Iron Supply by Subglacial Discharge Into a Fjord Near the Front of a Marine-Terminating Glacier in Northwestern Greenland

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Abstract Glacial fjords in Greenland show high productivity owing to the runoff of meltwater from the glaciers. Macronutrient dynamics (of nitrate, phosphate, and silicate) associated with subglacial discharge plumes in front of marine-terminating glaciers are widely cited as important drivers of summer phytoplankton blooms in the fjords. However, the dynamics of iron (Fe), an essential micronutrient for primary production, remain largely unstudied in glacial fjords. To investigate the role of subglacial discharge plumes in Fe supply processes in glacial fjords, a comprehensive survey of Bowdoin Fjord, adjacent to the marine-terminating Bowdoin Glacier in northwestern Greenland, was conducted. The subglacial discharge of Fe-rich meltwater induces a buoyancy-driven upwelling plume in front of the glacier that entrains nutrient-rich Arctic- and Atlantic-origin waters. The plume water potentially carried $4.5-8.7 \times 10^7$ g day$^{-1}$ of total dissolved Fe out of Bowdoin Fjord in summer. The concentration of dissolved Fe (dFe) in the plume water ($\sim 15.6$ nmol kg$^{-1}$) was 4 times higher than that in the water in the outer part of the fjord ($\sim 3.8$ nmol kg$^{-1}$). The dFe:nitrate + nitrite ratio (nmol mol$^{-1}$) in the plume water varied between 0.58 and 3.2, several orders of magnitude higher than the phytoplankton cellular Fe:nitrate ratio estimated using the hypothetical Fe:C ratio and observed particulate organic carbon:nitrate ratio of the fjord. Hence, the plume water is replete with Fe with respect to phytoplankton demands. Subglacial discharge drives the upwelling of Fe and macronutrients toward the euphotic zone, which is vital for the generation of summer phytoplankton growth in glacial fjords.

1. Introduction

Glacial fjords in Greenland are characterized by high productivity, represented by a high phytoplankton biomass as well as a high density of fish, seabirds, and marine mammals (Laidre et al., 2016; Meire et al., 2017; Nishizawa et al., 2019). The production of phytoplankton biomass within fjords shows seasonal patterns with a primary peak in spring, followed by a secondary peak in summer (Juul-Pedersen et al., 2015). The increase in production in summer (summer phytoplankton bloom) has been attributed to meltwater runoff from glaciers, evidenced by the distribution of meltwater correlating with satellite-derived surface chlorophyll a (Chl. a) concentration (Arrigo et al., 2017; Oliver et al., 2018). Therefore, meltwater runoff plays an important role in the summer phytoplankton bloom initiation.

Glacial meltwater runoff can either dilute or enhance the concentrations of biogeochemical components within the surface layers of fjords. In Greenland, runoff flows into the ocean through two main pathways, namely, the proglacial streams of land-terminating glaciers and the subglacial discharge from marine-terminating glaciers. Meltwater generated on the surface of land-terminating glaciers drains into proglacial streams and enters the ocean surface (Chu et al., 2012; Overeem et al., 2017). On marine-terminating glaciers, surface meltwater drains into a subglacial drainage network through crevasses and englacial channels. The subglacial meltwater then discharges into the ocean at depths of several hundred meters, upwelling along the calving front to form a buoyant plume (Mankoff et al., 2016; Motyka et al., 2013). Recent studies have indicated that glacial meltwaters are not a significant source of macronutrients, especially nitrate, thereby likely causing a dilution of nitrate in the surface waters of glacial fjords (Hopwood et al., 2018;
Kanna et al., 2018; Meire et al., 2016). Yet, upwelling of a subglacial discharge-driven plume plays a crucial role in the macronutrient supply to surface waters by entraining what is often nitrate-enriched deep water into the plume (Cape et al., 2019; Hopwood et al., 2018; Kanna et al., 2018; Meire et al., 2017). This upwelling mechanism can sustain high productivity by enhancing vertical fluxes of macronutrient throughout the summer in fjords (Hopwood et al., 2020; Kanna et al., 2018; Meire et al., 2017).

Currently, increasing ice mass loss from the Greenland ice sheet and peripheral glaciers (e.g., Bamber et al., 2018; Zemp et al., 2019) has raised interest in the Fe supply via meltwater runoff. In the ocean, Fe is an essential micronutrient for the growth of phytoplankton. Fe is associated with phytoplankton metabolism with respect to photosynthesis, chlorophyll synthesis, and macronutrient utilization, which ultimately control carbon fixation from the atmosphere into the ocean (Morel & Price, 2003). As the Fe concentration in glacial meltwaters is much higher than in the ocean, the meltwater runoff is thought to be a significant source of Fe supply into the ocean (Aciego et al., 2015; Bhatia et al., 2013; Hawkins et al., 2014; Hopwood et al., 2018). However, significant amounts of glacier-sourced Fe are removed by flocculation and precipitation upon mixing with seawater, a process commonly observed in estuaries (Hodson et al., 2017; Hopwood et al., 2016; Schroth et al., 2014; Zhang et al., 2015). Therefore, Fe limitation for phytoplankton growth is generally expected only in the outer oceans during summer, such as in the North Atlantic at high latitudes (e.g., Achterberg et al., 2018), and not in fjords.

