

1 Characteristics of snowpack chemistry on the coastal region in the 2 northwestern Greenland Ice Sheet facing the North Water

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11 Abstract

12 In the North Water, the opening of sea ice due to polynya formation may influences the surrounding water and aerosol
13 circulation. We conducted glaciological observations from seacoast to inland on the western side of Prudhoe Land, which is
14 located in the northwestern Greenland Ice Sheet close to the North Water, to elucidate water and aerosol circulation around
15 the North Water. The spatial variations in $\delta^{18}\text{O}$ and chemical substances in surface snow showed that water vapor and aerosols
16 were directly transported from the southern North Water to northern inland of areas on the western side of Prudhoe Land.
17 Unlike the inland area of the Greenland Ice Sheet, the snowpack on the western side of Prudhoe Land contained signals of
18 ocean biological and sea ice conditions in the North Water. The spring–summer snowpack contained high concentrations of
19 methanesulfonic acid, which likely derived from sea-to-air dimethylsulfide emission following a phytoplankton bloom. The
20 autumn–winter snowpack exhibited high concentrations of NH_4^+ , which could potentially have originated from sea-to-air
21 ammonia gas emissions during the decline and death phases of marine organisms. The Na^+ , Cl^- , K^+ , Mg^{2+} , SO_4^{2-} , and Ca^{2+}
22 concentrations in the snowpack rapidly increased in winter, which could be attributed to the emission of frost flowers from the
23 newly formed sea ice surface into the atmosphere due to intense storm activity in the North Water. We suggest that the chemical
24 substances identified in the snowpack or ice core from the western side of Prudhoe Land can be used to better understand past
25 changes in ocean biological and sea ice conditions in the North Water.

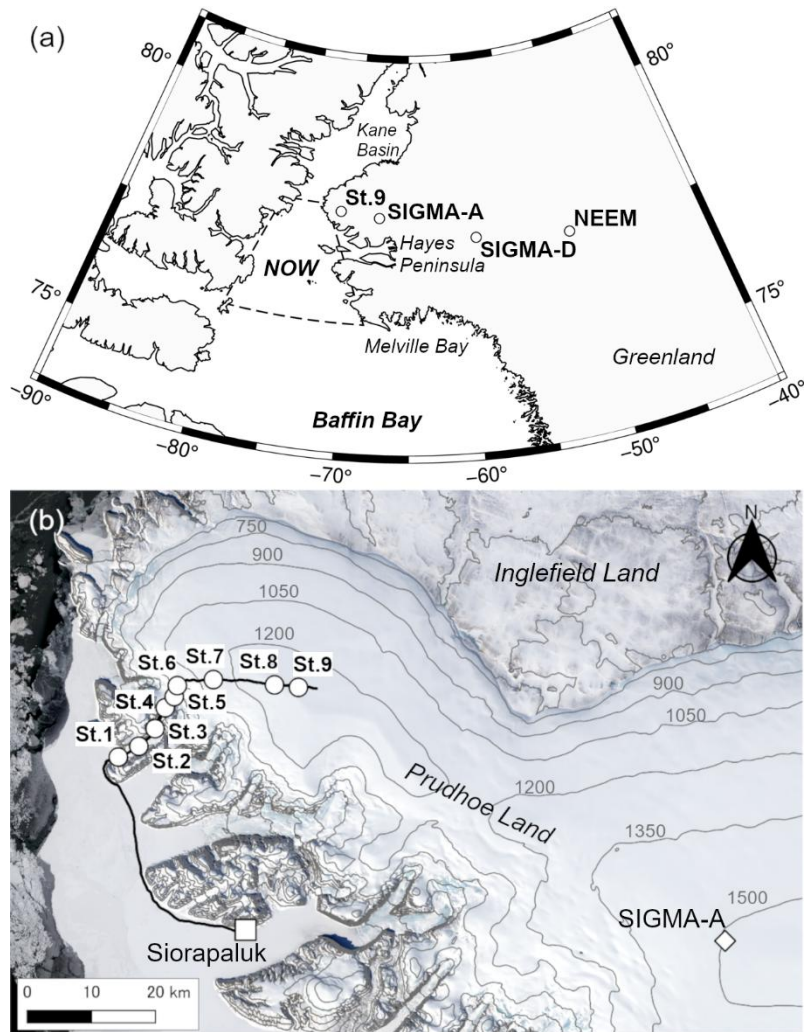
27 The North Water (NOW) is one of the largest polynyas in the Arctic and is located between northwestern Greenland and
28 Ellesmere Island in northern Baffin Bay (Barber and Massom, 2007) (Fig. 1a). The polynya in the NOW is formed by strong
29 northerly winds from the Nares Strait and the supply of relatively warm Atlantic waters yielded by a branch of the West
30 Greenland Current (Ito, 1982; Steffen and Ohmura, 1985; Mysak and Huang, 1992; Melling et al., 2001; Ingram et al., 2002;
31 Dumont et al., 2009; Vincent, 2019). Polynya formation in the NOW supplies heat and moisture to the atmosphere and
32 influences low-cloud formation around the NOW region. Monroe et al. (2021) reported that the low-cloud amount over
33 polynyas can reach up to twice as high as that over nearby sea ice in the NOW. Moreover, polynya formation in the NOW
34 influences ocean conditions such as vertical mixing by wind, the advection of nutrients into the euphotic layer, light availability,
35 and density stratification. These physical factors enhance primary production and collectively support high levels of
36 productivity in the NOW (Mei et al., 2002; Marchese et al., 2017). Additionally, the notable phytoplankton blooms promote
37 the production of dimethylsulfides (DMS), and sea-to-air DMS emission contributes to marine cloud formation (e.g. Charlson
38 et al., 1987). Sea ice fluctuations in the NOW influence not only water and aerosol circulation and ocean productivity but also
39 human activities. The recent increase in the frequency of sea ice breakup in front of Siorapaluk village, which is the
40 northernmost village in Greenland, influences hunting and fishing on sea ice and movement between villages on sea ice by
41 dog sledges and skidoos in winter and spring (Matoba and Yamasaki, 2018). Decreases in the sea ice thickness and
42 concentration in the NOW due to future climate warming could result in changes in water and aerosol circulation, ocean
43 productivity, and human lifestyle conditions.

44 To elucidate the changes in the source environmental conditions and the transport process of water vapor and aerosols over
45 northwestern Greenland, snow pit and ice core observations have been conducted in the northwestern Greenland Ice Sheet
46 (Kuramoto et al., 2011; NEEM Community members, 2013; Matoba et al., 2015; Matoba et al., 2018; Osterberg et al., 2015;
47 Kurosaki et al., 2020). The chemical components and water stable isotopes in the snowpack on the Greenland Ice Sheet provide
48 valuable information related to past changes in aerosol and water vapor conditions in source regions and their transport
49 processes. Snow pit observations (NEEM; 77.45°N, 51.06°W, 2445 m above sea level (a.s.l.) (Fig. 1a)) at the inland of the
50 northwestern Greenland Ice Sheet revealed seasonal variations in snowpack chemical compositions (Kuramoto et al., 2011).
51 For example, the concentration of Ca^{2+} originating from mineral dust peaked from late winter to spring, the concentrations of
52 SO_4^{2-} and NO_3^- originating from fossil fuel combustion peaked from winter to spring, and the concentration of methane
53 sulfonate⁻ (hereafter referred to as MSA) originating from marine phytoplankton peaked in spring and from late summer to
54 autumn (Kuramoto et al., 2011). Analysis of mineral dust in an ice core from SIGMA-D (77.636°N, 59.120°W, 2100 m a.s.l.),
55 which is located approximately 250 km from Qaanaaq on the northwestern coast of Greenland (Fig. 1a), revealed that the
56 amount of local dust originating from the Greenland coastal region increased recently due to a decrease in snow cover during
57 the warm season (Nagatsuka et al., 2021). Kurosaki et al. (2020) reported that the temporal variation in deuterium excess (d-
58 excess) in an ice core from SIGMA-A (78.05°N, 67.63°W, 1490 m a.s.l.), which is located approximately 70 km northeast of

59 Qaanaaq (Fig. 1a and b), was significantly correlated with the sea ice concentration in northern Baffin Bay from 1979–2015.
60 Via the use of relationship from the SIGMA-A ice core, the temporal variation in sea ice concentrations over 100 years in
61 northern Baffin Bay was reconstructed (Kurosaki et al., 2020). Because all of these studies were performed at relatively high
62 altitudes and far from the coast, the variations in chemical compositions and water stable isotopes were reflected by
63 environmental changes in relatively large areas driven by large-scale atmospheric circulation rather than local phenomena.
64 Matoba et al. (2014) conducted multiple snow pit observations at sites closer to the coast on the eastern side of Prudhoe Land,
65 which is located the northern part of Hayes peninsula in the northwestern Greenland facing the northeastern side of the NOW
66 (Fig. 1a and b), and they reported notable spatial variations in water stable isotopes and chemical compositions at lower
67 elevations and coastal regions. Water vapor and sea salt on the glacier facing the ocean were transported from the coast. On
68 the other hand, water vapor, mineral dust, anthropogenic substances, and MSA on the eastern side of Prudhoe Land were
69 transported from the central-west coast of Greenland, which is located around Disko Bay, via the central part of the Greenland
70 Ice Sheet (Matoba et al., 2014). The changes in the NOW could influence the water and aerosol circulation in the northwestern
71 Greenland. However, the past changes in the NOW have never been traced from the previous snow and ice core observations
72 in the northwestern Greenland Ice Sheet.

73 The western side of Prudhoe Land facing the northeastern side of the NOW is located between Inglefield Land in the north
74 and the southern part of Hayes peninsula in the south (Fig. 1a and b). The Inglefield Land is dry area where the precipitation
75 amount is quite low. On the other hand, the southern part of Hayes peninsula is one of the highest areas of precipitation amount
76 ($> 600 \text{ mm yr}^{-1}$) in the northwestern Greenland due to the southerly moisture advection (Bales et al., 2001). The western side
77 of Prudhoe Land, where locates between these regions with large contrast in precipitation, could be influenced by complicate
78 processes of the heat and moisture supply from the ocean around this region including the NOW. However, most of
79 meteorological and glaciological observations have been conducted on the southern part of Hayes peninsula in the northwestern
80 Greenland (Bales et al., 2001; Osterberg et al., 2015; Becagli et al., 2016; Akers et al., 2020), not on Prudhoe Land. In order
81 to make accurate future projections of environmental change in this region, where the influences of climate warming on the
82 life activities in this region is already emerging, it is necessary to clarify the impact of changes in the NOW on the processes
83 of the heat and moisture circulation in this region.

