic-origin waters from the central Arctic Ocean basins are held back. Similarly, the deep Atlantic waters in Baffin Bay (WAW and AW) cannot pass over these sills, and so, these waters are recirculated southward within Baffin Bay, subducting underneath the Arctic water.

Nitrate delivery has been shown to vary between the two major surface currents in Baffin Bay. Arctic water entering via the BIC was reported to deliver between 5 and 12 $\mu$mol kg$^{-1}$ of nitrate, whereas WGC waters entering from the southeast contained approximately 14.5 $\mu$mol kg$^{-1}$ nitrate in winter (Tremblay et al., 2002).

**Methods**

**Sampling**

Data presented here were collected between June 9 and July 10, 2016, onboard the CCGS Amundsen as part of the Green Edge expedition (Randelhoff et al., 2019). A series of seven longitudinal transects were sampled across the retreating ice edge. Each transect, numbered from 100 to 700 in the order of sampling (see Figure 1), traversed between modified Atlantic waters in eastern Baffin Bay and Arctic outflow waters in the west. Midway through the research cruise (between transects 300 and 400), the ship traveled south to the hamlet of Qikiqtarjuaq, Nunavut, for a personnel swap with the associated Green Edge ice camp (Oziel et al., 2019).

Throughout the cruise, the ship’s conductivity, temperature, and depth (CTD)/Rosette system, consisting of 24 × 12 L Niskin-type bottles (OmegaTest Equipment) with a Sea-Bird 911plus CTD, was deployed at a total of 135 unique station locations. The CTD was also equipped with a fluorometer for the detection of chlorophyll $a$ (SeaPoint). Nutrient samples (for analysis of nitrite, nitrate, ammonium, phosphate, and silicate) were collected at 92 of these stations, while discrete samples for salinity, stable oxygen isotope ($^{18}O$), and the inorganic carbon system (for analysis of total alkalinity [TA] and dissolved inorganic carbon [DIC]) were only collected at 11 stations in the 100 and 300 transects. Underway surface water measurements of $O_2$/Ar and $pCO_2$ were collected through two separate seawater intake lines, sampling at depths of approximately 5 and 7 m below the waterline, respectively. Underway measurements of surface $pCO_2$ and $O_2$ concentrations showed good correlation with surface measurements from the CTD/Rosette (see Figure S1), and therefore, we do not anticipate that biofouling within the seawater intake lines had any significant effect on our measurements.

**Analytical procedures**

Continuous underway $\Delta O_2$/Ar measurements were made using equilibrator inlet mass spectrometry (EIMS; Cassar et al., 2009). Briefly, seawater from the ship’s seawater intake line (sampling from a nominal depth of 5 m) is pumped through a gas-permeable membrane cartridge, allowing the gas in the headspace of the cartridge to equilibrate with the gases in seawater. The $O_2$/Ar ratio is then measured from the headspace gas using a quadrupole mass spectrometer (Pfeiffer Prisma model QMG 220 M1). The system is calibrated against the atmospheric $O_2$/Ar ratio every 4 h, and the instrument precision is $\pm 0.3\%$.  

![Figure 1. Map of the study area. The left map shows the general surface circulation patterns of Baffin Bay, with the West Greenland Current shown in red and the Baffin Island Current shown in blue. The inset map on the right shows the cruise track, colored according to sampling date, with east–west transects named sequentially in the order of sampling from 100 to 700. DOI: https://doi.org/10.1525/elementa.013.f1](http://online.ucpress.edu/elementa/article-pdf/8/1/013/439752/elementa.013.pdf)
or better (Cassar et al., 2009). Here we present 2-min averages of $\Delta O_2/Ar$ measurements collected every 8 s, following Eveleth et al. (2017).

Continuous measurements of sea surface $pCO_2$ were also made underway using an infrared-based showerhead equilibrator system (General Oceanics model 8050; Pierrot et al., 2009) using a different clean seawater intake line into the ship’s engine room at 7-m depth. The infrared gas analyzer (LI-COR model LI-7000) was calibrated twice daily against four certified gas standards ($CO_2$ concentrations of 0.0, 254.5, 431.3, and 554.2 ppm) traceable to World Meteorological Organization standards. The underway system also houses a flow-through CTD (Idronaut model Ocean Seven 315) that provided coincident salinity measurements. Due to the dependency of the $pCO_2$ measurement on temperature, a separate temperature measurement was made by a thermocouple located directly in the seawater intake line (accurate to 0.5 °C). These “in situ” temperature measurements were used to correct for any warming that may have occurred during transport of water to the $pCO_2$ system’s equilibrator using the method of Takahashi et al. (1993). The overall relative uncertainty of the temperature-corrected $pCO_2$ values is estimated to be 2%.

Carbon system seawater samples were collected following standard protocols (Dickson et al., 2007), with replicates collected from at least one Niskin bottle per rosette cast. Samples were collected in 250-mL glass bottles, preserved with 100 µL of saturated mercuric chloride solution, capped with ground glass stoppers greased with Apiezon M, and sealed with electrical tape. Samples were then stored in the dark at 4 °C until arrival at the Institute of Ocean Sciences in Sidney, British Columbia, where analysis for TA and DIC was conducted. The coulometric DIC analysis utilized either a SOMMA (Johnson et al., 1993) or VINDTA 3D (MARIANDA) extraction system. Measurements of TA were determined using an open-cell potentiometric titration with nonlinear least squares endpoint determination (Dickson et al., 2007). Both of these measurements were calibrated against certified reference materials (CRM Batch #106) provided by Andrew Dickson (Scripps Institute of Oceanography). Determination of duplicate DIC samples indicates a precision of $\pm 1$ µmol kg$^{-1}$, and duplicate TA samples show a precision of $\pm 3$ µmol kg$^{-1}$.