Fe supply processes are better understood for land-terminating glaciers in Greenland because investigations of stream discharge are logistically easier. For example, approximately 21.5 Gg yr$^{-1}$ of total dissolved Fe is delivered to the ocean surface by meltwater runoff from entire land-terminating parts of the Greenland ice sheet (Stevenson et al., 2017). Studies of the Fe supply process from marine-terminating glaciers are sparse; hence, details remain unknown about the role of subglacial discharge plumes on Fe transport and summer phytoplankton growth. Further, to accurately understand the supply of Fe into fjords, investigation is required for the removal of Fe occurring through mixing of subglacial discharge with ocean water in upwelling plumes.

This study presents the dynamics of Fe in a fjord near the calving front of Bowdoin Glacier, a marine-terminating glacier in northwestern Greenland. Based on data on Fe in glacial meltwater and seawater of the glacial fjord, the mechanism of Fe supply and the subsequent losses incurred by mixing meltwater with seawater in upwelling plumes was depicted. Other biogeochemical components (macronutrients, Chl. a, and particulate organic carbon [POC] and nitrogen [PON]) were also analyzed in the fjord to assess the potential influence of Fe in plumes on phytoplankton growth.

2. Materials and Methods

2.1. Study Area

Bowdoin Glacier (Kangerluarsuup Sermia in Greenlandic; Bjørk et al., 2015) is a 3-km-wide marine-terminating outlet glacier located 30 km northeast of Qaanaaq in northwestern Greenland (Figure 1). The glacier flows into the 20-km-long Bowdoin Fjord, which connects to a 100-km-long fjord system called Inglefield Bredning (Figure 1). At the calving front of Bowdoin Glacier, approximately 250-m-thick ice is grounded and about 90% of the ice is situated below sea level (Sugiyama et al., 2015). In summer, Bowdoin Glacier discharges turbid subglacial water into the fjord, forming a buoyant plume near the calving front (Jouvet et al., 2018). The impact of the subglacial discharge on the marine ecosystem in Bowdoin Fjord has been reported recently by Kanna et al. (2018), Naito et al. (2019), and Nishizawa et al. (2019).

2.2. Sampling Around the Glacier and in the Fjord

A trace metal clean method was used in the sampling of Fe in glacial meltwater and seawater. The 125-ml low-density polyethylene (LDPE) bottles (Thermo Fisher Scientific), 0.2-μm filters (Milllex-LG, Merck Millipore or AcroPak, Pall), and Teflon-coated 2.5-L Niskin-X samplers (General Oceanics) were thoroughly acid-cleaned in advance in a class-100 laminar flow hood in a clean-air laboratory following a previously reported method (Cutter et al., 2010; Kanna et al., 2014). Supraglacial water (Supraglacial B) and stream waters along the eastern margins of Bowdoin Glacier (Stream B; Figure S1 in the supporting information) and the land-terminating Tugto Glacier (Stream T; Figure S1) were collected during field campaigns in July 2016, 2017, and 2019 (Figure 1). Sampling was also conducted for the stream water fed into the
The glacial meltwater samples for total dissolvable Fe (TdFe) were directly collected in LDPE bottles by personnel wearing trace-metal-free gloves. The samples immediately passed through a 0.2-μm Millex syringe filter were used for dissolved Fe (dFe) analysis. Seawater samples were collected by the Niskin-X samplers suspended on a Kevlar rope in Bowdoin Fjord on 24 July and 1 August 2017, using a 5-m-long motor boat (Figure 1). Teflon messengers were used to close the samplers at the desired depths. Seawaters for the TdFe analyses were transferred to LDPE bottles using an acid-cleaned silicon tube and a polycarbonate filling bell to avoid contamination by airborne particles. Seawaters for the dFe analyses were immediately filtered through a 0.2-μm Millex syringe filter on board, except for the station located 1 km from the glacier, where a 0.2-μm Acropak capsule filter directly attached to spigots of the Niskin-X samplers was used. The glacial meltwater and seawater samples for the Fe analyses were transported to Japan and acidified to pH < 1.8 by adding ultrapure HNO₃ and HCl (Tamapure AA-100 or −10, Tama Chemicals), respectively, in a laminar flow clean-air bench in a clean-air laboratory. The acidified samples were stored at ambient temperature more than 2 months before analysis to detach all of the Fe from the walls of LDPE bottles and leach Fe out of the labile fraction.

Sample collection procedures for the measurements of macronutrients, POC and PON, and Chl. a in the glacial meltwater and seawater are detailed in Kanna et al. (2018). Briefly, samples for the macronutrient analyses were collected in 10-ml acrylic tubes without filtration and HgCl₂ was added to the samples to avoid biological activity (Kattner, 1999). Samples for the POC and PON analyses were filtered onto precombusted (450°C for 5 hr) GF/F filters (Whatman). Samples for the Chl. a analyses were filtered onto GF/F filters.
under a gentle vacuum (<0.013 MPa). The Chl. a on the filter was then extracted in N,N-dimethylformamide (FUJIFILM Wako Pure Chemical) for 24 hr in the dark (Suzuki & Ishimaru, 1990). During the field campaigns on the glacier, these processed samples were preserved with ice in an insulated container for less than 2 weeks because of logistical limitations. The samples were immediately frozen at −30°C in Qaanaaq, transported to Japan without melting, and stored at −20°C until their analyses. The degradation of these samples between collection and analysis was not likely to occur (see Kanna et al., 2018, for details). Samples for particle size and salinity analyses were collected into 100-ml polypropylene bottles and 250-ml glass bottles, respectively, and transported to Japan. The samples for particle size were stored at 4°C until analysis.