84 In the Prudhoe Land, a gentle valley separates the western region from the eastern region, where the SIGMA-A site is
85 situated. (Fig. 1b). This valley could serve as a pathway for air masses transported from inland of the Greenland Ice Sheet to
86 descend toward the coastal region, and these air masses are likely not transported to the western side of Prudhoe Land. If that
87 is the case, the snowpack on the western part of Prudhoe Land could contain aerosols originating from the NOW without being
88 mixed with aerosols from the interior of the Greenland Ice Sheet. We conducted glaciological observations on the western side
89 of Prudhoe Land facing the NOW to elucidate water and aerosol circulation around the NOW, which influences the
90 environmental changes and related human activities in this region.



91
 92 **Figure 1: Maps of the sampling sites.** (a) shows location of the snowpit and ice core sampling sites in this study (St. 9) and
 93 previous studies (SIGMA-A, SIGMA-D, and NEEM) in the northwestern Greenland Ice Sheet. The dashed polygon in (a)
 94 denotes the approximate location of the NOW. Hayes peninsula in the northwestern Greenland is located between Kane Basin
 95 in the north and Melville Bay in the south. (b) shows Landsat-8 image around St. 9 and SIGMA-A of Prudhoe Land, which is
 96 located on the northern part of Hayes peninsula, on 13 April 2023. The black circles in (b) denote the sampling sites from St.
 97 1 to St. 9, and the black line denotes dog sledge route. The gray contours in (b) are drawn from the Greenland Mapping Project
 98 2 (GIMP-2) Digital Elevation Model version 2.

100 **2.1 Observations and chemical analysis**

101 We conducted snow observations from 9–11 April 2023 on the western side of Prudhoe Land, which is a coastal site in the
102 northwestern part of the Greenland Ice Sheet (Fig. 1). The expedition base was established in Siorapaluk village at 77.78°N
103 and 70.75°W. Siolapaluk village faces Robertson Fjord, which is located on the northeastern side of the NOW (Fig. 1b). The
104 observation route is shown in Fig. 1b. We started from Siorapaluk on 9 April 2023 on dog sledges, moved on sea ice westward
105 from Sioraparuk, climbed the Clements Glacier, and advanced on the Greenland Ice Sheet. We arrived at the terminal point of
106 the route (78.13°N, 71.06°W, 1279 m above sea level (St. 9)) on 10 April 2023 and returned to Siorapaluk on 11 April 2023.
107 We conducted glaciological observations along the route. The locations and elevations of the observation sites are shown in
108 Fig. 1b and Fig. S1. We conducted snow pit observations on the glacier (St. 3) and the ice sheet (St. 9) along the route to
109 measure the chemical species, stratigraphy, and density of the snowpack. Snow samples for chemical analysis at St. 3 and St.
110 9 were collected from the snow walls of the snow pit using a precleaned stainless-steel sampler into ®Whirl-Pak polyethylene
111 bags (Nasco). The snow sampling intervals at St. 3 were 0.02 m from 0.00 to 0.20 m and 0.03 m from 0.20 to 1.01 m, and the
112 snow sampling intervals at St. 9 were 0.02 m from 0.00 to 0.30 m and 0.03 m from 0.30 to 1.08 m. Additionally, an ice core
113 was collected at St. 9 from 1.08 to 4.20 m using a hand corer. After the stratigraphy and density of the ice core were measured,
114 the ice core samples were cut at 0.05–0.10 m intervals using a precleaned saw, were polished using a precleaned ceramic knife
115 to remove contamination on the ice core samples and were then placed into ®Whirl-Pak polyethylene bags (Nasco). Surface
116 snow samples for chemical analysis were also collected at 9 sites along the route using a precleaned stainless-steel sampler
117 into Whirl-Pak polyethylene bags. We removed the potential contamination by organic compounds in general on the material
118 and tools used for contamination removal using ethanol, and then performed ultrasonic cleaning in ultrapure water. The snow
119 samples were kept frozen and were transported to Siorapaluk. The snow samples were melted at ambient temperature, were
120 placed in 50 mL clean polypropylene bottles, and were then kept frozen in a portable freezer (approximately –18 °C). The
121 samples were then transported to a cold room (–20 °C) at the Institute of Low Temperature Science (ILTS) of Hokkaido
122 University, Japan, and were then kept frozen until chemical analysis.

123 The concentrations of Na^+ , NH_4^+ , K^+ , Mg^{2+} , Ca^{2+} , MSA, Cl^- , SO_4^{2-} , and NO_3^- were measured by ion chromatography (ICS-
124 2100, Thermo Scientific). For the cations, separation was carried out with a Dionex CG-12 (4 × 50 mm) guard column,
125 followed by a Dionex CS12-A (4 × 250 mm) separation column. Injection volume of samples was 500 µL. MSA (20 mM) was
126 used as eluent, and flow-rate was kept 1.0 mL min⁻¹. Dionex CDRS600 dynamically regenerated suppressor was used for
127 conductivity suppression before conductivity cell. For the anions, separation was obtained with a Dionex AG-18 (4 × 50 mm)
128 guard column and Dionex AS-18 (4 × 250 mm) separation column. Injection volume of samples was 1000 µL. KOH (23 mM)
129 was used as eluent, and flow-rate was kept 1.0 mL min⁻¹. Dionex ADRS 600 dynamically regenerated suppressor was used for
130 conductivity suppression before conductivity cell. The absolute calibration curve method was used for quantitative

determination of each ion concentration. For the absolute calibration determination by ion chromatography, we used the standard solutions for ion chromatography produced by FUJIFILM Wako Pure Chemical corporation, diluted to 20, 50, 100, and 200 ppb with ultra-pure water. If the ion concentration of samples were outside the calibration range (> 200 ppb), it was remeasured using 500, 1000, 2000–3000, and 6000 ppb standard for the anions and 500, 1000, 2000, and 4000 ppb standard for the cations. Blanks were always evaluated before the calibration procedure. The analytical precision of the ion chromatography was $< 5\%$ (at the measurement of 20 ppb standard). The limit of detection (LOD) was < 0.1 ppb. The limit of quantification (LOQ) was < 0.5 ppb. The ion chromatography method in this study had been evaluated by our previous study (Kurosaki et al., 2020; Kurosaki et al., 2022).

The non-sea-salt (nss) components of Ca^{2+} and SO_4^{2-} were calculated using the seawater ratios of these ions with respect to Na^+ , thereby assuming that Na^+ is solely of a sea-salt origin. For example, the concentration of nssSO_4^{2-} was calculated following Eq. (1):

$$[\text{nssSO}_4^{2-}] = [\text{SO}_4^{2-}] - (\text{SO}_4^{2-}/\text{Na}^+)_{\text{sea}} \times [\text{Na}^+], \quad (1)$$

We applied the concentration unit as $\mu\text{eq L}^{-1}$ in the Eq: (1). The $(\text{SO}_4^{2-}/\text{Na}^+)_{\text{sea}}$ is the equivalent concentration ratio of $\text{SO}_4^{2-}/\text{Na}^+$ in the sea water, which is 0.12. Similarly, the concentrations of nssCa^{2+} were calculated using sea water ratios of 0.044 (Wilson, 1975; Legrand and Mayewski, 1997).

The stable oxygen and hydrogen isotopic compositions of the water samples were measured by a water stable isotopes analyzer (L2130-I, Picarro Inc.) with an evaporating device (A0211, Picarro Inc.). We used the ultrapure water ($\delta^{18}\text{O} = -11.583$ and $\delta\text{D} = -77.2$), Antarctic iceberg ($\delta^{18}\text{O} = -20.4$ and $\delta\text{D} = -158.7$), snowpack on the Antarctic ice Sheet ($\delta^{18}\text{O} = -46.694$ and $\delta\text{D} = -370.7$) for calibration. The analysis precisions for $\delta^{18}\text{O}$ and δD were 0.08‰ and 0.8‰, respectively.

2.2 Meteorological and sea ice data

We used meteorological and sea ice data from northwestern Greenland and Baffin Bay to analyze the factors contributing to the variations in chemical species and water stable isotopes in snow on the northwestern coast of Greenland. The temperature, relative humidity, wind speed, wind direction, and atmospheric pressure at Siorapaluk were measured by an automatic weather station (AWS) (WXT530, Vaisala) from April 2021 to March 2024 (Matoba et al., 2024). The AWS was installed 2 m above ground level (a.g.l.), and these meteorological factors were observed at 10-minutes intervals. We used data of the air temperature, geopotential height, wind speed, and wind direction around the northwestern coast of Greenland and Baffin Bay from ERA5 reanalysis dataset supplied by the European Center for Medium Range Weather Forecasts (ECMWF) (Hersbach et al., 2020). Additionally, we used 2 m air temperature in the western side of the Prudhoe Land from ERA5-Land reanalysis dataset supplied by the ECMWF (Muñoz-Sabater et al., 2021).

163 **2.3 Backward trajectory**

164 To investigate the source region and transport pathway of water vapor and aerosols contained in ice core at the St. 9 site, we
165 analyzed air mass position along the backward trajectory from the St. 9 site during the past 7 days using the National
166 Oceanographic and Atmospheric Administration (NOAA) Hybrid Single-Particle Lagrangian Integrated Trajectory
167 (HYSPPLIT) model (Stein et al., 2015) and National Centers for Environmental Prediction (NCEP) reanalysis data. The initial
168 positions of air mass were set at 50, 500, 1000, 1500 m above ground level over the St. 9 site. The initial date and time were
169 every 6 h from 2019 to 2023. We calculated existing probability of an air mass with a 1° resolution. The existence probability
170 was defined as the proportion of backward trajectories originating from the St. 9 that passed through each 1°×1° grid cell. We
171 counted the number of times the backward trajectories originating from St. 9 passed through each grid cell. Subsequently, we
172 normalized the count for each grid cell by dividing it by the total number of the trajectory passes across all grid cells.
173 Considering the water vapor and aerosols supply from the ocean and land surface, we excluded air mass over 1000 m above
174 ground level. The existence probability was weighted by the daily amount of precipitation when the air mass arrived at the St.
175 9 site. The daily amount of precipitation was extracted from the ERA5 reanalysis dataset (Hersbach et al., 2020).