Samples for the determination of $\delta^{18}O$ were analyzed at the GEOTOP stable isotope laboratory at the Université du Québec à Montréal. Measurements were made using the $CO_2$ equilibration method, where 200 µL of sample water was equilibrated with $CO_2$ for 7 h at 40 °C. The $CO_2$ was then analyzed on a Micromass Isoprime™ universal triple collector mass spectrometer in dual inlet mode with an AquaPrep™ system (Isoprime Ltd., Cheadle, UK). Results are expressed in the $\delta$ notation in % versus Vienna Standard Mean Ocean Water. For each analytical sequence, two internal reference water samples were used to normalize the sample data ($\delta^{18}O = -6.71\%$ and $-20.31\%$). Uncertainties in replicate measurements were $\pm 0.05\%$ (1σ).

Vertically resolved subsamples for nutrients determination were collected directly from the Niskin bottles with syringes and filtered through a GF/F filter inserted in a Swinnex. The concentrations of phosphate and nitrate (obtained by the difference between nitrite and nitrate + nitrite) were measured fresh on an AutoAnalyzer 3 (Bran and Luebbe) and used to estimate the fraction of Pacific water (FPW) at each sampling depth. Here we used the N-P regression lines from Jones et al. (1998), as Tremblay et al. (2015) showed the N-P relationships in northern Baffin Bay closely follow those derived by Jones et al. (1998). In order to estimate surface FPW along the ship track, average mixed layer FPW values were calculated at each sampling station (as the average of all FPW values within the defined mixed layer depth). These mixed layer average values were then linearly interpolated between sampling station locations. Surface-mixed layer depths were defined following Shaw et al. (2009).

### NCP from $O_2/Ar$

Rates of NCP calculated from $O_2/Ar$ measurements exploit the similar physical properties but different biological behavior of oxygen and argon (Craig and Hayward, 1987). Since physical processes affecting oxygen concentrations (e.g., bubble entrainment and temperature changes) similarly affect argon (Ar), the biological oxygen saturation anomaly ($\Delta O_2/Ar$) can be calculated as

$$\Delta O_2/Ar = \left[ \frac{[(O_2/][Ar])_{sat} - 1}{[(O_2/][Ar])_{sat}} \right],$$  \hspace{1cm} (1)

where $[(O_2/][Ar])_{sat}$ is the measured $O_2/Ar$ ratio from the EIMS and $[(O_2/][Ar])_{sat}$ is the ratio of equilibrium concentrations of $O_2$ (Garcia and Gordon, 1992) and Ar (Hamme and Emerson, 2004). $\Delta O_2/Ar$ represents the percent deviation in seawater $O_2/Ar$ from atmospheric equilibrium, with Ar normalization used to remove physical effects on $O_2$ saturation (Craig and Hayward, 1987). Assuming that the surface-mixed layer is isolated from the deeper water column, NCP can be calculated as the sea-to-air flux of biologically produced oxygen:

$$\text{NCP}_{O_2/Ar} \text{(mmol C m}^{-2} \text{d}^{-1}) \approx \frac{k_{O_2} \cdot [O_2]_{sat} \cdot \Delta(O_2/Ar)}{O_2 \cdot C},$$  \hspace{1cm} (2)

where $k_{O_2}$ is the gas transfer velocity for $O_2$ (m day$^{-1}$), $[O_2]_{sat}$ is the equilibrium concentration (mmol O$_2$ m$^{-3}$), and $O_2 : C$ is the stoichiometric ratio of 1.4 from Laws (1991). Within Equation 2, the term $[O_2]_{sat} \cdot \Delta(O_2/Ar)$ represents the sea-to-air gradient of biologically derived oxygen. The transfer velocity ($k_{O_2}$) is calculated from the wind speed parameterization of Wanninkhof (2014) using $4 \times$ daily wind components from the National Center for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) reanalysis data (Kalnay et al., 1996) and is weighted to account for wind speed history over the 60 days prior to sampling following the weighting scheme of Reuer et al. (2007) as modified by Teeter et al. (2018). The $k_{O_2}$ was also linearly corrected for sea-ice cover using sea-ice concentrations (SICs) from the Advanced Microwave Scanning Radiometer 2 (AMSR2; Srensen et al., 2008) matched to the ship’s sampling time and location. We chose to apply a linear scaling of $k_{O_2}$ by...
This large uncertainty is again applied by our residence-time multiplier to report seasonal NCP in units of mol C m⁻². We find that the relative uncertainty of our NCP₂⁻/Ar estimates may be more than 100% (especially in ice-covered waters). However, these estimates represent the best approximations that we can make using the O₂/Ar technique in partially ice-covered waters.

**NCP from pCO₂**

Calculated values of NCP₃=pCO₂ utilize the seasonal drawdown of DIC in the surface-mixed layer to estimate the total growing season NCP (mol C m⁻²). NCP₃=pCO₂ was calculated as

\[
NCP_{pCO}_2 = MLD(DIC_W \cdot \rho_W - nDIC_S \cdot \rho_S) + FCO_2,
\]

where MLD is the mixed layer depth, DIC_W is the winter water DIC concentration, nDIC_S is the salinity-normalized surface DIC concentration (normalized to the winter water salinity) derived from pCO₂, and ρ_W and ρ_S are the winter and surface water densities, respectively. The FCO₂ term represents a correction for the air–sea exchange of CO₂ over the days of open water (DOW; see Methods section on SIC) prior to the sampling date. MLD values were estimated from CTD profiles and were defined as the depth at which the density increases from its surface value to 20% of the difference between 100 m and the surface following Shaw et al. (2009). MLD values were then interpolated linearly along the ship track between stations and were assumed to remain constant seasonally.

