

1           Holocene glacier activity reconstructed from proglacial  
2           lake Gjøavatnet on Amsterdamøya, NW Svalbard  
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37    **Keywords:**

38           Holocene Paleoclimate, Svalbard, Lake Sediments, Glaciers, Fram Strait, Freshwater Forcing  
39

40 **Abstract**

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42 Well-dated and highly resolved paleoclimate records from high latitudes allow for a better  
43 understanding of past climate change. Lake sediments are excellent archives of environmental change,  
44 and can record processes occurring within the catchment, such as the growth or demise of an upstream  
45 glacier. Here we present a Holocene-length, multi-proxy lake sediment record from proglacial lake  
46 Gjøavatnet on the island of Amsterdamøya, northwest Svalbard. Today, Gjøavatnet receives meltwater  
47 from the Annabreen glacier and contains a record of changes in glacier activity linked to regional climate  
48 conditions. We measured changes in organic matter content, dry bulk density, bulk carbon isotopes,  
49 elemental concentrations via Itrax core-scanning, and diatom community composition to reconstruct  
50 variability in glacier extent back through time. Our reconstruction indicates that glacially derived  
51 sedimentation in the lake decreased markedly at ~11.1 cal kyr BP, although a glacier likely persisted in  
52 the catchment until ~8.4 cal kyr BP. During the mid-Holocene (~8.4-1.0 cal kyr BP) there was  
53 significantly limited glacial influence in the catchment and enhanced deposition of organic-rich sediment  
54 in the lake. The deposition of organic rich sediments during this time was interrupted by at least three  
55 multi-centennial intervals of reduced organic matter accumulation (~5.9-5.0, 2.7-2.0, and 1.7-1.5 cal kyr  
56 BP). Considering the chronological information and a sedimentological comparison with intervals of  
57 enhanced glacier input, we interpret these intervals not as glacial advances, but rather as cold/dry episodes  
58 that inhibited organic matter production in the lake and surrounding catchment. At ~1.0 cal kyr BP, input  
59 of glacially derived sediment to Gjøavatnet abruptly increased, representing the rapid expansion of the  
60 Annabreen glacier.

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65 **1) Introduction**

66           Reliable and highly resolved paleoclimate reconstructions are necessary in order to better  
67 contextualize recent and predicted climate change (Kaplan and Wolfe, 2006), especially from high  
68 latitudes where these changes are expected to be greatest (Callaghan et al., 2010; Serreze and Barry,  
69 2011; Stocker et al., 2013). Glaciers and small ice caps respond rapidly to climate variations, but direct  
70 evidence for past changes in ice extent in remote areas is temporally and spatially limited. The Holocene  
71 Epoch (past ~11.7 thousand calendar years before present (cal kyr BP)) provides important context to  
72 understand Arctic climate dynamics because atmospheric and ocean circulation patterns were similar to  
73 their current configuration and natural insolation forcing caused widespread environmental change  
74 without the overprint of significant anthropogenic influence (until recent decades) (Johnsen et al., 2001;  
75 Kaufman et al., 2004; Mayewski et al., 2004). Broad scale Arctic climate throughout this interval is  
76 understood to have been characterized by early Holocene warmth that progressively transitioned toward a  
77 colder late Holocene (i.e., the Neoglacial), driven mainly by declining high latitude summer insolation  
78 (Laskar et al., 2004; Miller et al., 2010; Briner et al., 2016). Superimposed on this trend are spatial  
79 heterogeneities linked to the decaying Northern Hemisphere ice sheets and associated meltwater pulses  
80 (e.g. Sejrup et al., 2016). High-resolution paleoclimate records are important to better examine the timing,  
81 expression, and magnitude of Holocene climate change throughout the Arctic.

82           The Svalbard archipelago is uniquely situated at the intersection between the northern North  
83 Atlantic and the Arctic Ocean basins (**Figure 1**). Near Svalbard, warm Atlantic Water transported via the  
84 West Spitsbergen Current (WSC) mixes with cold, less saline water from the Arctic Ocean. Atmospheric  
85 conditions are influenced by the relative influence of cold, polar-derived air masses from the north and  
86 east and warmer subpolar maritime air masses from the south and south west (Førland et al., 2011). The  
87 relative locations of these important boundaries, along with associated sea ice feedbacks, has been shown  
88 to vary throughout the Holocene and influence climate in Svalbard (e.g Müller et al., 2012; Rasmussen et  
89 al., 2014; Werner et al., 2013, 2015).