176 **3 Results and Discussion**

177 **3.1 Snowpack dating and annual accumulation**

178 The snowpack at St. 3 and St. 9 was dated on the basis of the snow stratigraphy and seasonal variations in $\delta^{18}\text{O}$, d-excess
179 ($\text{d-excess} = \delta\text{D} - 8\delta^{18}\text{O}$) and MSA. The vertical profiles of the snow stratigraphy at St. 3 and St. 9 are shown in Fig. S2.
180 Rounded grains, faceted crystals, and depth hoars were observed from 0.00 to 1.01 m in the glacier at St. 3 (Fig. S2). Glacier
181 ice was present below 1.01 m at St. 3. The snowpack from 0.00 to 1.01 m at St. 3 corresponds to seasonal snow for the periods
182 from autumn in 2022 to spring in 2023. Rounded grains, faceted crystals, and depth hoars were observed from 0.00 to 0.96 m
183 at St. 9 (Fig. S2). Below 0.96 m at St. 9, melt forms prevailed, and some ice layers with thickness ranging from several
184 millimeters to 10 cm were found (Fig. S2). The ice layers indicated that the surface or subsurface of the snowpack melted in
185 summer. Therefore, we assumed that the depth of 0.96 m at St. 9 corresponds to the end of summer in 2022. Vertical profiles
186 of $\delta^{18}\text{O}$, d-excess and MSA at St. 9 are shown in Fig. 2. The seasonal variation in $\delta^{18}\text{O}$ reflects the air temperature and exhibits
187 a maximum value in summer and a minimum value in winter at several Greenland sites (Steffensen et al., 1988; Legrand and
188 Mayewski, 1997; Kuramoto et al., 2011; Kurosaki et al., 2020; Nakawaza et al., 2021). The seasonal variation in d-excess
189 reflects the sea surface temperature and sea-to-air evaporation process in the water vapor source region of precipitation
190 (Merlivat and Jouzel, 1979; Uemura et al., 2007; Kurita, 2011; Kopec et al., 2019). At several sites in Greenland, the d-excess
191 of snowpack exhibits maximum values in autumn and minimum values from spring to early summer (Johnsen et al., 1989;
192 Kuramoto et al., 2011). MSA is an oxidization product from dimethyl sulfide (DMS), which originates from ocean

193 phytoplankton. The seasonal variation in MSA exhibits maximum values from spring to summer during ocean phytoplankton
194 blooms and minimum values in winter (Jaffrezo et al., 1994; Legrand and Mayewski, 1997; Kuramoto et al., 2011; Osterberg
195 et al., 2015; Nakazawa et al., 2021; Kurosaki et al., 2022). We determined the seasonality of the snowpack from 0.00 to 1.15
196 m as shown below. The date of snow layer from 0.00–0.04 m was close to the observation date (April 2023). We determined
197 that snowpack in 0.04–0.72 m corresponded to the autumn–winter period of 2022–2023, based on the presence of negative
198 $\delta^{18}\text{O}$ peak, positive d-excess peak and low MSA concentrations (Fig. 2). The snowpack from 0.72 m–1.15 m was interpreted
199 to correspond to the spring–summer period in 2022, due to the presence of ice layer, high $\delta^{18}\text{O}$ value and high MSA
200 concentrations (Fig. 2 and Fig. S2). Snowpack below 0.96 m corresponded to previous summer and before it because the
201 amplitude of seasonal variation of $\delta^{18}\text{O}$ and d-excess below 0.96 m were smaller than those in the shallower layers from 0.00–
202 0.96 m because of summer melting (Fig. 2a and b). The MSA concentration showed obvious seasonal variations, and the $\delta^{18}\text{O}$
203 values exhibited slight seasonal variations below 0.96 m (Fig. 2a and c), although the values of the water stable isotopes were
204 smoothed by summer melting and some chemical species showed high peaks in the ice layers owing to relocation processes
205 by meltwater refrozen. We assigned the snow layers with high MSA and $\delta^{18}\text{O}$ values to the spring to summer period and
206 assigned the snow layers with low MSA and $\delta^{18}\text{O}$ values to autumn to winter period. Consequently, the firm layers from 1.15–
207 1.45 m, 2.02–2.28 m, and 3.12–3.39 m corresponded to the autumn–winter periods from 2021–22, 2020–21, and 2019–20,
208 respectively. The snowpacks from 1.45–2.02 m and 2.28–3.12 m corresponded to the spring–summer periods in 2021 and
209 2020, respectively. According to the snowpack dating approach described above, the snow densities, which were the average
210 of the bulk density of ice core at St. 9 (Fig. S3), from autumn–winter 2020 to autumn–winter 2021 and from autumn–winter
211 2019 to autumn–winter 2020 were 477 kg m^{-3} and 497 kg m^{-3} , respectively. Using these mean densities, the annual
212 accumulations were calculated to be $0.41 \text{ m w.eq. yr}^{-1}$ and $0.56 \text{ m w.eq. yr}^{-1}$, respectively. The Fifth Generation Mesoscale
213 Model modified for polar climates (Polar MM5) using firm core and meteorological station data estimated that the
214 climatological value of the annual accumulation rate on the inland of the western side of Prudhoe Land ranged from $0.31\text{--}0.43$
215 m w.eq. yr^{-1} for the 1958–2007 period (Burgess, et al., 2007). The annual accumulation rate derived from the snowpack at St.
216 9 in this study was comparable to or slightly higher than the climatological value.

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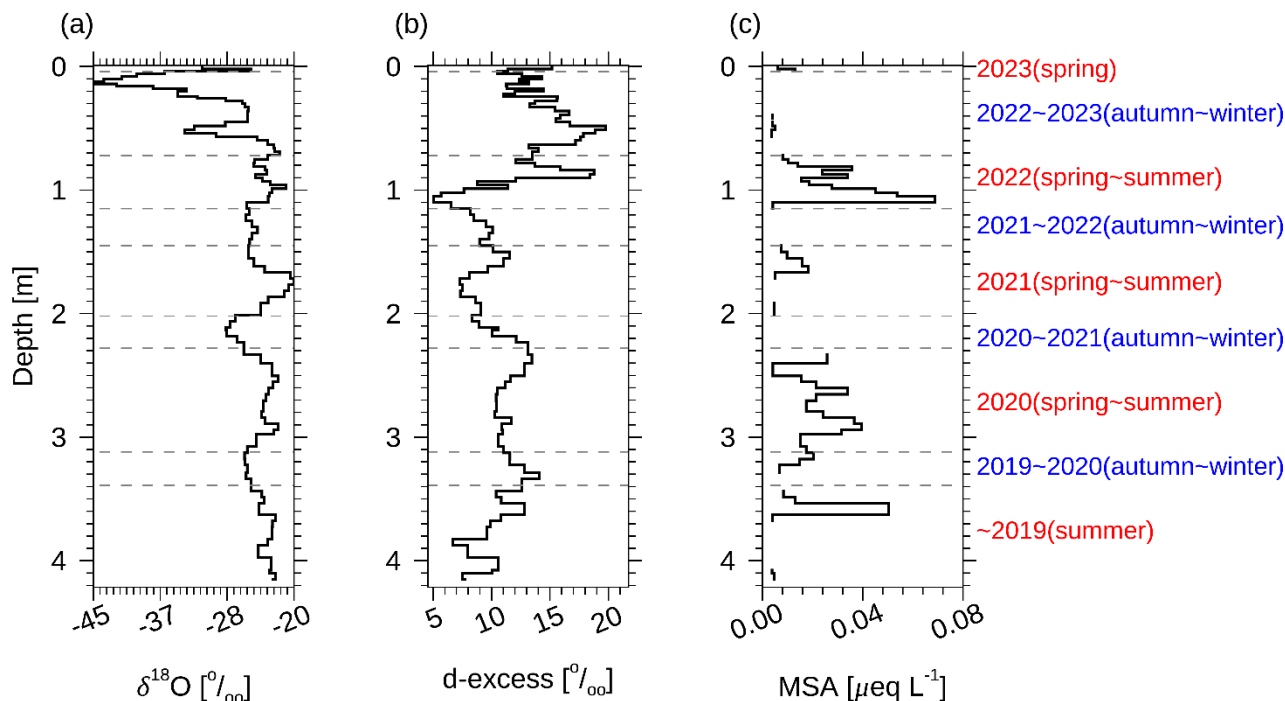


Figure 2. Vertical profiles of (a) $\delta^{18}\text{O}$, (b) d-excess, and (c) MSA at St. 9. The LOD of MSA was $< 0.0011 \mu\text{eq L}^{-1}$.