Surface measurements of pCO₂, temperature, salinity, and salinity-based TA estimates were used to calculate surface water DIC concentrations (DIC_S) using CO2SYS (van Heuven et al., 2011) with the carbonic acid dissociation constants of Lueker et al. (2000) and those of Dickson (1990) for bisulfate. Two separate linear salinity-TA relationships were defined: one for Arctic outflow waters in western Baffin Bay (TA_ATL = 43.69 + Sd + 816.98, R² = 0.67, root mean square error [RMSE] = 11.2 µmol kg⁻¹) and a separate relationship for modified Atlantic waters in eastern Baffin Bay (TA_ATL = 58.27 + Sd + 287.82, R² = 0.86, RMSE = 7.3 µmol kg⁻¹). All samples used to define the Arctic outflow salinity-TA relationship coincided with FPWs greater than or equal to 0.3, whereas all samples used to define the Atlantic-modified water relationship

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**Table 1. Summary of relevant integration timescales and depths of each NCP approach. DOI: https://doi.org/10.1525/elementa.013.t1**

| NCP Approach   | Integration Timescale (Residence Time of Tracer) | Integration Depth       |
|----------------|-----------------------------------------------|-------------------------|
| NCP₂⁻/Ar       | Approximately 20 days* in open water (σ = 6 days; range of 8–32 days); approximately 60 days* under ice (σ = 45 days; range of 12–193 days) | Surface-mixed layer     |
| NCP₃=pCO₂      | Months to years                                |                          |
| NCP₄           | Months to years                                | Euphotic zone            |

*Approximate average values for open water and ice-covered conditions. All locations with sea-ice concentration above zero are considered ice-covered.
had FPW = 0 (see Figures S2 and S3). When estimating surface TA at any point along the ship track, we thus used interpolated MLD-average FPW values (FPWMLD) to determine which linear salinity-TA relationship to use. For FPWMLD ≥ 0.3, the TAARC equation was used; for FPWMLD = 0, the TAATL equation was used; and for 0.3 > FPWMLD > 0, TA was determined as a weighted average, as follows:

\[
TA = (FPWMLD/0.3)TAARC + \left(1 - \frac{FPWMLD}{0.3}\right)TAATL. \tag{4}
\]

Winter surface DIC concentrations (\(DIC_W\)) were estimated from the DIC concentration at the temperature minimum (\(T_{\text{min}}\)) of each station where carbon system samples were collected (100 and 300 transects only). The \(T_{\text{min}}\) is assumed to represent the convection depth of the previous winter mixed layer (Rudels et al., 1996); however, as Randelhoff et al. (2019) pointed out, the absolute \(T_{\text{min}}\) in western Baffin Bay (found at 100–150 m) likely does not represent the true depth of winter convection but is a remnant of winter convection in eastern Baffin Bay that was subsequently subducted underneath Arctic waters (as also evidenced by the low FPW values of this layer). Therefore, we follow the adapted method proposed by Randelhoff et al. (2019) to identify the winter convection depth as the shallowest local temperature minima. This method is somewhat arbitrary and leads to a range of winter mixed layer depths in western Baffin Bay from 15 to 40 m and \(DIC_W\) of 2,104.6 to 2,136.8 \(\mu\text{mol kg}^{-1}\), but it provides the best approximation that we can make with the available data.

Due to the lower number of stations where discrete DIC samples were collected, we could not simply interpolate \(DIC_W\) between stations along the ship track. Therefore, we first determined average \(DIC_W\) and \(\rho_W\) values for the Atlantic and Arctic domains, which were averaged from DIC values with FPW = 0 and FPW > 0.3 at the identified winter convection depth, respectively. Calculated average \(DIC_W\) values for the Atlantic and Arctic domains were 2,111.5 ± 11 \(\mu\text{mol kg}^{-1}\) and 2,127.2 ± 13 \(\mu\text{mol kg}^{-1}\), respectively. Then, the interpolated values of FPWMLD were used to determine \(DIC_W\) and \(\rho_W\) at all points along the ship track in the same manner as TA above. We were able to use interpolated FPWMLD values as a proxy for winter mixed layer conditions, as we found that only small differences in FPW existed between the MLD and depth of winter convection (average difference of 0.04 ± 0.03) and that these differences did not result in a change of classification between Atlantic or Arctic domains at any station.

To account for the dilution of surface water DIC by freshwater inputs, \(DIC_S\) values were salinity-normalized to the salinity of the winter mixed layer (the salinity associated with \(DIC_W\)). We performed the salinity normalization following Fris (2003), such that:

\[
nDIC_s = \frac{(DIC_s - DIC_0)}{S_s}S_w + DIC_0. \tag{5}
\]

where \(DIC_s\) is the surface DIC derived from surface \(pCO_2\) measurements, \(S_s\) is the surface salinity, \(S_w\) is the salinity of the winter mixed layer, and \(DIC_0\) is the estimated DIC concentration at a salinity of zero. \(DIC_0\) values were determined for Arctic outflow waters (\(DIC_{0,\text{ARC}} = 914 \, \mu\text{mol kg}^{-1}, \text{RMSE} = 13.0\)) and Atlantic-modified waters (\(DIC_{0,\text{ATL}} = 0 \, \mu\text{mol kg}^{-1}, \text{RMSE} = 35.1\)) from linear DIC-S relationships (see Figure S4). The appropriate \(DIC_0\) value to use at each point along the ship track was determined by FPWMLD in the same manner as TA and \(DIC_W\).

The flux correction term (\(FCO\) in Equation 3) integrates daily air-to-sea \(CO_2\) flux estimates over the DOW prior to sampling. Daily fluxes were calculated as

\[
F_{\text{daily}} = \alpha k(pCO_{2\text{sw}} - pCO_{2\text{atm}}), \tag{6}
\]

where \(k\) is the transfer velocity, estimated using the parameterization of Wanninkhof (2014) and daily wind speeds from NCEP/NCAR reanalysis data (Kalnay et al., 1996); \(\alpha\) is the solubility of \(CO_2\) in seawater (Weiss, 1974); and \(pCO_{2\text{sw}}\) and \(pCO_{2\text{atm}}\) are the partial pressures of \(CO_2\) in the surface seawater and atmosphere, respectively. \(pCO_{2\text{atm}}\) is assigned a constant value of 404 \(\mu\text{atm}\); the average value recorded at Alert, Nunavut, between June 7 and July 14, 2016 (World Data Centre for Greenhouse Gases, 2018). As we do not have measurements of surface \(pCO_{2\text{sw}}\) prior to our date of sampling, we make the assumption that surface \(pCO_{2\text{sw}}\) would not have changed significantly between the day of retreat (DOR) and our sampling date (i.e., we hold the observed \(pCO_{2\text{sw}}\) values constant over DOW). This assumption is supported by the results of previous investigations in Canadian Arctic waters (Shadwick et al., 2011; Else et al., 2012, 2019), which show that surface \(pCO_{2\text{sw}}\) generally undergoes a rapid decrease associated with an under-ice or ice-edge bloom, quickly reaching its annual minimum value before beginning a slow increase throughout the open water season. The results of Else et al. (2019) demonstrate this trend occurring within our study region, showing a rapid drop in surface \(pCO_{2\text{sw}}\) and nitrate concentrations associated with the onset of an under-ice bloom in early July.