### 2.3. Hydrographic Observation in Fjord

During the oceanographic fieldwork, a conductivity-temperature-depth (CTD) profiler equipped with turbidity and dissolved oxygen (DO) sensors (ASTD 102, JFE Advantech) was lowered into the fjord to measure water properties at approximately 0.1-m intervals. Accuracies in conductivity, temperature, depth, turbidity, and DO measurements were ±0.01 mS cm⁻¹, ±0.01°C, ±1.8 m, ±0.3 FTU, and ± 0.4 mg L⁻¹, respectively.

#### 2.4. Sample Analyses

Acidiﬁed TdFe and dFe samples in the glacial meltwater were directly analyzed with a graphite furnace and an atomic absorption spectrophotometer (GFAAS) (Z-2700, Hitachi High Technologies). For the seawater samples, acidiﬁed TdFe and dFe were ﬁrst preconcentrated with a manual solid-phase extraction system equipped with a column of Nobias Chelate-PA1 resin (Hitachi High Technologies), modiﬁed after Sohrin et al. (2008) and Evans and Nishioka (2018). Only ultrapure grade reagents (Optima Acids, Thermo Fisher Scientiﬁc or Tamapure AA-10, Tama Chemicals) were used for the preconcentration and extraction steps. The seawater samples were adjusted to pH 6.0 ± 0.1 by adding 3.6 M of HAcO–NH₄AcO buffer solution prepared from acetic acid and ammonia solution. The Fe concentrated onto the resin was eluted with 1 M HNO₃ and determined with the GFAAS. Following the preconcentration procedures, the procedure blank was evaluated by using 18.2 MΩ ultrapure water. The procedure blank and the corresponding detection limit (3 times the standard deviation [3sd] of the procedure blank, n = 8) were 0.07 ± 0.07 and 0.2 nmol kg⁻¹, respectively. A recovery test was performed by spiking ultrapure water with inorganic Fe of 2.76 nmol kg⁻¹, which showed a recovery of 94 ± 4% (n = 3). Furthermore, a certiﬁed reference material for the trace metals, NASS-6 (National Research Council Canada) containing 8.65 ± 0.4 nmol kg⁻¹ of Fe was used to validate the preconcentration process, which showed an analytical value of 8.78 ± 0.3 nmol kg⁻¹ (n = 3). The recovery test and reference material results indicated that the procedures were suitable for measurements of Fe in the seawater samples.

Macronutrients were measured using an autoanalyzer (QuAAtro, BL TEC) with a continuous ﬂow system. The measurements were calibrated with reference seawater materials (KANSO Technos). The detection limits (3sd of low-nutrient-seawater, n = 10) were 0.01 μmol kg⁻¹ for nitrate + nitrite, 0.04 μmol kg⁻¹ for phosphate, and 0.13 μmol kg⁻¹ for silicate. POC and PON were measured using an elemental analyzer (NA-1500, Fison). The standard deviations of POC and PON obtained by repeated measurements (n = 3) of a standard material (acetanilide, C₈H₈NO) were 0.3 and 0.1 μmol, respectively. Chl. a concentrations were determined with a fluorometer (10-AU, Turner Designs) using the nonacidification method described in Welshmeyer (1994). Salinity measurement was conducted with a salinometer (AUTOSAL 8400B, Guildline Instruments), which was calibrated with IAPSO reference seawater materials (Ocean Scientiﬁc International). The measured salinity in grab samples was used for calibration of a CTD proﬁler. Particle sizes (mean equivalent spherical diameters, ESD) of suspended particles were determined within a range of 0.7–18 μm with a Coulter Counter (Coulter Multisizer III, Beckman). Results of the water sample analyses are summarized in Table S1.

### 3. Results

#### 3.1. Hydrography in Bowdoin Fjord

Potential temperature (Θ) and S measurements in seawater showed three distinct layers of fresh and warm surface waters, Arctic-origin Polar water, and Atlantic-origin deep water, as previously reported in the study region (Kanna et al., 2018; Porter et al., 2014; Willis et al., 2018) (Figures 2a and 2b). The fresh and warm surface water (up to around 5°C) was distributed in the layer shallower than 50 m, which was underlain
by a colder layer (< −1°C) of Arctic-origin Polar water at depths ranging between 50 and 100 m. Relatively warm (ranging from 0°C to 1°C) Atlantic-origin deep water was distributed below 300 m. Turbidity and DO concentration data showed clear variations from the glacier to Inglefield Bredning (>20 km from the glacier) (Figures 2c and 2d). The most turbid water was found at a depth of about 10 m within 4 km from the glacier (Figure 2c). Turbidity at this depth decreased with distance from the glacier. The DO concentration near the glacier was low (approximately 270 μmol L⁻¹ at 10–50 m (Figure 2d). At greater distances from the glacier, seawater in the same range showed higher DO concentrations (400–450 μmol L⁻¹) than those at other depths.