3.2 Spatial and temporal variations in $\delta^{18}\text{O}$ and chemical species

3.2.1 $\delta^{18}\text{O}$

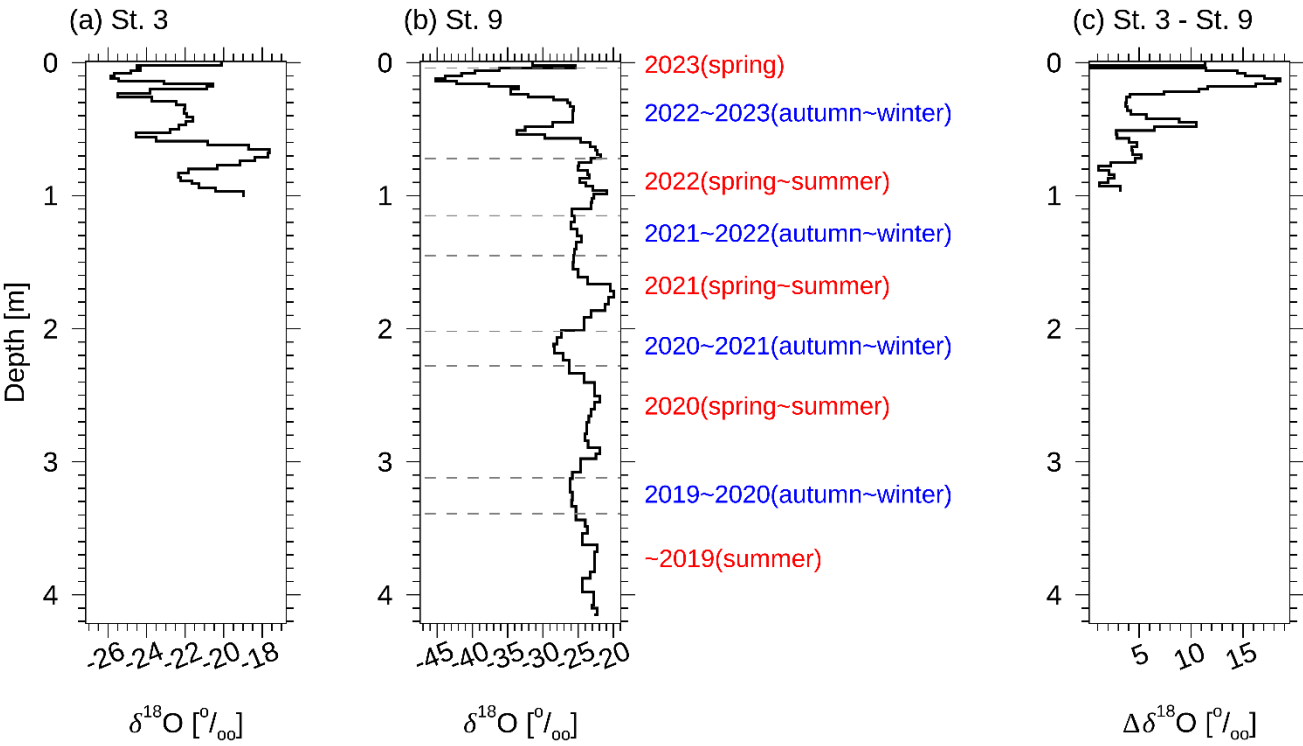
The spatial variations in $\delta^{18}\text{O}$ in the surface snow showed maximum and minimum values at St. 3 (-19.12 ‰) and St. 9 (-37.21 ‰), respectively (Fig. S4a). The average $\delta^{18}\text{O}$ value from 0.00 to 1.01 m at St. 3 was greater than that at St. 9 (St. 3: -22.03 ‰ ; St. 9: -29.12 ‰) (Table 1). The $\delta^{18}\text{O}$ values in surface snow and the snowpack decreased from the seacoast toward the inland site. The past 7 days backward trajectory arriving at the St. 9 also exhibited that majority of air mass was transported from the south of the St. 9, situated on the northern Baffin Bay and eastern NOW (Fig. S5). We suggest that the south-to-north gradient of $\delta^{18}\text{O}$ results from water vapor, which originates from northern Baffin Bay and eastern NOW, transport from the southern coast to the northern inland area by southerly winds. On the other hand, Matoba et al. (2014) reported different spatial variations in $\delta^{18}\text{O}$ in surface snow on the eastern side of Prudhoe Land. The $\delta^{18}\text{O}$ value of surface snow on the northwestern coast of Greenland decreased along the elevation on a glacier from sea level to 1,100 m a.s.l. and increased toward the inland on the Greenland Ice Sheet. Matoba et al. (2014) suggested that the distribution of $\delta^{18}\text{O}$ values in surface

snow resulted from the water vapor pathway, which originates from southern Baffin Bay, after which it first progress toward the northeast and then moves the inland of the Greenland Ice Sheet, and if finally turns northwest on the western coast of Greenland because of the cyclonic flow into a depression at Melville Bay. Because the observation site in this study is located in the westernmost part of Prudhoe Land, it is not affected by water vapor passing through the interior of the Greenland Ice Sheet, and water vapor transport from the south mainly influences the spatial variation in $\delta^{18}\text{O}$.

Fig. 3 shows the vertical profile of $\delta^{18}\text{O}$ in the snowpack at St. 3 and St. 9, and the difference in $\delta^{18}\text{O}$ between St. 9 and St. 3. The vertical profile of $\delta^{18}\text{O}$ in the snowpacks at St. 9 was similar to that at St. 3. The depths of the negative and positive peaks of $\delta^{18}\text{O}$ at St. 9 agreed well with those at St. 3, and the vertical profile of $\delta^{18}\text{O}$ between 0.00 and 1.01 m at St. 9 correlated significantly with that at St. 3 ($r = 0.69$, $p < 0.01$). The snowpack corresponding to autumn–winter from 2022–2023 at St. 3 and St. 9 at the same snow depth were most likely accumulated with precipitation attributed to the same snowfall events, and $\delta^{18}\text{O}$ in the snowpack had not been changed by the post depositional processes, which is melting and refreezing, water molecule diffusion, wind blowing, and sublimation. Therefore, we propose that the vertical profile of $\delta^{18}\text{O}$ between 0.00 and 1.01 m at St. 9 can be reasonably compared with the profile at St. 3 based on their differences. The difference in $\delta^{18}\text{O}$ values between St. 3 and St. 9 increased from spring–summer to autumn–winter from 2022–2023 and decreased until spring in 2023 (Fig. 3c). Matoba et al. (2002a) reported that the $\delta^{18}\text{O}$ values of precipitation in Siorapaluk was positively correlated with the surface air temperature. We suggest that the altitude gradient of the surface air temperature in winter was greater than that in summer in the western region of Prudhoe Land. We estimated the difference in surface air temperature between St. 3 and St. 9 using ERA5-Land reanalysis dataset (Fig. S6 and Fig. S7). The temperature difference between St. 3 and St. 9 was smallest in summer and increased toward winter. The mean temperature differences in autumn and winter were larger than those in summer. This result supports our suggestion, based on water stable isotope, that the altitude gradient of surface air temperature in the western side of Prudhoe Land was steeper in winter than in summer.

Table 1. Mean values and standard deviations of several ion species and water stable isotopes between St. 3 and St. 9.
The mean values at St.9 were obtained at depths from 0.00 to 1.01 m.

	Na ⁺ (μeq L ⁻¹)	NH ₄ ⁺ (μeq L ⁻¹)	K ⁺ (μeq L ⁻¹)	Mg ²⁺ (μeq L ⁻¹)	nssCa ²⁺ (μeq L ⁻¹)	MSA (μeq L ⁻¹)	Cl ⁻ (μeq L ⁻¹)	nssSO ₄ ²⁻ (μeq L ⁻¹)	NO ₃ ⁻ (μeq L ⁻¹)	Cl ⁻ /Na ⁺ (μeq L ⁻¹)	δ ¹⁸ O (‰)	δD (‰)	d-excess (‰)
mean(St.3)	133.70	0.61	2.77	22.69	-0.46	0.01	155.53	1.51	1.32	1.19	-22.03	-160.27	15.93
std. (St.3)	181.51	0.45	4.26	27.51	0.87	0.02	206.34	3.45	1.19	0.15	2.22	17.58	2.12
mean(St.9)	34.17	0.04	0.68	7.11	0.85	0.01	41.37	0.40	0.60	1.17	-29.12	-218.69	14.29
std. (St.9)	54.12	0.22	1.25	9.12	3.67	0.01	65.33	0.92	0.44	0.12	6.46	52.27	2.66



263

264 **Figure 3: Vertical profile of $\delta^{18}\text{O}$. (a) and (b) show $\delta^{18}\text{O}$ values at St. 3 and St.9, respectively. (c) shows difference**
265 **between St.3 and St.9 in terms of $\delta^{18}\text{O}$. i–vii denote seasons from 2019 to 2023. i, iii, v, and vii denote from autumn to winter**
266 **period from 2022–2023, 2021–2022, 2020–2021, and 2019–2020, respectively. ii, iv, and vi denote from spring to summer in**
267 **2022, 2021, and 2020, respectively.**

268

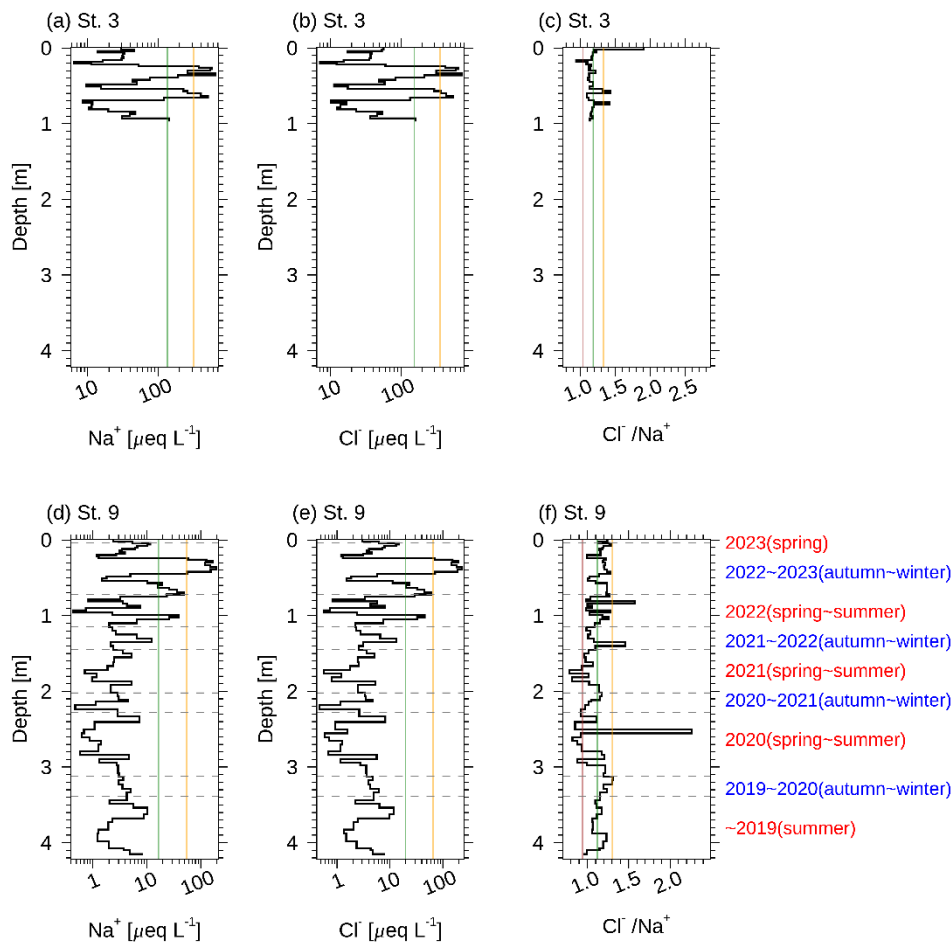
270 The possible sources of Na⁺ and Cl⁻ are sea salt and frost flowers (Legrand and Mayewski, 1997). The Na⁺ and Cl⁻
 271 concentrations in surface snow decreased with increasing distance from the seacoast (Fig. S4b and g). The average values of
 272 Na⁺ and Cl⁻ between 0.00 and 1.01 m at St. 3 were four times greater than those at St. 9 (Table 1). High values of these ions
 273 at St. 3 and St. 9 were found from autumn–winter (Fig. 4). The spatial variations in Na⁺ and Cl⁻ in surface snow were similar
 274 to those in the eastern region of Prudhoe Land (Matoba et al., 2014). The Na⁺ and Cl⁻ concentrations at several sites in the
 275 inland of the Greenland Ice Sheet also exhibit maximum values in winter (Steffensen, 1988; Fischer and Wagenbach, 1996;
 276 Dibb et al., 2007; Kuramoto et al., 2011).