We estimate the relative uncertainty of our calculated \(NCP_{pCO_2}\) values to be ±20% based on recalculating \(NCP_{pCO_2}\) with variations in TA, DIC\(_W\) and DIC\(_0\) within their stated limits of variability. This estimate does not take into account the uncertainty associated with the flux correction (\(FCO\)); however, the relative increase in \(NCP_{pCO_2}\) due to the correction was very small (maximum of 8%) and therefore has only a small impact on the uncertainty of \(NCP_{pCO_2}\). Additionally, the sensitivity of \(FCO\) to a different definition of DOW, using a 50% \(SIC\) threshold instead of 15% (as described in the \(SIC\) section below), was found to be very small, altering \(FCO\) by a maximum of only 0.01 mol C m\(^{-2}\) and a maximum relative change of 0.8%. This limited sensitivity is likely the case because of the relatively small geographical area occupied by the marginal ice zone in comparison to open water of heavily ice-covered areas (see Figure S5).
**NCP from nitrate drawdown**

Calculated values of NCP$_N$ (mol C m$^{-2}$) correspond to the cumulated water-column drawdown of nitrate since the onset of the growing season multiplied by the molar Redfield C: N ratio of 6.7. The nitrate drawdown is estimated from the difference between observed nitrate inventories ($N_{obs}$) at the time of sampling and a reconstruction of pre-bloom inventories ($N_p$) based on the salinity-nutrient approach introduced by Bergeron and Tremblay (2014). These authors observed that linear salinity–nutrient relationships extend through the upper 200 m during late winter in northern Baffin Bay and that pre-bloom nutrient concentrations at each depth can be estimated from the salinity values at the time of sampling, with a correction for conservative dilution when salinity is lower than the winter minimum at the surface. See Bergeron and Tremblay (2014) and the associated supporting information for details.

**Freshwater fractions from $\delta^{18}O$**

To classify low-salinity waters as being affected by sea-ice melt or meteoric waters, $\delta^{18}O$ and salinity measurements were used in a simple three end-member mass balance whereby each water sample is assumed to be a mixture of Atlantic water ($S_{AW}$ = 34.9, $\delta^{18}O_{AW}$ = 0.3%), sea-ice melt ($S_{SIM}$ = 4, $\delta^{18}O_{SIM}$ = 0.5%), and meteoric water ($S_{MW}$ = 0, $\delta^{18}O_{MW}$ = −20%). End-member values for sea-ice melt and meteoric water are based on those of Østlund and Hut (1984) and Fairbanks (1982), respectively, whereas the Atlantic water’s end-members follow Dodd et al. (2012). Here we cannot distinguish low-salinity Pacific water from the mixture of Atlantic water and meteoric waters because, as stated by Yamamoto-Kawai et al. (2005), the salinity-$\delta^{18}O$ properties of Pacific water lie on the mixing line of Atlantic water with Arctic meteoric water. Therefore, in our results, the presence of Pacific water will appear in the meteoric water signal.

**Sea-ice concentration**

Using AMSR2 SIC data (Spreen et al., 2008) on a 6.250-km grid, we identified the day of ice retreat for each sampling location. We defined DOR as the last day that SIC dropped below 15% and remained below that threshold for the rest of the season. Defining DOR in this way is a commonly used climate metric in studies characterizing variability in sea-ice extent and the length of the open water season using passive microwave measurements such as AMSR2 (Peng et al., 2018; Bliss et al., 2019). We then calculated the DOW at each location at the time of sampling by subtracting DOR from our sampling date (both expressed as year-day). This approach resulted in positive and negative values of DOW in locations that had become ice-free or remained ice-covered at the time of sampling, respectively.

**Results**

**Sea surface properties**

Figure 2 shows sea surface properties throughout the study area with all measurements displaying both spatial (east–west) and temporal variations. Figure 2A shows the east-to-west retreat of the ice edge across southern Baffin Bay (shown as DOR contours). At the beginning of the cruise (day 160), the ice edge was located approximately along the West Greenland shelf break, and by the end of the cruise (day 190), it was quickly receding toward the Baffin Island shelf break. It was not until July 28 (day 210) that the entire study area became ice-free (not shown).

Surface waters in western Baffin Bay were generally colder and fresher than those in the east (Figure 2E, F). This finding is consistent with the presence of cold, less saline Arctic outflow waters in the west, and warmer, more saline, modified Atlantic waters in the east. Henceforth, we will refer to these two water masses as “Arctic” and “Atlantic” waters, respectively. Observed patterns in sea surface temperature and salinity are also consistent with observed SIC along the ship track (Figure 2C), with ice-covered waters displaying colder temperatures. As the cruise progressed from early June into July (DOY 160–190; Figure 2B), sea surface temperatures in the east progressively became warmer due to solar insolation, while surface salinities in the west continued to decrease as sea-ice melt progressed.

Underway measurements of surface $pCO_2$ ranged from 124 to 389 µatm (Figure 2G) and also displayed an east-to-west gradient with generally lower $pCO_2$ observed in the east (approximately 124–250 µatm) and higher $pCO_2$ in the west (approximately 250–389 µatm). The range of this east-to-west gradient decreased over time mainly due to decreasing surface $pCO_2$ in the west. The first four (southernmost) transects to be sampled displayed the greatest range in $pCO_2$ measurements compared to the latter three (northernmost) transects of the cruise. All underway $pCO_2$ measurements were consistently undersaturated with respect to the atmosphere ($pCO_2_{atm}$ = 404 µatm; average recorded value at Alert, Nunavut, between June 7 and July 14, 2016 (World Data Centre for Greenhouse Gases, 2018).