The θ and DO versus S diagrams revealed that seawater in the deeper layer (>50 m) had similar values of θ, S, and DO regardless of the distance from the glacier (Figures 3a and 3b). On the other hand, seawater properties above 50 m showed spatial variations, that is, water at 1 km from the glacier was colder, saltier, and had lower DO than the waters 26 km from the glacier. Most likely, the difference in water properties among the observed samples is attributable to the magnitude of mixing of seawater with meltwater supplied by Bowdoin Glacier. To estimate the meltwater fraction in seawater, water mass analysis was conducted in accordance with previous studies (Gade, 1979; Mankoff et al., 2016; Mortensen et al., 2013; Straneo et al., 2012). Briefly, the observed θ and S of the samples were compared to those expected in the fjord seawater after mixing with meltwater supplied by the glacier. On a θ-S diagram, two mixing lines were overlain: mixing with subglacial discharge (SD line), and mixing with submarine meltwater (SM line) (Figure 3a). Here, we assumed that seawater at a depth of 200 m in the fjord was entrained and mixed with subglacial discharge and submarine meltwater. The water at a depth of 200 m at 1 km from the glacier front was chosen as the entrained seawater because the subglacial discharge was expected at depth of 200 m (Sugiyama et al., 2015) and our CTD profile was available. Using the θ and S data of these three water masses, the volumetric water mass fraction of the subglacial discharge (f_sg, Figure 3a) was computed by solving the mass, θ, and S balance equations for the fractions of the three water components. Results from the water mass analysis showed that the cold, saline, and low-DO water above 50 m found at 1 km from the glacier contained up to approximately 5% of subglacial discharge (see f_sg in Figure 3a and the supporting information).
3.2. Distributions of Fe, Macronutrients, POC, PON, Chl. a, and Particle Sizes in Seawater

At a depth of 10 m, the TdFe concentration 1 km from the glacier (3,501 nmol kg$^{-1}$) was 2 orders of magnitude larger than that 20 km from the glacier (17 nmol kg$^{-1}$) (Figure 4a). The TdFe concentration showed clear variations with depth within Bowdoin Fjord (1–15 km). The concentration decreased gradually from the surface to the deeper layer and approached the value obtained in Inglefield Bredning (20 km) at a depth of 200 m. The dFe concentration above 100 m was higher at the sites within Bowdoin Fjord (approximately 16 nmol kg$^{-1}$) than in Inglefield Bredning (approximately 4 nmol kg$^{-1}$) regardless of the distance from the glacier (Figure 4b).

Macronutrients also showed higher concentrations near the glacier (Figure 5). For example, the nitrate + nitrite concentration near the glacier was significantly higher than that within Inglefield Bredning above 50 m (Figure 5a). Within this layer, the concentration 1 km from the glacier showed a steep
vertical gradient from the surface (<0.1 μmol kg\(^{-1}\)) to 20 m (11.5 μmol kg\(^{-1}\)) (Figure 5a), whereas the one within Inglefield Bredning showed a weaker vertical gradient. Phosphate and silicate concentrations showed similar vertical variations to those for nitrate + nitrite (Figures 5b and 5c). POC and PON concentrations were relatively uniform horizontally and vertically, except near the surface (Figures 5d and 5e). At 1 km from the glacier, the concentrations of POC and PON near the surface were 54 and 8.0 μmol L\(^{-1}\), respectively. These concentrations within this layer gradually decreased toward Inglefield Bredning. The surface layer in the fjord closest to the glacier showed a low Chl. a concentration (Figure 5f). Interestingly, the highest concentration of Chl. a (4.8 μg L\(^{-1}\)) was observed at a depth of 20 m, 12 km from the glacier. The largest

**Figure 5.** Vertical profiles of (a) nitrate + nitrite, (b) phosphate, (c) silicate, (d) POC, (e) PON, (f) Chl. a (x axis on a logarithmic scale), and (g) mean ESD of suspended particles in seawater. The color scale indicates the distance from Bowdoin Glacier. The data collected at 1 and 23 km from Bowdoin Glacier are shown by polylines.

**Figure 6.** Comparison of concentrations of (a) TdFe and (b) dFe in supraglacial water, periglacial stream waters, and seawater. Each concentration is shown by its mean value with standard deviation.
mean equivalent spherical diameter (ESD) of the suspended particles (1.4 μm) was observed at the surface and 1 km from the glacier, and gradually decreased toward Inglefield Bredning (Figure 5g).

3.3. Comparisons of Fe, Macronutrients, POC, PON, and Particle Sizes Between Seawater and Glacial Meltwater

The concentrations of Fe in the supraglacial and periglacial stream waters were compared with those in seawater at a depth of 100 m in the fjord (Figure 6). The depth of 100 m was chosen for the comparison because the seawater in the shallower layer (<50 m) was influenced by glacial meltwater as indicated by specific changes in values of θ, S, and DO (Figure 3). The TdFe and dFe concentrations in the Supraglacial B, Stream B, and Stream T waters were 2–3 orders of magnitude higher than those in the seawater (Figures 6a and 6b). In particular, the Stream B and T waters were richer in TdFe and dFe than the Supraglacial B water. The TdFe and dFe concentrations in Stream B water were comparable with those in the Stream T water (Figures 6a and 6b).

In the same manner, macronutrients, POC, PON, and mean ESD ranges of the suspended particles in the supraglacial and periglacial stream waters were compared with those in the seawater (Table 1). Macronutrient concentrations were lower in the glacial meltwater than in the seawater (Table 1), as previously described by Kanna et al. (2018). The stream water was richer in POC and PON than the seawater (Table 1). The maximum value of mean ESD of the suspended particles in the glacial meltwater was greater than that in the seawater (Table 1). In particular, the mean ESD of the particles in the Stream B Est. water (approximately 5 μm) was the largest among all the sampled waters.