277 The Na⁺ and Cl⁻ concentrations in snow are modified with respect to the bulk seawater composition due to heterogeneous
 278 reactions with atmospheric acids such as H₂SO₄ and HNO₃ (Cadle, 1972; Delmas et al., 1982; Kerminen et al., 2000):



281 Na₂SO₄ and NaNO₃ are removed from the atmosphere through the rainout process following the formation of cloud
 282 condensation nuclei and precipitation. Therefore, the difference of the Cl⁻/Na⁺ ratio in snowpack from sea-water ratio often
 283 reveals sea salt modification within the atmosphere. In general, if sea-salt is transported for a long time in the atmosphere, Na⁺
 284 is scavenged to a greater degree than Cl⁻ due to reactions with H₂SO₄ and HNO₃, and Cl⁻/Na⁺ is thus higher than the sea-water
 285 ratio (1.17). The Cl⁻/Na⁺ ratio in the snowpack at St. 9 was close to the seawater ratio for all seasons (Fig. 4). Additionally,
 286 the values of Cl⁻/Na⁺ in the snowpack at St. 3 and in surface snow at all the sites were close to the seawater ratio (Figs. S3k
 287 and 4). At several sites in the northwestern Greenland Ice Sheet, Cl⁻/Na⁺ exhibited maximum values in summer (Kuramoto et
 288 al., 2011; Kurosaki et al., 2020). In contrast, the Cl⁻/Na⁺ ratio at St.9 did not show clear seasonality. Moreover, the spatial
 289 distribution of air mass transport to the St.9 exhibited that primary source of sea salt was the area near the St. 9, situated on
 290 the eastern NOW, throughout the year (Fig. S5b). Therefore, the sea salt observed in this study could be transported along a
 291 short-distance pathway without reactions with H₂SO₄ and HNO₃ in the atmosphere throughout the year, and the possible source
 292 region is the NOW polynya.

293



294

295 **Figure 4: Vertical profile of Na^+ , Cl^- , and Cl^-/Na^+ at (a–c) St. 3 and (d–f) St. 9.** Green lines denote mean ion
 296 concentration across all observation depths. Orange and brown lines denote the mean ion concentrations plus and minus
 297 standard deviation across all observation depths, respectively. The gray dashed lines in (c) and (f) denote Cl^-/Na^+ in seawater
 298 (1.17). The LOD of Na^+ and Cl^- was < 0.0043 and $< 0.0028 \mu\text{eq L}^{-1}$, respectively.

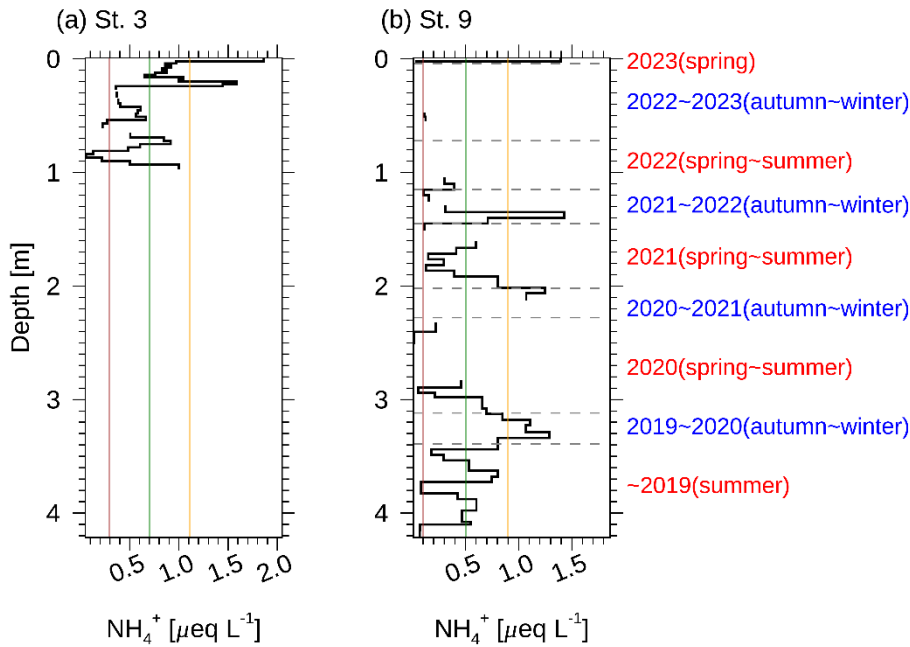
299

300

301 3.2.3 NH_4^+

302 The possible source of NH_4^+ is biomass burning (Fuhrer and Legrand, 1997; Legrand and Mayewski, 1997; Kuramoto et al.,
 303 2011). NH_4^+ in surface snow reached maximum values at St. 5 (Fig. S4c). NH_4^+ in the snowpack at St. 3 and St. 9 exhibited
 304 minimum values from spring–summer and maximum values from autumn–winter (Fig. 5). Positive peak values of NH_4^+ were
 305 exhibited at depths of 1.38, 2.04, and 3.31 m corresponding to autumn–winter periods from 2021–2022, 2020–2021, and 2019–

306 2020, respectively (Fig. 5). According to the stratigraphy of the snowpack at St. 9 (Fig. S2), we attributed the peaks at 1.38,
 307 2.04, and 3.31 m to the deposition of atmospheric ammonium, without the influence of melting and refreezing process. At
 308 several sites in the inland of the Greenland Ice Sheet, the concentration of NH_4^+ peaked from spring–summer owing to biomass-
 309 burning events (Dibb et al., 2007; Nakazawa et al., 2021; Kjær, et al., 2022). In contrast, the concentrations of NH_4^+ at St. 9
 310 did not peak in spring–summer. The autumn–winter maximum of NH_4^+ at St. 3 and St. 9 have not been observed at the inland
 311 sites of the Greenland Ice Sheet. According to an incubation experiment of nitrogen-fixing organisms using artificial seawater,
 312 a substantial increase in the atmospheric emission of ammonia gas and gas-phase basic water-soluble organic nitrogen (WSO_N)
 313 was found during the decline and death phases of the organisms (Dobashi, 2023). The NOW polynya facing our study sites is
 314 one of the most biologically productive marine areas of the Arctic Ocean (Klein et al., 2002; Odate et al., 2002). Phytoplankton
 315 blooms in the NOW onsets in April, are sustained for three months, and then decline and death from autumn–winter (Mei et
 316 al., 2002; Marchese et al., 2017). We suggest that the increase in NH_4^+ in the snowpack at St. 3 and St. 9 from autumn–winter
 317 was caused by the increase in sea-to-air emission of ammonia gas during the decline and death phases of phytoplankton in the
 318 NOW polynya.
 319

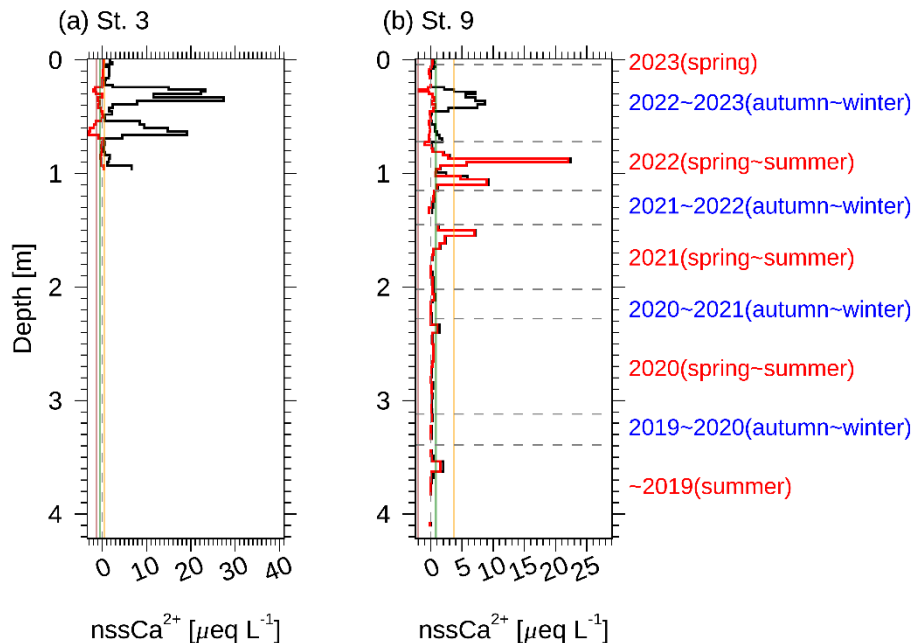


320
 321 **Figure 5: Vertical profile of NH_4^+ at (a) St. 3 and (b) St. 9.** Green lines denote mean NH_4^+ across all observation depths.
 322 Orange and brown lines denote the mean NH_4^+ plus and minus one standard deviation across all observation depths,
 323 respectively. The LOD of NH_4^+ was $< 0.0055 \mu\text{eq L}^{-1}$.

325 **3.2.4 nssCa²⁺**

326 The possible source of nssCa²⁺ is terrestrial dust (Dibb et al., 2007; Kuramoto et al., 2011). In surface snow, the maximum
327 value of nssCa²⁺ occurred at St. 2 (Fig. S4e). Positive peak values of nssCa²⁺ were exhibited at depths of 0.89, 1.07, and 1.53
328 m corresponding to early and late spring–summer period in 2021 and 2022 (Fig. 6). According to the stratigraphy of the
329 snowpack at St. 9 (Fig. S2), we attributed the peaks at 0.89, 1.07, and 1.53 m to the deposition of atmospheric nssCa²⁺, without
330 the influence of melting and refreezing process. In the inland areas of the Greenland Ice Sheet, mineral dust, which originates
331 mainly from Asia and North Africa (hereafter referred to as remote dust), was observed in spring layers in snowpacks or ice
332 cores (Steffensen et al., 1988; Whitlow et al., 1992; Mosher et al., 1993; Drab et al., 2002; Kuramoto et al., 2011; Nakazawa
333 et al., 2021; Nagatsuka et al., 2021). The early spring–summer peaks of nssCa²⁺ in the snowpack at St. 9 could have originated
334 from remote dust. However, the late spring–summer peaks of nssCa²⁺ in the snowpack at St. 9 have not been observed in the
335 inland of the Greenland Ice Sheet. Recently, the decreasing snow cover area in the Arctic associated with Arctic warming,
336 which is processing at a rate two to four times faster than the global average, has increased mineral dust emissions from bare
337 land in the Arctic during the summer–autumn period (hereafter referred to as local dust) (Amino et al., 2020; Matsui et al.,
338 2024). We assumed that the late spring–summer peaks of nssCa²⁺ in the snowpack at St. 9 could be attributed to local dust on
339 the northwestern coast of Greenland.