Surface measurements of biological oxygen supersaturation ($\Delta O_2/Ar$) ranged from −5.6% to 25.4% across the region (Figure 2H) and generally showed a negative correlation with surface $pCO_2$ measurements ($n = 5,374, R^2 = 0.64, P < 0.001$), suggesting that biological processes largely drove $pCO_2$ variations. In the open waters of eastern Baffin Bay, $\Delta O_2/Ar$ was consistently positive (ranging between 1.2% and 25.4%) signaling that surface waters were net autotrophic. In the western ice-covered waters we observed some negative $\Delta O_2/Ar$ values with values ranging between −5.6% and 12.8%, implying that some ice-covered areas may have either been net heterotrophic at the time of sampling or impacted by upwelling.

**Water column characteristics**

Figure 3 shows water column properties over the upper 100 m of transect 100. East–west trends shown in this transect are generally representative of all seven longitudinal transects discussed in this study.

The east–west divide between Atlantic and Arctic waters was evidenced by more saline ($S > 33.5$) Atlantic...
water extending to the surface in the east, whereas lower salinity Arctic waters were present in the upper water column west of 61°W (Figure 3A). Atlantic waters were relatively warm ($T > -1$ °C) over the upper 40 m in areas that had already become ice-free, with colder winter waters below (Figure 3B). Atlantic waters generally
displayed deeper mixed layers and greater depths of winter convection compared to Arctic waters. Within the Atlantic domain, we observed a pocket of lower salinity surface waters at the ice edge (centered at 59°W), which was shown to be due to sea-ice melt (Figure 3C) and coincided with a decrease in surface-mixed layer depths. Arctic waters showed a predominant brine signal (negative $F_{\text{SIM}}$) with positive fractions of meteoric water extending below the upper 100 m (Figure 3D). This meteoric water signal was closely aligned with the location of Pacific waters (Figure 3F), suggesting that it results primarily from the presence of low-salinity Pacific waters.

The distribution of nitrate (Figure 3E) was influenced by both water mass distribution and biogeochemical processes. A strong imprint of primary production was observed in the upper water column of the easternmost stations where nitrate concentrations have been depleted to depths of 40 m. The observed fluorescence signal also demonstrated the downward migration of the surface bloom at the easternmost stations, becoming an SCM located at depths below 40 m (Figure 3G). The depth range of the SCM was found to be perched between the nitracline depth and the 0.415 mol m$^{-2}$ day$^{-2}$ isolume (see figure 9 of Randelhoff et al., 2019), balancing the need for sufficient nutrients and light. Closer to the ice edge, the fluorescence signal only indicated near-surface primary production. Surface nutrients in the west had not yet been depleted, and the fluorescence signal showed little primary production at the time of sampling. As expected, nitrate concentrations at depth were higher in Atlantic waters.
waters (approximately 10–12 μmol kg⁻¹) than in Arctic waters (approximately 8 μmol kg⁻¹).

NCP estimates

Figure 4 shows spatial trends in NCP estimated by our three different approaches. Generally, all three approaches showed higher rates of NCP in eastern Baffin Bay than in the west, with the westernmost sampling locations displaying the lowest NCP estimates. The range of values estimated by each approach varied greatly (note different scales in Figure 4). NCPp displayed the greatest range from 0 to 5.7 mol C m⁻², with NCPpCO₂ ranging from –0.7 to 3.3 mol C m⁻², and NCPO₂/Ar from –0.3 to 1.2 mol C m⁻².

Values of NCPpCO₂ and NCPN generally increased from west to east across each transect with one exception seen in transect 100 (see Figure 1 for transect identifications), where both NCPpCO₂ and NCPN decreased slightly at their easternmost sampling locations. Values of NCPO₂/Ar did not follow the same increasing trend from west to east. Instead, NCPO₂/Ar generally peaked near the West Greenland Shelf break.

Figure 5 presents a time-series view of NCP estimates with coinciding measurements of SIC and surface FPW. All NCP estimates were generally highest in open water conditions with low FPW and decreased under high SIC, which often coincided with high FPW. NCPO₂/Ar consistently showed lower values compared to other approaches, except under high SIC, where NCPN and NCPp displayed good agreement with NCPpCO₂. Throughout the first three transects of the cruise (100–300), NCPpCO₂ values dramatically changed at the ice edge, increasing from lower values in close agreement with NCPO₂/Ar under ice to higher values nearing NCPN estimates in ice-free waters. During the later transects (400–700), NCPpCO₂ estimates in open water remained well below NCPN values. Throughout the cruise, NCPpCO₂ estimates displayed a strong inverse relationship to FPW. Along transects 600 and 700, under lower overall SIC conditions, NCPpCO₂ values also demonstrated an inverse relationship with FPW.