4. Discussion

4.1. Supply and Removal Processes of Glacially Sourced Iron in the Estuary

Our data indicated that glacial meltwater has the potential to greatly enhance the Fe content in Bowdoin Fjord (Figures 6a and 6b). Compared with the fjord seawater, the TdFe and dFe were richer in the Supraglacial B water generated on the surface of Bowdoin Glacier. After mixing with Supraglacial B water, the TdFe and dFe concentrations in the periglacial stream water (Stream B) were enhanced by interaction with glacial till, as suggested by the higher sediment inclusion in the Stream B water (79.0 ± 3.5 mg L⁻¹) compared with Supraglacial B water (11.9 ± 8.7 mg L⁻¹) (Table S1). The TdFe and dFe in the Stream T water originating from Tugto Glacier were also richer than those in the seawater. In particular, labile particulate Fe (pFe), which is calculated by subtracting the dFe concentration from TdFe, is a major form in these stream waters (accounting for >99% of TdFe).

The TdFe and dFe concentrations in the Stream B Est. water were lower than those in the Supraglacial B and Stream B water (Figure 6). The water salinity resulted to be approximately 3, indicating that the Stream B Est. water was mixed with seawater. The relatively lowered Fe concentrations in the Stream B Est. water was probably due to flocculation of Fe by the destabilizing influence of seawater electrolytes (e.g.,

### Table 1

|                | Macronutrients (μmol kg⁻¹) | Particulate organic matter (μmol L⁻¹) | Mean ESD (μm) |
|----------------|---------------------------|--------------------------------------|--------------|
|                | Nitrate + nitrite         | Phosphate                            | Silicate     |
| Supraglacial B (Supraglacial water) | 0.3 ± 0.2 (n = 13) | 0.2 ± 0.2 (n = 13) | <DL | 6.1 ± 3.4 (n = 9) | 0.6 ± 0.3 (n = 9) | 0.88-1.42 (n = 13) |
| Stream B (Periglacial stream water) | 1.9 ± 0.4 (n = 3) | 0.3 ± 0.1 (n = 3) | 0.7 ± 0.2 (n = 3) | 48 ± 14 (n = 3) | 3.0 ± 1.2 (n = 3) | 1.14-1.44 (n = 3) |
| Stream B Est. (S < 3) (Periglacial stream water) | 1.0 ± 0.7 (n = 3) | 0.1 ± 0.1 (n = 3) | 4.4 ± 2.1 (n = 3) | 14 ± 6.8 (n = 3) | 1.5 ± 0.6 (n = 3) | 1.41-4.86 (n = 3) |
| Stream T (Periglacial stream water) | 0.4 ± 0.1 (n = 3) | 0.3 ± 0.04 (n = 3) | 0.9 ± 0.1 (n = 3) | N/A | N/A | 1.40 (n = 1) |
| Seawater (100 m, S = ~34) | 12.3 ± 0.8 (n = 8) | 1.1 ± 0.1 (n = 8) | 12.9 ± 1.4 (n = 8) | 6.3 ± 3.1 (n = 8) | 0.4 ± 0.1 (n = 8) | 0.87-1.11 (n = 8) |

Note. N denotes the number of samples. The detection limit of silicate is 0.13 μmol kg⁻¹. N/A: not available.
Hunter, 1983) and precipitation of nanoparticulate/colloidal Fe at the fresh/seawater mixing zone, processes known as estuarine removal (Hopwood et al., 2016; Schroth et al., 2014; Zhang et al., 2015). Another possibility is the biological removal of Fe, although we do not have the data to confirm its significance. Despite such removal processes, the TdFe and dFe concentrations in the Stream B Est. water were higher than those in the fjord seawater (Figure 6), indicating that glacial meltwater is a significant source of Fe in Bowdoin Fjord.

4.2. Supply and Removal Processes of Glacially Sourced Iron in Plume Water

As discussed in section 4.1, we also expect that glacially sourced Fe is delivered to the fjord at depths of several hundred meters through subglacial channels of Bowdoin Glacier. Although Fe is supplied to fjords in Greenland from land-terminating (e.g., Bhatia et al., 2013; Hawkings et al., 2014) and marine-terminating glaciers (Hopwood et al., 2016), the removal process for glacially sourced Fe in fjords upon physical mixing processes induced by subglacial discharge plumes remains unknown. Here, the distribution of seawater influenced by subglacial discharge in the fjord is discussed, in addition to the dynamics of Fe in subglacial discharge plumes.

As shown in Figure 3a, our water mass analysis revealed that the seawater above 50 m near the glacier contains at least 5% of subglacial discharge. We refer to this near-surface water as plume water. The S and θ characteristics of the described plume water in Figure 3a can be interpreted as a combination of mixing of subglacial discharge with ambient seawater and meltwater from submerged ice. Near the calving front of the glacier, the subglacial discharge encountered ambient seawater at the depth of 200 m, causing changes in S and θ of the ambient seawater along a SD line depending on a mixing ratio $f_{sg}$. The ambient seawater entrained in the subglacial discharge subsequently upwells along the vertical face of the calving front, where the melting of submerged ice takes place, resulting in changes in S and θ of the seawater along a SM line (see Figure S2 for more details). The upwelling water finally reaches the shallow layer (<50 m) to form the plume water. On a θ-S diagram, the properties of the plume water fell in the section bounded by the SD and SM lines (Figure 3a), suggesting that the plume water contained submarine meltwater as well as subglacial discharge. The remainder in the plume water comes from the entrained seawater, which has low DO concentrations (approximately 280 μmol L$^{-1}$) (Figure 3b).