340



341
 342 **Figure 6: Vertical profile of nssCa^{2+} at (a) St. 3 and (b) St. 9.** Green lines denote mean nssCa^{2+} across all observation
 343 depths. Orange and brown lines denote the mean nssCa^{2+} plus and minus one standard deviation across all observation depths,
 344 respectively. The black and red curves denote the total concentration and non-sea-salt (nss) fractions, respectively. The gray
 345 dashed lines in (a) and (b) denote values of $0 \mu\text{eq. L}^{-1}$. The LOD of Ca^{2+} was $< 0.0050 \mu\text{eq L}^{-1}$.

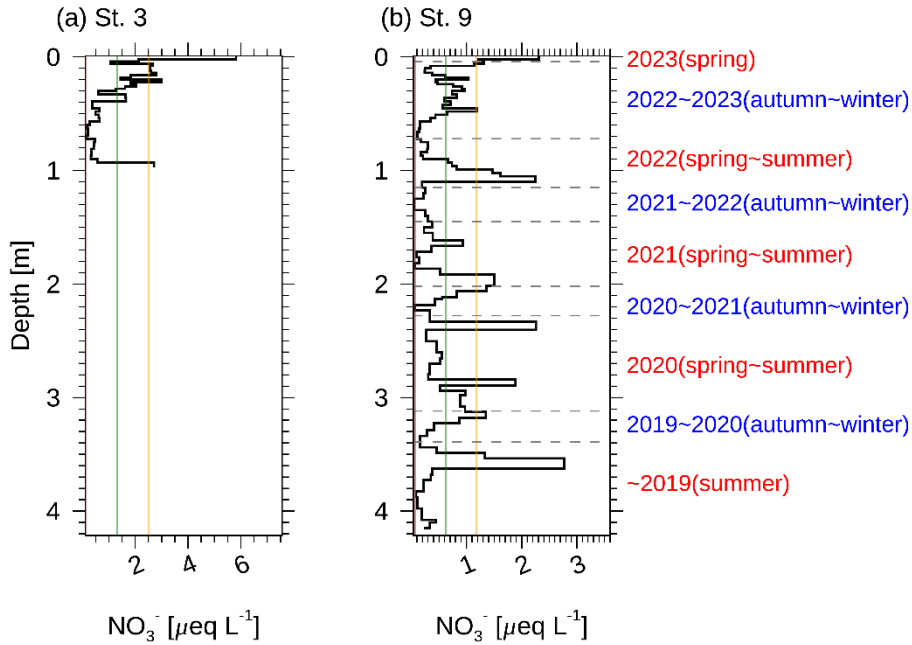
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347

348 3.2.5 NO_3^-

349 The possible sources of NO_3^- are fossil fuel combustion, biogenic soil emission, biomass burning, and photochemical
 350 reactions due to lightning (Legrand and Mayewski, 1997; Hastings et al., 2004). The concentration of NO_3^- in surface snow
 351 on Clements Glacier decreased with increasing distance toward the top of the glacier (Fig. S4h). The average concentrations
 352 of NO_3^- at depths from 0.00–1.01 m at St. 3 were greater than that at St. 9 (Table 1). The decreasing trend with distance from
 353 the seacoast was similar to that in the eastern region of Prudhoe Land (Matoba et al. 2014). The peak of NO_3^- at St. 3 was
 354 observed on the snow surface corresponding to the spring of 2023 (Fig. 7). The concentrations of NO_3^- at St. 9 showed several
 355 peaks at depths of 0.01, 1.07, 1.94, 2.37, 2.87, 3.15, and 3.57 m (Fig. 7). According to the stratigraphy of the snowpack at St.
 356 9 (Fig. S2), we attributed the peaks at 0.01, 1.07, 1.94 and 3.15 m to the deposition of atmospheric nitrate and the peaks at
 357 2.37, 2.87 and 3.57 m to melting and refreezing process. The NO_3^- tend to move easily with meltwater and become

358 concentrated during refreezing (Matoba et al., 2002b). Excluding peaks in the ice layers, the seasonal variations in the
 359 snowpack at St. 3 and St. 9 exhibited maximum values from late autumn–winter and early spring–summer and low values in
 360 late spring–summer. At several inland sites in Greenland, positive peaks of NO_3^- occurred in summer and from winter–spring
 361 (Kuramoto et al., 2011; Oyabu et al., 2016; Nakazawa et al., 2021). On the basis of the seasonal variation in the nitrogen and
 362 oxygen isotopes of NO_3^- in the snowpack at Summit, the source of NO_3^- in summer is natural NO_x produced by biomass
 363 burning, biogenic soil emissions, and photochemical reactions due to lightning, whereas that in winter is anthropogenic NO_x
 364 from fossil fuel combustion (Hastings et al., 2004). We assumed that the snowpack at St. 3 and St. 9 included aerosols
 365 originating from fossil fuel combustion and did not include aerosols originating from biomass burning, unlike the inland sites
 366 in Greenland.
 367

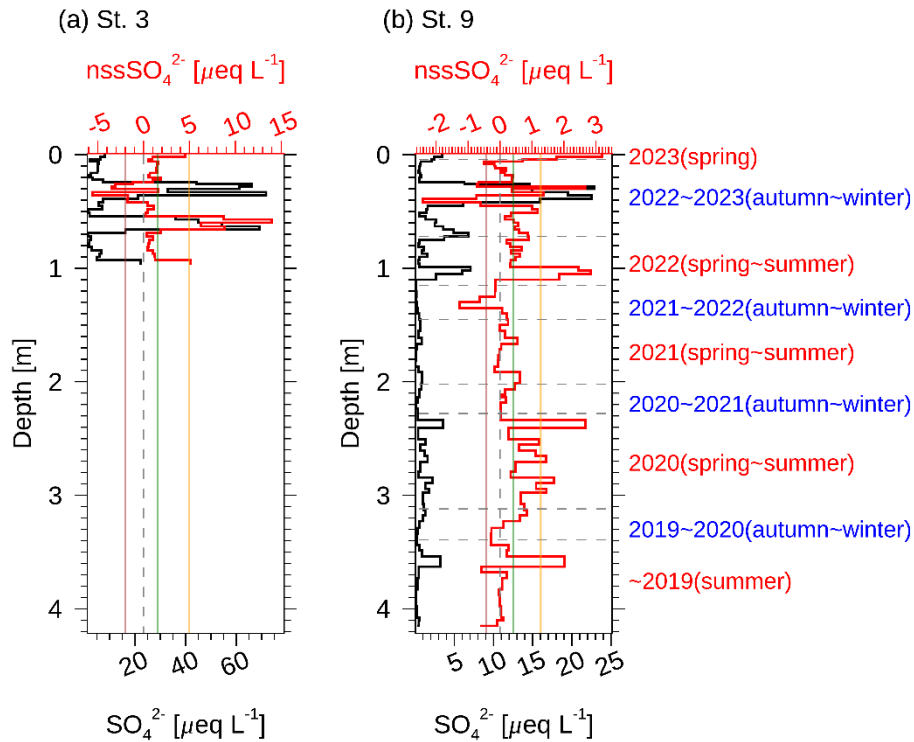


368
 369 **Figure 7: Vertical profile of NO_3^- at (a) St. 3 and (b) St. 9.** Green lines denote mean NO_3^- across all observation depths.
 370 Orange and brown lines denote the mean NO_3^- plus and minus one standard deviation across all observation depths,
 371 respectively. The LOD of NO_3^- was $< 0.0016 \mu\text{eq L}^{-1}$.
 372
 373

375 The possible sources of nssSO₄²⁻ are volcanic eruption, fossil fuel combustion, mineral dust, and DMS emissions produced
376 by marine phytoplankton (Jaffrezo et al., 1994; Legrand and Mayewski, 1997). The concentrations of nssSO₄²⁻ in surface snow
377 increased from St. 1 to St. 9 (Fig. S4i). The increasing trend with distance from the seacoast was similar to that in the eastern
378 region of Prudhoe Land (Matoba et al. 2014). The concentrations of nssSO₄²⁻ at St. 9 showed several positive peaks at depths
379 of 0.01, 1.04, 2.37, 2.68, 2.87, 2.96, and 3.57 m (Fig. 8). According to the stratigraphy of the snowpack at St. 9 (Fig. S2), we
380 attributed the peaks at 0.01, 1.04, 2.68 and 2.96 m to the deposition of atmospheric sulfate and the peaks at 2.37, 2.87 and 3.57
381 m to melting and refreezing process (Fig. 8). The seasonal variations in the snowpack at St. 3 and St. 9 peaked from late
382 autumn–winter and early spring–summer, excluding positive peaks in the ice layers. At several sites in Greenland, nssSO₄²⁻
383 exhibited maximum values from winter–spring, which is attributed to anthropogenic fossil fuel combustion (Dibb et al., 2007;
384 Kuramoto et al., 2011; Oyabu et al., 2016). Therefore, we believe that the source of positive nssSO₄²⁻ peaks in winter and
385 spring at St. 9 could be fossil fuel combustion, such as that at the inland sites in Greenland.