Discussion

Comparison of estimates from various NCP approaches

As shown in Figures 4 and 5, the ranges of the NCP estimates from each approach differed, which makes sense when considering the different integration timescales and depths of each approach (see Table 1). Of our three approaches, NCPO₂/Ar integrates over both the shortest timescale and depth, meaning that the O₂/Ar approach is representative of NCP in surface waters over a relatively short window (approximately 20 days in open water; longer under ice) prior to the time of sampling. NCPO₂/Ar displayed its highest values in open water in proximity to the receding ice edge (Figures 4 and 5), capturing the signal of the transient ice-edge bloom as it progressed westward. In many transects, NCPO₂/Ar showed lower rates eastward of the ice edge (into higher DOW), likely signaling that the surface bloom was declining after using up the available stock of nutrients in the surface-mixed layer. NCPpCO₂ and NCPN represent seasonally integrated NCP estimates throughout the surface-mixed layer and total euphotic zone, respectively. By "seasonally integrated," we are referring to the fact that the residence times of these tracers (nitrate and DIC) are long enough to capture all NCP that has occurred since the end of winter. Therefore, we would expect both of these NCP estimates to continually increase as the bloom continues, with NCPN being greater than NCPpCO₂ in areas where net primary production is also occurring below the surface-mixed layer. For the most part, NCPpCO₂ and NCPN increase from west to east, following the trend of increasing DOW, with one exception in transect 100 where both these approaches show lower NCP estimates at the easternmost sampling station. Figure 3 also reveals a shallowing of the nitracline at the easternmost station of transect 100, along with increasing surface temperatures and a weaker fluorescence signal throughout the surface-mixed layer. This pattern is perhaps signaling a transition to a different water mass regime over the West Greenland continental shelf.
Controls on the timing and magnitude of NCP
As positive rates of NCP (net autotrophy) cannot occur in the absence of net primary production, it follows that the timing and succession of the spring phytoplankton bloom will play a major role in determining the observed temporal and spatial trends in NCP. To gain insight into the impacts of bloom timing and succession, we can exploit the different integration timescales and depths of our NCP approaches. As we have established, $NCP_{O_{2}/Ar}$ estimates are representative of production over a relatively short window prior to sampling (approximately 20 days in open water; greater under ice) within the surface-mixed layer, whereas $NCP_{pCO_{2}}$ and $NCP_{N}$ represent seasonal NCP integrated over the surface-mixed layer and total euphotic zone, respectively. Based on these distinctions, if $NCP_{O_{2}/Ar} \approx NCP_{pCO_{2}}$, we can assume net autotrophic conditions began roughly around the time of sampling, whereas if $NCP_{pCO_{2}} > NCP_{O_{2}/Ar}$, net autotrophy had already begun prior to the time of sampling. Both the $NCP_{O_{2}/Ar}$ and $NCP_{pCO_{2}}$ approaches integrate over the surface-mixed layer and therefore will not capture any deeper production by the SCM. However, the $NCP_{N}$ approach does capture the SCM by integrating over the total euphotic zone. Therefore, comparing $NCP_{pCO_{2}}$ and $NCP_{N}$ estimates can provide insight into how much of the total NCP has occurred within the surface-mixed layer or deeper in the water column. If $NCP_{pCO_{2}} \approx NCP_{N}$, then all NCP has occurred within the surface-mixed layer, whereas if $NCP_{N} > NCP_{pCO_{2}}$, then the portion of NCP occurring below the surface-mixed layer can be approximated by $1 - NCP_{pCO_{2}}/NCP_{N}$.

Figure 6 directly compares the $NCP_{O_{2}/Ar}$ estimates with those from seasonally integrated approaches. Figure 6A shows a strong inverse relationship between DOW and the slope of the $NCP_{O_{2}/Ar}/NCP_{pCO_{2}}$. Most $NCP_{O_{2}/Ar}$ values that fall along the 1:1 line with $NCP_{pCO_{2}}$ are associated with negative DOW, indicating that net autotrophic conditions were initiated by an under-ice bloom. Based on the range of DOW values near the 1:1 line, net autotrophy began anywhere from 2 to 15 days prior to ice retreat (not considering the positions that remained net heterotrophic, i.e., NCP < 0). We observed some diverging bands of data points with negative DOW and $NCP_{pCO_{2}} > NCP_{O_{2}/Ar}$, which we hypothesize may be due to the coarse resolution of the AMSR2 SIC data that were used to calculate DOW. We find a sharper relationship between in situ sea surface temperature and the slope of $NCP_{O_{2}/Ar}/NCP_{pCO_{2}}$ (Figure S6), suggesting that gridded satellite data are unable to capture small-scale decreases in sea-ice cover that would lead to warmer sea surface temperatures by solar insolation. The coarse resolution of our sea-ice concentration data would also impact our estimates of air-sea gas exchange in partially ice-covered waters, which may lead to slight underestimates in both $NCP_{O_{2}/Ar}$ and $NCP_{pCO_{2}}$ within the marginal ice zone.

Under open water conditions (DOW > 0), $NCP_{pCO_{2}}$ continued to increase, while $NCP_{O_{2}/Ar}$ either remained approximately constant or decreased. This pattern is consistent with the short-lived ice-edge bloom commonly observed in satellite chlorophyll $a$ observations, which quickly depletes the nutrient availability of the shallow surface layer stratified by sea-ice meltwater (Perrette et al., 2011). High $NCP_{O_{2}/Ar}$ values coinciding with open water conditions were mainly observed at earlier sampling dates (year-day 160–175; transects 100–300) when the ice edge remained in the center of our study region, whereas locations sampled later in the season (year-day 185–190)
displayed $NCP_{O_{2}/Ar}$ values near zero in open water (Figure S7) likely due to post-bloom nutrient-depleted surface waters.

As DOW increased, NCP$_N$ estimates generally became increasingly greater than $NCP_{pCO_2}$, indicating the continuation of primary production below the surface-mixed layer in SCM. During the first three transects of the cruise (100–300), NCP$_N$ values were on average 1.3 mol C m$^{-2}$ greater than $NCP_{pCO_2}$ ($\sigma = 0.7$), and in later transects, NCP$_N$ values averaged 1.6 mol C m$^{-2}$ greater than $NCP_{pCO_2}$ ($\sigma = 1.1$). These differences are consistent with the pattern of bloom succession described in the literature (Sakshaug, 2004; Martin et al., 2010). The fluorescence signal in Figure 3G also showed this progression from surface bloom to SCM in Atlantic waters, coinciding with a deepening nitracline. Based on the differences between NCP$_N$ and $NCP_{pCO_2}$ estimates in Atlantic waters, we estimate that SCMs may have contributed up to 70% of total NCP in the latter half of our study period (transects 400–700; see Figure S8).

Figure 6B shows a scatterplot of our NCP estimates, with data points colored according to the FPW, demonstrating that seasonal NCP was consistently lower in the Arctic waters in western Baffin Bay than in the Atlantic waters in eastern Baffin Bay. All seasonal NCP estimates in locations with FPW $> 0.2$ were lower than 1 mol C m$^{-2}$, with NCP$_N$ estimates not differing greatly from $NCP_{pCO_2}$ (average difference of 0.3 mol C m$^{-2}$, $\sigma = 0.2$). These results suggest that the majority of NCP that had occurred in this region was located near the surface. As most sampling locations in Arctic outflow waters remained ice-covered at the time of sampling, given more time NCP would likely have increased. However, the water column characteristics observed in western Baffin Bay suggest that future primary production will likely be limited compared to that in eastern Baffin Bay. Arctic waters displayed higher freshwater content over the upper 80 m of the water column (Figure 3A) and displayed lower pre-bloom nitrate concentrations in comparison to Atlantic waters (Figure 3E; Randelhoff et al., 2019). Nitrate concentrations were especially low (<3 µmol kg$^{-1}$) in Arctic waters over the upper 20–30 m of the water column, with strong stratification inhibiting the replenishment of nutrients from below. Based on these characteristics, seasonal NCP rates in Arctic waters will likely remain below those observed in the east.