Turbidity has also been used as a tracer of subglacial discharge in previous studies in Greenland (e.g., Chauché et al., 2014; Kanna et al., 2018). The most turbid water in this study was distributed at 10 m over the entire transect along Bowdoin Fjord, whereas the turbidity gradually decreased with distance from the glacier (Figure 2c). Based on the water mass analysis, we conclude that the plume water distributed in a depth range of 10–50 m in Bowdoin Fjord at the time of the observation. This is consistent with the past two summer (late July/early August in 2014 and 2016) observations in this area, which also showed the existence of the plume water in the same depth range across the fjord (Ohashi et al., 2020).

Turbidity in seawater was positively correlated with the TdFe, dFe, and plFe concentrations (Figures 7a–7c). The relationship between turbidity and Fe suggested that Fe in the plume water originated from sediments from glacial weathering carried by subglacial discharge. Given the difference in TdFe and dFe concentrations in the seawater of the fjord, more than 97% of the TdFe was accounted for plFe.

Near the glacier, TdFe concentrations showed enrichment toward the surface and depletion in deep waters (Figure 4a). Given the formation process of the plume water, the vertical distribution of TdFe in the seawater within Bowdoin Fjord was likely formed as a result of an external input of Fe, that is, subglacial discharge supplies Fe deep into the fjord, and subsequently brings Fe near the surface through upwelling in front of the glacier. The plume water indicated the highest TdFe concentrations at 1 km from the glacier, and the concentration rapidly decreased with distance from the glacier (Figure 4a). Because Fe is rapidly adsorbed onto suspended particles (Bruland et al., 1994), Fe is likely removed from the water column near the glacier front.

In order to quantify the removal of plFe (i.e., concentration differences between TdFe and dFe), a major form of Fe in the fjord, the settling velocity for the plFe was calculated. The settling velocity depends on the size, density, and shape of the plFe according to the following equation (Stokes, 1851):
\[ V = g \times D^2 \times \left( \rho_p - \rho_w \right) / (18 \times \eta) \]

Where \( V \) is the setting velocity (m s\(^{-1}\)), \( g \) is the gravitational acceleration (9.8 m s\(^{-2}\)), \( D \) is the assumed mean ESD (m) of the plFe (see Figure 7c), \( \rho_p \) and \( \rho_w \) are the density (kg m\(^{-3}\)) of plFe and seawater at the depth of the sampled waters, respectively, and \( \eta \) is the viscosity of seawater at 1.88 \times 10^{-3} \text{ kg} \text{(m s)}^{-1}. It is assumed that \( \rho_p \) is 3.3 \times 10^{3} \text{ kg} \text{m}^{-3}, which was reported for amphibole (Zhang et al., 2013).

This is because the high amphibole (and rock fragments) contents in the plume water, as well as in surface sediments deposited near the lateral moraine of Bowdoin Glacier, were detected by X-ray diffraction (XRD) and scanning electron microscopy-energy dispersive spectroscopy (SEM-EDS) (Ando et al., 2019). For the plFe of 0.90–1.27 \( \mu \text{m} \) ESD in the plume water, \( V \) was estimated to be 4.6–9.2 cm day\(^{-1}\) (Figure 8). This calculated low settling velocity may keep the plFe in suspension in the plume water for a prolonged period of time. The shape and character of the flocs formed by particle aggregation also influence the settling velocity as previously observed in Disko Fjord, western Greenland (Markussen et al., 2016).

### 4.3. Horizontal Transport of Iron by Plume Water

The horizontal transports of Fe from Bowdoin Fjord to Inglefield Bredning were estimated from the vertical profiles of TdFe and dFe in the fjord. A recent study revealed that fjord stratification and a subglacial discharge flux in Bowdoin Fjord are important factors controlling the vertical distribution of turbidity (Ohashi et al., 2020), which also controls the vertical distributions of Fe. This implies that the depiction of the horizontal transports of Fe would change depending on the summer conditions in our studied region. The reduction in Fe during horizontal transport was defined as the distance over which Fe concentration dropped to 1/\( e \) of the value evaluated closest to the glacier. In order to calculate the total amount of TdFe within the plume water, TdFe concentrations obtained at observation sites were numerically integrated from 10 to 50 m depths using a linear regression with depth (Figure 9a). Following the procedure used by Johnson et al. (1997), natural log values of the integrated TdFe versus distance from the glacier were plotted (Figure 9c). Linear regression of the data yields a slope of \(-0.1295\), which corresponds to a scale length (computed as 1/slope) of 7.7 km. At a distance of 26 km in Inglefield Bredning, the total amount of TdFe in the plume water reduces to 3\% \((e^{-0.1295 \times 26} = 0.03)\) of the value attained 1 km from the glacier (100.4 mmol m\(^{-2}\)). The total amount of dFe within the...
plume water was calculated in the same manner, except that trapezoidal integration was used because linear regression between concentration and depth was not appropriate (Figure 9b). This should probably be interpreted with caution as the dFe are influenced by the biological uptake and by flocculation and precipitation. Natural log values of the integrated dFe versus distance from the glacier therefore yields linear regression with a slope of $-0.0545$ (Figure 9d), which corresponds to a scale length of 18.3 km. At a distance of 26 km in Inglefield Bredning, the total amount of dFe reduces to 24% ($e^{-0.0545 \times 26} = 0.24$) of the value at 1 km from the glacier (0.33 mmol m$^{-2}$). The foregoing analysis reveals that 3% of TdFe and 24% of dFe in the plume water are transported to Inglefield Bredning.