386 Several negative nssSO₄²⁻ peaks were found from autumn–winter in the snowpack at St. 3 and St. 9 (Fig. 8). In a coastal
387 Antarctic ice core, negative peaks have also been observed. The negative values of nssSO₄²⁻ in the snowpack result from the
388 low SO₄²⁻/Na⁺ in aerosols emitted from frost flowers on newly formed sea ice (Rankin et al., 2002; Woolf et al., 2003; Rankin
389 et al., 2004). Frost flowers are depleted in sulfate relative to sodium owing to the precipitation of mirabilite (Na₂SO₄·10H₂O)
390 during the formation of sea ice at temperatures below –8 °C (Richardson, 1976; Rankin et al., 2000; Rankin et al., 2002; Hara
391 et al., 2017). In northwestern Greenland, the frost flowers collected from newly formed sea ice in front of Siorapaluk indicated
392 depletion in sulfate relative to that in seawater (Hara et al., 2017). The negative nssSO₄²⁻ peaks observed at our study sites
393 could be attributed to the atmospheric emission of frost flowers from newly formed sea ice in the NOW polynya, where new
394 sea ice is often created and broken up by strong northerly winds in winter.

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403 3.2.7 MSA

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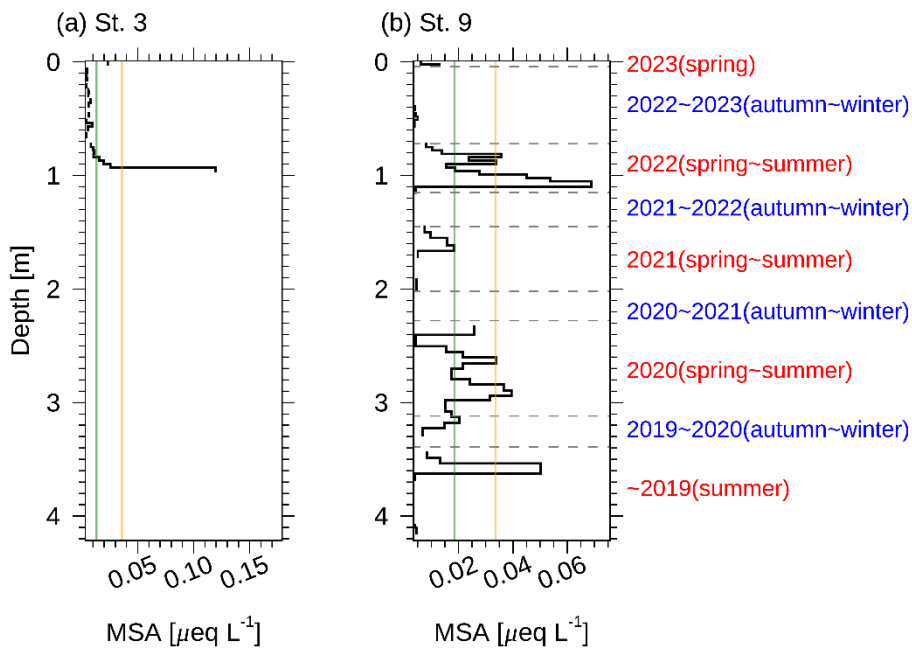
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412

Figure 8: Vertical profile of nssSO_4^{2-} at (a) St. 3 and (b) St. 9. Green lines denote mean nssSO_4^{2-} across all observation depths. Orange and brown lines denote the mean nssSO_4^{2-} plus and minus one standard deviation across all observation depths, respectively. The black and red curves denote the total concentration and non-sea-salt (nss) fractions, respectively. The gray dashed lines in (a) and (b) denote values of $0 \mu\text{eq L}^{-1}$. The LOD of SO_4^{2-} was $< 0.0021 \mu\text{eq L}^{-1}$.

MSA is an oxidation product of DMS, which originates from ocean phytoplankton (Charlson et al., 1987; Jaffrezo et al., 1994). The concentration of MSA in surface snow reached a maximum value at St. 6 (Fig. S4j). At St. 3, an exceptionally high MSA was observed at a depth of 1.01 m, corresponding to the summer of 2022 (Fig. 9a). The snowpack at St. 9 exhibited multiple distinct MSA peaks at depths of 0.83, 1.07, 2.68, 2.91, and 3.57 m (Fig. 9b). According to the stratigraphy of the snowpack at St. 9 (Fig. S2), we attributed the peaks at 0.83, 1.07, and 2.68 m to the deposition of atmospheric MSA and the peaks at 2.91 and 3.57 m to melting and refreezing processes. The positive MSA peak at 1.07 m coincided with the depth of the positive NO_3^- and nssSO_4^{2-} peaks. This finding indicates that MSA originating from ocean phytoplankton activity at middle and low latitudes could have been transported to St. 9 with anthropogenic products, including NO_3^- and nssSO_4^{2-} . In the snowpack at depths of 0.83 and 2.63 m corresponding to spring–summer layers, MSA and/or nssSO_4^{2-} showed positive peaks,

413 whereas NO_3^- did not indicate any peaks. Because NO_3^- peaks did not occur, the air masses containing MSA were not
 414 transported from mid-latitudes, which are areas of human activity. Therefore, the positive peaks of MSA at 0.83 and 2.63 m
 415 could be reflected in sea-to-air DMS emission attributed to the ocean phytoplankton bloom in the NOW polynya.
 416



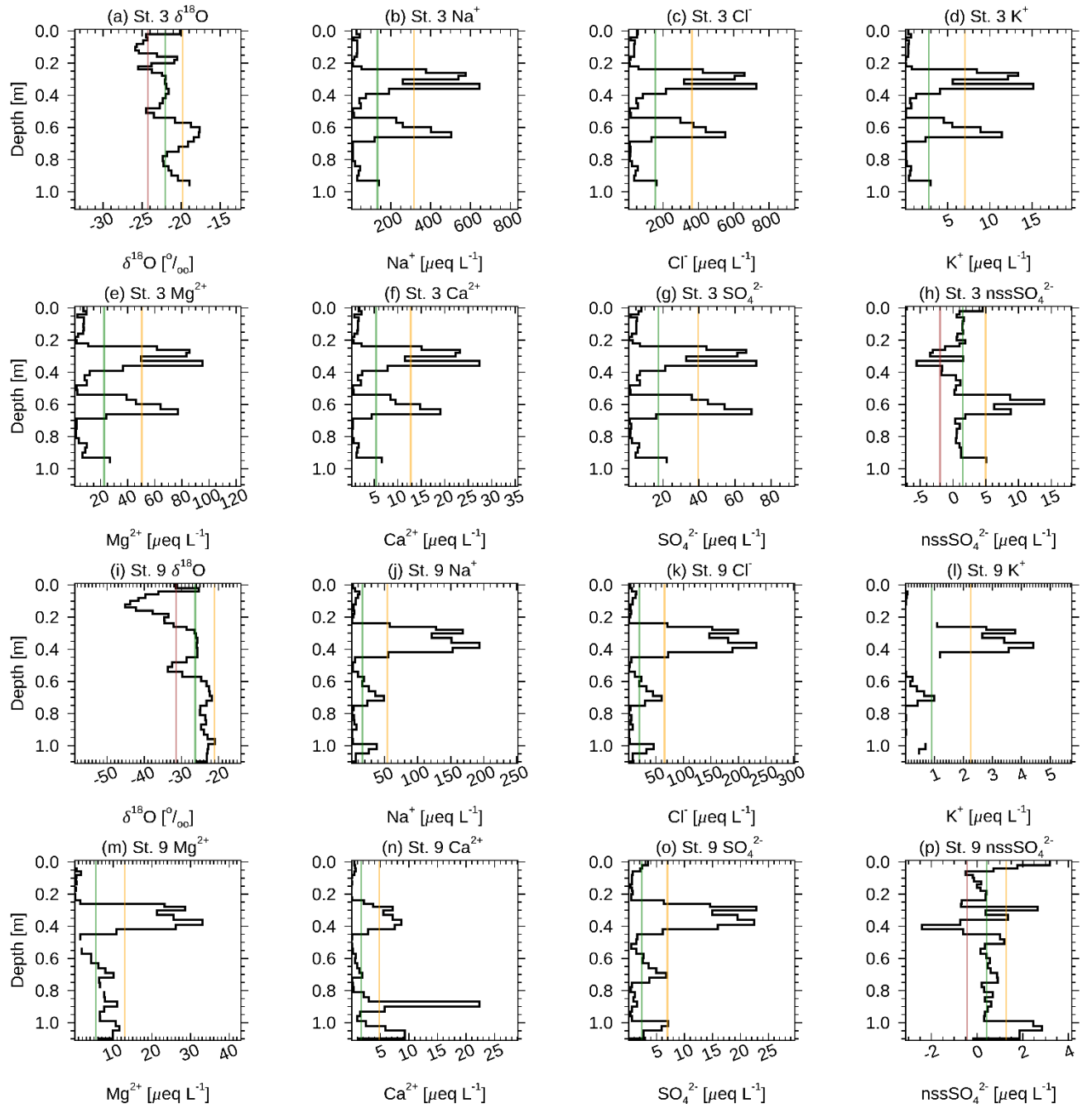
417
 418 **Figure 9: Vertical profile of nssSO_4^{2-} at (a) St. 3 and (b) St. 9.** Green lines denote mean MSA across all observation
 419 depths. Orange and brown lines denote the mean MSA plus and minus one standard deviation across all observation depths,
 420 respectively. The LOD of MSA was $< 0.0011 \mu\text{eq L}^{-1}$.
 421
 422

424 **3.3 Intense poleward heat and moisture transport events in winter from 2022–2023**