Comparison with previous studies

A number of previous studies utilizing multiyear observations in northern Baffin Bay have revealed declining trends in primary production and/or NCP (Bélanger et al., 2013; Bergeron and Tremblay, 2014; Blais et al., 2017; Marchese et al., 2017). Bergeron and Tremblay (2014) and Blais et al. (2017) reported declining rates of NCP and phytoplankton biomass, respectively, to coincide with decreasing salinity in the upper water column. Both investigations also noted that the strongest declining trends occurred at western stations in Arctic outflow waters. These reported trends align well with our observations of low NCP associated with Arctic outflow waters in western Baffin Bay. Using a satellite-based model to assess primary production trends from 1998 to 2010, Bélanger et al. (2013) also found decreasing trends in primary production along major exit routes of Arctic waters, such as the outflow shelves of East Greenland, the CAA, and northwestern Baffin Bay.

A limited number of studies have previously published estimates of NCP in Baffin Bay. The synthesis study of Codispoti et al. (2013) reported average NCP values over large subregions of the Arctic Ocean, utilizing nutrient data collected between 1970 and 2007. These authors found a higher average NCP value of 32 g C m$^{-2}$ (2.7 mol C m$^{-2}$) for their “Canadian Archipelago” region, which included western Baffin Bay and the channels of the CAA compared to 10 g C m$^{-2}$ (0.8 mol C m$^{-2}$) over their “Greenland Shelf” region, which encompassed both the East and West Greenland shelves. This result is opposite to the east–west NCP gradient we report in Baffin Bay.
Codispoti et al. (2013) found surface waters in the CAA to have higher pre-bloom nitrate concentrations (approximately 10 µM) compared to those over the Greenland shelves (<5 µM) and reported lower surface salinities and stronger stratification over the Greenland shelves. We again find the opposite trend, with our results in western Baffin Bay displaying lower pre-bloom nitrate concentrations and stronger stratification (Figure 3). This disparity of results is likely a manifestation of comparing multi-annual averaged values over immense areas, with those calculated over smaller spatial and temporal scales (this study), making meaningful comparisons between these two studies difficult. However, interannual changes in the freshwater inventory of both the BIC and WGC systems may have also contributed somewhat to the observed disparity.

We find better agreement between our NCPN estimates in western Baffin Bay and those reported by Bergeron and Tremblay (2014) near the end of their roughly decadal study (1997–2011) in northern Baffin Bay. These authors reported a minimum annual NCP value of 214 mmol NO3

m–2 (or 1.42 mol C m–2) from 2009. Our NCPN values, calculated using the same approach, produce a mean value of 0.4 mol C m–2 (σ = 0.3) in Arctic waters (surface FPW > 0.2), which is still significantly lower than the 2009 value of Bergeron and Tremblay (2014), but this difference may be expected considering the earlier sampling dates of our study and the declining NCP trend reported by Bergeron and Tremblay (2014).

Finally, the study of Tremblay et al. (2006) reported NCP estimates in the eastern North Water Polynya region during June and July 1998 (comparable with our Atlantic domain in eastern Baffin Bay). These authors calculated seasonal NCP rates based on nitrate and DIC drawdown over the upper 150 m of the water column and reported values of approximately 9.5 mol C m–2 in June and approximately 10.5 mol C m–2 in July. These values are much higher than our maximum NCP rate in eastern Baffin Bay of 5.7 mol C m–2; however, given the declining NCP and primary production trends over the past decade within the North Water region (Bélangier et al., 2013; Bergeron and Tremblay, 2014; Blais et al., 2017; Marchese et al., 2017), such high rates of NCP are no longer occurring within that region. Our reported seasonal NCP rates within the Atlantic sector of Baffin Bay (1.4–5.7 mol C m–2) fall in the same approximate range as those reported more recently in the productive Barents Sea (2.3–4.2 mol C m–2; Fransson et al., 2001; Codispoti et al., 2013; Chierici et al., 2019).

**Caveats of inferring pre-bloom conditions from the temperature minimum**

Due to the scarcity of winter measurements within the Arctic region, former studies of nutrient or DIC drawdown have relied upon winter surface concentrations inferred from summer profiles (Ulfsbo et al., 2014; Randelhoff et al., 2019). Following Rudels et al. (1996), the temperature minimum located below the summer mixed layer is generally assumed to indicate the depth of winter convection during the previous winter, and therefore, nutrient concentrations at the temperature minimum are representative of the previous winter’s surface-mixed layer. However, depending on the water mass assembly of certain Arctic Seas, accurately identifying the depth of winter convection using the temperature minimum can be very challenging. We found this to be the case in western Baffin Bay (Figure 3), where the upper 100–150 m of the water column remains near the freezing point temperature with several temperature minima, suggesting advection from neighboring areas with slightly different convection depths and salinities. Even in eastern Baffin Bay, where the location of the temperature minimum layer is more easily identifiable, linearly interpolating between sparse sample depths may lead to significant uncertainty in defining winter surface water nutrient concentrations.