Next, the Fe flux from the plume water to Inglefield Bredning was evaluated. Jouvet et al. (2018) measured the flow speed of the plume in Bowdoin Fjord using an unmanned aerial vehicle in July 2016. Surface water flowed from a plume near the glacier front toward the fjord at a speed of 1–2 m s$^{-1}$ (Jouvet et al., 2018). According to ocean measurements performed in the same period at 1 km from the glacier, the plume water distributed at depths of 10–50 m in Bowdoin Fjord (Kanna et al., 2018). The spatial distribution of the plume water obtained in this study was similar to that during the previous observation in July 2016 (Kanna...
et al., 2018), which also showed a thickness of the plume water of 40 m. Assuming a uniform distribution of the plume water over the 3-km-wide fjord and a current off the glacier of 1–2 m s\(^{-1}\), the TdFe transported by the plume water was 1.5–2.9 \times 10^9 g day\(^{-1}\). Given the 97% reduction in concentration over the course of the transport, the plume water carried 4.5–8.7 \times 10^7 g day\(^{-1}\) of TdFe out of Bowdoin Fjord. In the same manner, dFe carried by the plume water out of Bowdoin Fjord was calculated to be 1.1–2.3 \times 10^6 g day\(^{-1}\). This represents the first quantitative description of the Fe transport in the buoyancy-driven upwelling system at the glacier front in Greenland. This is likely to be most representative of late July/early August, because temporal variation of the Fe transport in this system could be large throughout the summer melt season. Nevertheless, given that our studied period coincides with peak runoff at outlet stream from Qaanaaq Ice Cap (Mankoff et al., 2020), the result represents the Fe flux under high flux conditions of stream discharge.

### 4.4. Influence of Plume Water on Phytoplankton Growth

The subglacial discharge of Fe-rich meltwater entrains nutrient-rich deep waters and induces the buoyancy-driven upwelling at the glacial front, resulting in the formation of Fe- and nutrient-enriched plume water (Figure 10). Higher macronutrient concentrations are found at depths of 10–50 m in Bowdoin Fjord compared to Inglefield Bredning (Figures 5a–5c). This is because Bowdoin Fjord is affected by plume water upwelling with macronutrients to the subsurface layer. This section therefore discusses the potential influence of the plume water distribution on the growth of phytoplankton in the fjord.

The plume water is highly turbid from subglacial sediments, which could potentially suppress primary production by limiting light. Figure 11b depicts low phytoplankton abundance as shown by the depth-integrated Chl. \(a\) in the fjord within 6 km from the glacier, where high turbidity was observed. The phytoplankton was most abundant at 12 km from the glacier, resulting in increases in macronutrients consumption (Figure 11a). The maximum abundance of phytoplankton (phytoplankton bloom) was mainly associated with light availability in the relatively less turbid waters observed in this region. The bloom is expected to occur at sites 15 and 17 km from the glacier because turbidity in the water is low relative to the regions within 6 km from the glacier (Figure 11b) and the water contains sufficient amounts of macronutrients and dFe (Figure 11a). However, the phytoplankton abundance at these sites was as low as that at sites within 6 km from the glacier (Figure 11b). This is presumably due to the limitation of our single “snapshot” observation, which was performed during the initial stage of the bloom.

![Figure 10. Schematic of formation and distribution of Fe- and nutrient-enriched plume water off the front of Bowdoin Glacier.](image-url)
The phytoplankton abundance further decreased toward Inglefield Bredning (>20 km) (Figure 11b), likely because of nitrate limitation in the seawater shallower than 30 m (Figure 11a). Macronutrients are important resources for phytoplankton growth. Nitrate, in particular, is a main summer-time growth-limiting resource in regions around Greenland (Hopwood et al., 2018). Increase in phytoplankton biomass in spring results in increases in nitrate consumption, and consequently nitrate is depleted in early summer (Juul-Pedersen et al., 2015). Even though the water contains sufficient amounts of nitrate, nitrogen metabolic pathways of phytoplankton such as nitrogen fixation and nitrate + nitrite reduction are inhibited under very low concentrations of Fe (e.g., Morel & Price, 2003; Twining & Baines, 2013).

The dFe:nitrate + nitrite ratio in seawaters is a comprehensive tool to identify the factor limiting the phytoplankton community by comparing the ratio in seawaters to cellular Fe:nitrogen (N) ratios in phytoplankton (Birchill et al., 2019). Limited information is available on phytoplankton cellular Fe:N ratios in the coastal regions around Greenland. Nevertheless, cellular Fe:N ratios can be estimated by multiplying carbon (C):N ratios in the water column by previously reported phytoplankton cellular Fe:C ratios (>3.0 μmol mol\(^{-1}\)) (Ho et al., 2003; Maldonado & Price, 1996; Sunda & Huntsman, 1995; Twining et al., 2004).