425 The concentrations of $\delta^{18}\text{O}$, Na^+ , Cl^- , K^+ , Mg^{2+} , SO_4^{2-} , and Ca^{2+} increased substantially and the nssSO_4^{2-} exhibited negative
426 values at depths from 0.26–0.42 m at St. 3 and St. 9, corresponding to autumn–winter from 2022–2023 (Fig. 10). According
427 to the stratigraphy of the snowpack at St. 3 and St. 9, the peaks of $\delta^{18}\text{O}$ and ion concentrations observed from 0.26–0.42 m had
428 not been changed by the melting and refreezing process. Because the depth of the positive peaks of $\delta^{18}\text{O}$ and these ions
429 coincided between St. 3 and St. 9, we believe that the snowpacks from 0.26–0.42 m originated from the same precipitation
430 event during the autumn–winter period from 2022–2023. In previous observations of precipitation isotopes in Siorapaluk, the
431 $\delta^{18}\text{O}$ value in precipitation was significantly correlated with the surface air temperature from winter–spring (Matoba et al.,
432 2002a). Therefore, the rapid positive increase in $\delta^{18}\text{O}$ at depths from 0.26–0.42 m at St. 3 and St. 9 suggested that the air
433 temperature at our study sites increased during low temperature period, leading to the coldest winter season. Continuous
434 weather monitoring in Siorapaluk revealed that the air temperature increased by more than 0 °C and that the southern
435 component of the wind increased from 5–6 December 2023 (Fig. 11a and c). During this period, the atmospheric pressure
436 gradient at 500 hPa increased along Baffin Bay due to high pressure in Greenland and low pressure in northern Canada, and
437 the southerly wind speed along the western coast of Greenland increased during this period (Fig. 11d and Fig. 12).
438 Consequently, we determined that the rapid increase in $\delta^{18}\text{O}$ at depths from 0.26–0.42 m corresponded to the intense poleward
439 heat and moisture transport event from 5–6 December 2023. The Na^+ , Cl^- , K^+ , Mg^{2+} , SO_4^{2-} , and Ca^{2+} concentrations increased
440 substantially, and nssSO_4^{2-} exhibited negative values at depths from 0.26–0.42 m at St. 3 and St. 9 (Fig. 10). The Na^+ , Cl^- , K^+ ,
441 Mg^{2+} , SO_4^{2-} , and Ca^{2+} concentrations in frost flowers are comparable to or greater than those in seawater, and the $\text{SO}_4^{2-}/\text{Na}^+$
442 ratio of frost flowers is lower than that of seawater (Hara et al., 2017). On the eastern side of the NOW, the southern component
443 of the wind at 2 m a.g.l. increased from 5–6 December 2023, whereas the sea ice concentration remained high (Fig. 11a and
444 c). The increase in the surface wind speed across the sea ice in the NOW from 5–6 December 2023 promoted the atmospheric
445 emission of frost flowers from sea ice, after which the Na^+ , Cl^- , K^+ , Mg^{2+} , SO_4^{2-} , and Ca^{2+} concentrations increased, and
446 nssSO_4^{2-} exhibited negative values at depths from 0.26–0.42 m at St. 3 and St. 9. We suggest that intense poleward heat and
447 moisture transport events attributed to the high pressure in Greenland and low pressure in northern Canada caused precipitation,
448 including heavier isotopes and frost flowers on the northwestern coast of Greenland Ice Sheet.

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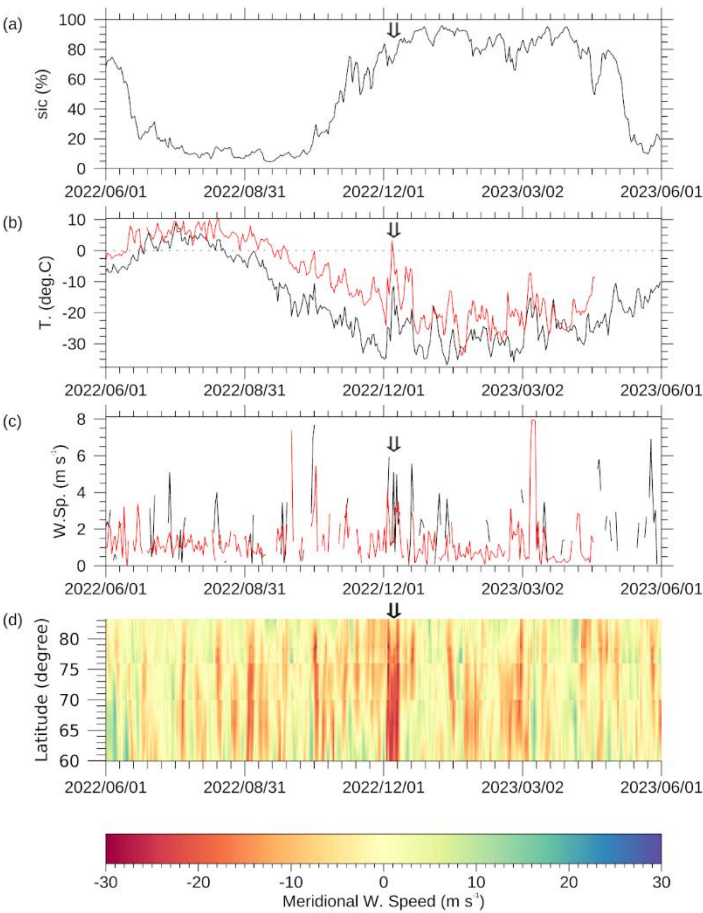
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451

452 **Figure 10: Vertical profiles of $\delta^{18}\text{O}$, Na^+ , Cl^- , K^+ , Mg^{2+} , Ca^{2+} , SO_4^{2-} , and nssSO_4^{2-} at St. 3 and St. 9. Green lines denote**
 453 **mean ion concentrations across all observation depths. Orange and brown lines denote the mean ion concentrations plus and**
 454 **minus one standard deviation across all observation depths, respectively.**

455



457

458 **Figure 11: Sea ice and meteorological conditions from April 2022 to April 2023 corresponding to snow depths ranging**
459 **from 0.00–1.15 m at St. 9.** (a) shows the sea ice concentration on the eastern side of NOW. (b) shows the temperature at 2 m
460 a.g.l. in Siorapaluk (red line) and St. 9 (black line). (c) shows the southern component of the wind speed at 10 m a.g.l. in
461 Siorapaluk (red line) and on the eastern side of the NOW (black line). (d) shows the meridional wind speed at 700 hPa along
462 the eastern side of Baffin Bay. The values on the eastern side of the NOW were averaged over 75.0°–80.0°N and 75.0°–65.0°W.
463 The values on the eastern side of Baffin Bay were averaged over 75.0°–65.0°W from 75.0° to 80.0°N, 75.0°–65.0°W from
464 75.0° to 80.0°N, 75.0°–65.0°W from 75.0° to 80.0°N and 75.0°–65.0°W from 75.0° to 80.0°N. The positive and negative
465 values of the meridional wind speed are the southern and northern parts of the wind, respectively. The black arrows denote the
466 date from 5–6 December 2023.

467

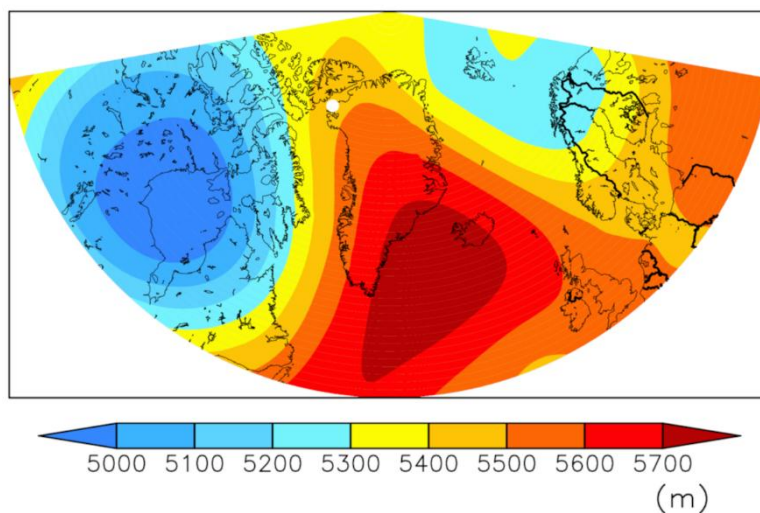


Figure 12: Geopotential height at 500 hPa averaged from 5 to 6 December 2022. The white circle indicates the location of St.9.

4 Conclusion

We conducted glaciological observations from 9–11 April 2023 on the western side of Prudhoe Land in northwestern Greenland facing the NOW to elucidate the source conditions and transportation processes of water vapor and aerosols in this region. The dating of the snowpack at St. 9, which is located at the inland of the western side of Prudhoe Land, revealed that the layer at a depth of 4.20 m corresponded to 3.5 years. The average annual accumulation at St. 9 was $0.49 \text{ m w.eq. yr}^{-1}$.

The snowpacks on the western side of Prudhoe Land contained aerosols from distant sources, such as remote dust and anthropogenic aerosols, in early spring–summer layers. On the other hand, they also contained aerosols from local sources such as ocean biological activity and frost flowers in the NOW and local dust around the coast of northwestern Greenland during other seasons, unlike the inland of the Greenland Ice Sheet. Moreover, we noted that the snowpacks were able to trace the poleward heat and moisture transport event along Baffin Bay during winter.

Arctic climate warming caused decreases in the sea ice thickness and concentration over the last few decades in the NOW and could influence clouds and precipitation following changes in sea ice and biological activities in the NOW. We found for the first time that the environmental changes in the NOW can be elucidated by the snowpack and ice core on the western side of the Prudhoe Land. We suggest that the chemical substances in the deeper ice core from this region could help explain the multidecadal variations in the sea ice, biological activities, and related water and aerosol circulation around the NOW and could develop to understand the accurate future projections of environmental change in this region.

488 **Code availability**

489 Data analyzes were performed using Interactive Data Language (IDL) (version 8.7.1) developed by Exelis Visual
490 Information Solutions, Boulder, Colorado. Map figures were generated using Generic-Mapping Tools (GMT) (version 6.1.0)
491 (<https://www.generic-mapping-tools.org/>) and Quantum Geographic Information System (QGIS) (<https://qgis.org/>).

492 **Data availability**

493 The data used in this are available from the Hokkaido University Collection of Scholarly and Academic Papers (HUSCAP)
494 (<http://hdl.handle.net/2115/94742>). The AWS data are available from the Arctic Data archive System (ADS)
495 (<https://ads.nipr.ac.jp/data/meta/A20241031-013>). The ERA5 and ERA5-Land data are available from the Copernicus Climate
496 Change Service (C3S) Climate Data Store (<https://cds.climate.copernicus.eu#!/search?text=ERA5&type=dataset>). The
497 Landsat-8 image is provided by United States Geological Survey (<https://earthexplorer.usgs.gov/>). The Greenland Mapping
498 Project 2 (GIMP-2) digital elevation model version 2 is provided by the National Snow and Ice Data Center (NSIDC)
499 (<https://nsidc.org/data/nsidc-0715/versions/2>).

500 **Author contributions**

501 Y.K. and S.M. designed this study. Y.K., M.M, T.K, and H.I. collected and processed the snow samples during the field
502 observations. Y.K. and S.M. analyzed water stable isotopes and chemical substances in the snow samples. Y.K. analyzed the
503 meteorological and sea ice data. Y.K. and S.M. wrote the manuscript draft. Y.I. revise the manuscript.

504 **Competing interests**

505 The contact author has declared that none of the authors has any competing interests.

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513 and (3) the National Institute of Polar Research through Special Collaboration Project no. B24-02.

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