Using our nitrate measurements, we performed a comparison of NCPN estimates using two different approaches: (1) the salinity–nutrient relationship approach of Bergeron and Tremblay (2014), which is the same as the NCPN method used in this work, but will be called the “BT14” method in this section and (2) the approach of using temperature minimum (Tmin) characteristics of each station to represent surface water conditions of the previous winter. The Tmin method integrates nitrate drawdown over the winter MLW (all depths above the Tmin) and assumes that winter surface nitrate concentrations remained constant throughout this layer unless diluted by freshwater inputs (see Text S1 for details on the Tmin method). Comparing these two methods makes immediately apparent that the BT14 method produced much greater NCP estimates (up to a factor of 2) than the Tmin approach (see Figure 7A). In many cases, NCP values from the BT14 approach were double of those produced using Tmin, characteristics. The greatest discrepancies were found at sampling locations in eastern Baffin Bay, suggesting that biological nutrient drawdown may have already occurred below the Tmin depth at these locations. This suggestion is also supported by the pre-bloom (or winter) surface nitrate concentrations predicted by each method (Figure 7B), with the Tmin approach underestimating pre-bloom surface nitrate concentrations in eastern Baffin Bay compared to the BT14 estimates. Additionally, fluorescence measurements clearly showed the downward migration of the bloom in eastern Baffin Bay to the Tmin depth (see Figure 3G). It follows that the Tmin approach may lead to underestimations of NCP and/or new production in situations where nutrient drawdown occurs at or below the Tmin depth.

Our NCPNO3 approach also relies on Tmin characteristics to define pre-bloom DIC concentrations in the Atlantic and Arctic domains of Baffin Bay. Interestingly, our NCPNO3 estimates agree well with the BT14 method integrated over the surface-mixed layer (see Figure S9). However, this good agreement may be largely due to the relatively shallow integration depth. We expect that if our NCPNO3 approach were applied over a greater depth range (e.g., to the depth of winter convection estimated by Tmin), it would also underestimate the true seasonal NCP.
Implications for marine ecosystems and carbon cycling

Our results highlight a strong contrast in spring NCP between the Arctic and Atlantic water domains of Baffin Bay. Arctic outflow waters displayed relatively low spring NCP (<1 mol C m⁻²) associated with persistent sea-ice cover, high freshwater content, and strong stratification of the upper water column. Meanwhile, modified Atlantic waters in eastern Baffin Bay were largely ice-free, with comparatively deep surface-mixed layers, and displayed relatively high NCP (up to 5.7 mol C m⁻²).

The low NCP rates of Arctic outflow waters will also have secondary effects on the marine ecosystem and the cycling of carbon in this region. Lower spring NCP decreases food export to higher trophic levels, with potentially adverse consequences for pelagic and benthic organisms. Blais et al. (2017) also observed decreasing phytoplankton cell sizes associated with Arctic outflow waters in northern Baffin Bay. Smaller cell sizes have lower sinking velocities and are therefore more likely to be recycled in the upper water column, feeding the microbial food web instead of passing organic material to the pelagic and/or benthic food webs (Legendre and Rassoulzadegan, 1995). Assuming this same trend is affecting all Arctic outflow waters in western Baffin Bay, we are likely to observe decreasing biodiversity of pelagic and benthic organisms in this region.

Low NCP also has implications for the marine carbon cycle in western Baffin Bay, limiting the ability of these waters to act as an atmospheric CO₂ sink (Azetsu-Scott et al., 2010) have already shown that increasingly fresh Arctic outflow waters are prone to low seawater pH and calcium carbonate saturation states. On the other hand, high rates of NCP in eastern Baffin Bay will maintain the ability of this region to act as an atmospheric CO₂ sink throughout the productive season, and the associated export of organic material to depth will sustain populations of other pelagic and benthic organisms.

Conclusions and future recommendations

Based on our NCPₚₙ estimates, we found an average spring NCP of 3.7 and 0.4 mol C m⁻² in the Atlantic and Arctic water domains of Baffin Bay, respectively. These results illustrate a strong dichotomy in the potential strength of the biological pump between eastern and western Baffin Bay. Low NCP in western Baffin Bay was likely caused by the high freshwater content of Arctic outflow waters and resultant strong stratification, which simultaneously decreased surface nitrate concentrations and restricted nutrient replenishment from below. These low rates of NCP may have adverse consequences for the pelagic and benthic food webs in this region and decrease the ability of these waters to act as an atmospheric CO₂ sink. With the continuation of increasing glacial meltwater runoff into Baffin Bay (Dahl-Jensen et al., 2011; Sharp et al., 2011; Bamber et al., 2018), this trend will likely continue and perhaps also begin to strengthen surface stratification in eastern Baffin Bay (Castro de la Guardia et al., 2015), potentially also decreasing NCP rates on the West Greenland shelf in the future.

By comparing multiple NCP approaches with different integration depths and timescales, we were able to track the spatial and temporal evolution of the spring bloom in southern Baffin Bay. We found the surface bloom was
initiated under ice up to 15 days prior to ice retreat. Once the bloom progressed into open water, it quickly depleted surface nutrients and progressed deeper into the water column becoming an SCM. We estimate that SCMs may have contributed up to 70% of the seasonal NCP within the Atlantic domain further establishing the importance of these features to the marine carbon cycle in Arctic Seas. We have also demonstrated that nutrient drawdown in the depth range of the SCM may coincide with, or surpass, the depth of the temperature minimum, a feature that has generally been considered a remnant of winter surface conditions. We hope this finding will serve to caution future investigations that conditions at the temperature minimum may not accurately represent surface conditions of the previous winter. Periodic sample collection in late winter would greatly improve the accuracy of seasonal NCP estimates.

Data accessibility statement
All data are available through the Green Edge database (http://www.obs-vlfr.fr/proof/php/GREENEDGE/greenedge.php) and will be made available prior to publication. Additionally, surface underway pCO₂ data are available through the Surface Ocean CO₂ Atlas at https://www.socat.info, and discrete bottle measurements are available through the Ocean Carbon Data System at https://www.nodc.noaa.gov/ocads/

Supplemental files
The supplemental files for this article can be found as follows:
All supplemental materials for this article can be found in one file, and all references to the supplemental material are prefixed with an “S.”
Text S1. Figures S1–S9. docx

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The authors have no competing interests to declare. JET is Elementa editor and he was not involved in the review of this manuscript.

Author contributions
Contributed to conception and design: TB, JET, and TP.
Contributed to acquisition of data: Author and all coauthors.
Contributed to analysis and interpretation of data: TMB and JET.
Drafted and/or revised the article: Author and all coauthors.
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