In the present study, the C:N ratio was represented as a POC:PON ratio of 9.4 in 10–50 m at >18 km from the glacier, where the influence of terrigenous POC and PON through meltwater runoff is small. Indeed, the concentrations of POC and PON at >18 km from the glacier were positively correlated with Chl. \(a\) (Pearson correlation coefficients \(R = 0.71\) \((p < 0.01)\) and \(R = 0.78\) \((p < 0.01)\), respectively), suggesting that POC and PON are produced mainly by marine phytoplankton. Therefore, the lower limit of phytoplankton cellular Fe:N ratio (mmol mol\(^{-1}\)) was estimated to be roughly 0.03 by multiplying the reported phytoplankton cellular Fe:C ratio by the POC:PON ratio measured in this study.

The dFe:nitrate + nitrite ratio within the plume water in Bowdoin Fjord (1–15 km) varied from 0.58 to 3.3 (Figure 12), which was higher than the phytoplankton cellular Fe:N ratio of 0.03 (dashed red line in Figure 12). This result suggests that the plume water is dFe replete with respect to the phytoplankton demand. The dFe:nitrate + nitrite ratio of water in Inglefield Bredning (20 km) was also higher than the...
phytoplankton cellular Fe:N ratio. Therefore, the limiting factor for the phytoplankton growth both in Bowdoin Fjord and Inglefield Bredning was nitrate + nitrite and not the dFe concentration. It is worth noting that Inglefield Bredning was not affected by the nutrient-rich plume water, and thus, the nitrate + nitrite concentration remained at low levels through 10–20 m of the subsurface layer (Figure 5a). As a result, the dFe:nitrate + nitrite ratio in Inglefield Bredning showed a higher value in 10–20 m compared with that in Bowdoin Fjord, where nitrate + nitrite was supplemented with plume water (Figure 12). On the other hand, the dFe:nitrate + nitrite ratio in Irminger Sea, which is located off southeastern Greenland (see inset of Figure 1; Rijkenberg et al., 2014; Schlitzer et al., 2018), is comparable to the phytoplankton cellular Fe:N ratio (Figure 12). The dFe concentration in Irminger Sea is notably low; that is, the concentration is <0.5 nmol kg$^{-1}$ in 10–50 m (Rijkenberg et al., 2014; Schlitzer et al., 2018). This suggests that the water is nearly Fe depleted with respect to the phytoplankton demand. Because the water density of the East Greenland current is largely different from that of offshore Irminger Sea, transport of glacially derived Fe, as well as sediment-derived Fe from the East Greenland current to Irminger Sea is insignificant (Achterberg et al., 2018). The higher concentration of dFe in Bowdoin Fjord indicates the importance of subglacial discharge on Fe enrichment in a closed fjord system. Moreover, Inglefield Bredning, an approximately 100-km-long fjord system, is also likely affected by substantial amounts of glacially sourced Fe exported by eight glaciers terminating in the fjord (Figure 1), including the relatively large, fast-flowing Heilprin and Tracy Glaciers (Sakakibara & Sugiyama, 2018).

In other polar oceans, Fe derived from melting of ice shelves and floating glacier termini potentially fuels phytoplankton blooms. For example, Fe supplied by the melting of the marine ice layer at the base of the Amery Ice Shelf is an order of magnitude larger than the required Fe necessary to sustain the large annual phytoplankton bloom in Prydz Bay, East Antarctica (Herraez-Borreguero et al., 2016). Additionally, Fe input from the Pine Island Glacier is the largest Fe source in the Pine Island Bay, West Antarctica, which could satisfy the Fe demand for the phytoplankton bloom (Gerringa et al., 2012). Similar to studies off the coast of Antarctica, Fe input from a marine-terminating glacier in Greenland, which is supplied by a subglacial discharge plume, has the potential to fuel phytoplankton blooms in a glacial fjord.

This study presents the first quantitative description of the supply of Fe relative to that of the macronutrients in the buoyancy-driven upwelling system in front of a marine-terminating glacier. Commonly, it is known that the supply of Fe, relative to that of the macronutrients, plays an important role in allowing extensive phytoplankton blooms to develop in the wind-driven coastal upwelling systems (Bruland et al., 2001; Fitzwater et al., 2003; Johnson et al., 1999). Likewise, subglacial discharge and upwelling plumes supply abundant Fe and macronutrients to the euphotic zone, which sustains high productivity in glacial fjords throughout the melt season.

5. Conclusions

The Fe-rich subglacial discharge from the marine-terminating Bowdoin Glacier and the subsequent formation of a buoyant plume enhance the Fe concentration in the seawater of a fjord system, thereby enhancing the phytoplankton growth in the system. A significant portion of Fe delivered by subglacial discharge is removed after it mixes into the fjord seawater, probably through flocculation and precipitation processes. Only 3% of the total Fe ($4.5–8.7 \times 10^7$ g day$^{-1}$) was transported by the plume water from Bowdoin Fjord to Inglefield Bredning. Nevertheless, the plume water is Fe replete with respect to phytoplankton demands. Importantly, the effect of the buoyant plume upon the fjord circulation greatly enhances the nitrate delivery,
as well as renders the fjord ecosystem replete with Fe. The nitrate delivered by the plume water eventually mitigates the nitrate limitation of phytoplankton throughout the summer in the fjord. Our result highlights the importance of a subglacial discharge plume for triggering summer phytoplankton blooms in front of a marine-terminating glacier in Greenland.

Data Availability Statement

Data used in this study are available on the Arctic Data archive System (ADS), Japan (https://ads.nipr.ac.jp/dataset/A20200324-001; https://ads.nipr.ac.jp/dataset/A20200324-002).

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