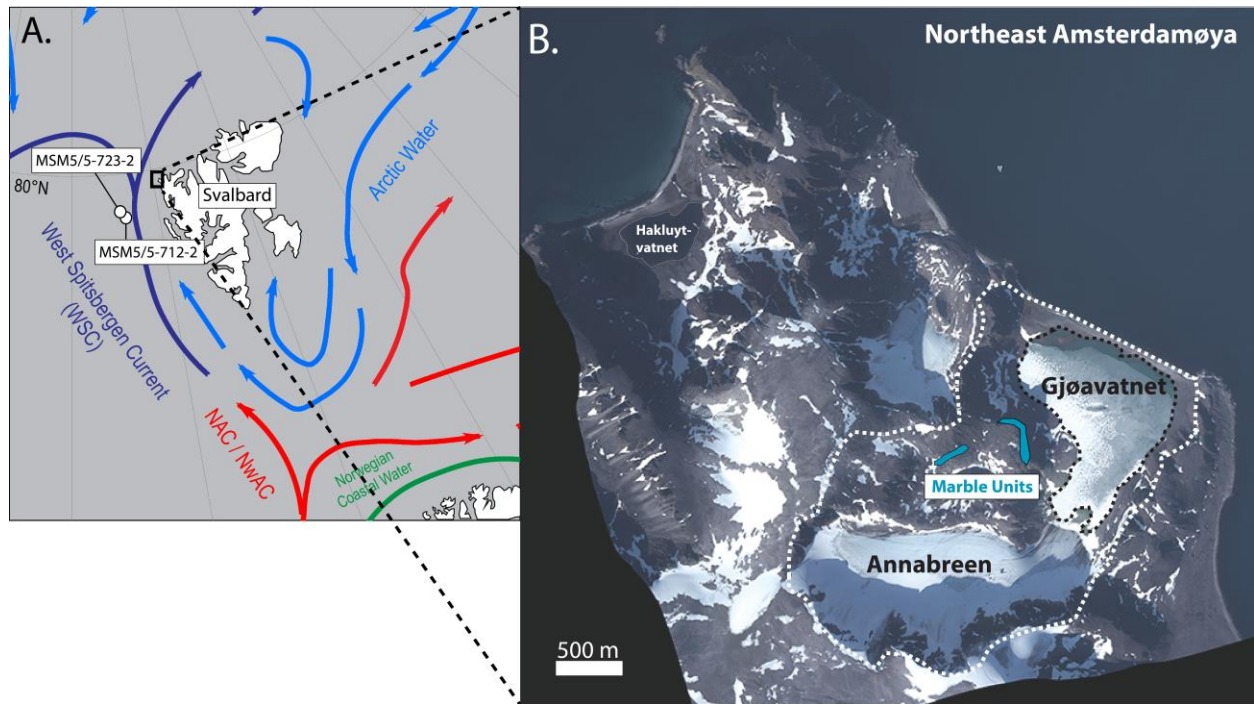
90           Broadly, evidence from marine records near Svalbard and in the northern North Atlantic suggests  
91 a warm early Holocene period (~11-8 kyr BP), characterized by an increased flux of Atlantic water to

92 high latitudes (Aagaard-Sørensen et al., 2014; Forwick and Vorren, 2009; Hald et al., 2004, 2007; Müller  
93 et al., 2012; Rasmussen et al., 2014; Risebrobakken et al., 2011; Sarnthein et al., 2003; Skirbekk et al.,  
94 2010; Ślubowska et al., 2005; Werner et al., 2013, 2015). The transition to cooler (Neoglacial) conditions  
95 during the middle Holocene is not well constrained, with some records suggesting cooling began as early  
96 8.8 cal kyr BP (e.g. Hald et al., 2004), while others point to cooling beginning ~6 kyr BP (e.g. Rasmussen  
97 et al., 2014). Oceanic conditions near Svalbard during the late Holocene are generally characterized by  
98 cold temperatures overprinted by fluctuations linked to changes in the advection of warm Atlantic Water  
99 in the WSC (e.g. Aagaard-Sørensen et al., 2014; Berben et al., 2014; Ślubowska et al., 2005; Werner et  
100 al., 2013).

101           While there has been an increase in the number of terrestrial Holocene paleoclimate studies from  
102 Svalbard in recent years (e.g. D'Andrea et al., 2012; Reusche et al., 2014; Røthe et al., 2015; van der Bilt  
103 et al., 2015; van der Bilt et al., 2016; Gjerde et al., *in press*; Balascio et al., *in press*), this region generally  
104 lacks continuous, well-dated paleoenvironmental reconstructions. Retreat of the Barents Sea ice sheet had  
105 begun during the end of the Pleistocene (~20-14 cal kyr BP) (Gjermundsen et al., 2013; Hormes et al.,  
106 2013; Ingólfsson and Landvik, 2013), and numerous records suggest many smaller glaciers in Svalbard  
107 retreated or completely melted during the early Holocene (Reusche et al., 2014; Røthe et al., 2015; Snyder  
108 et al., 2000; Svendsen and Mangerud, 1997; van der Bilt et al., 2015). The timing of late-Holocene glacial  
109 re-advance however, remains poorly constrained; some studies suggest glaciers advanced ~3-4 cal kyr BP  
110 (Reusche et al., 2014; Røthe et al., 2015; Svendsen and Mangerud, 1997), while others point to a later  
111 advance closer to 1 cal kyr BP (Humlum et al., 2005; Snyder et al., 2000; van der Bilt et al., 2015).

112           Here we present a glacier reconstruction from proglacial lake Gjøvatnet on the island of  
113 Amsterdamøya in NW Svalbard spanning the Holocene (**Figure 1**). The objective of this study is to  
114 reconstruct the history of the upstream Annabreen glacier (**Figure 1**) towards a better understanding of  
115 Holocene climate change in the region. The maritime climate of Amsterdamøya, and the presence of a  
116 glacier in the Gjøvatnet catchment, makes this record valuable for addressing questions about past  
117 climate variations in the High Arctic North Atlantic. Our record suggests that Annabreen disappeared or

118 was dramatically reduced in size by ~8.4 cal kyr BP. During the interval 8.4 – 1 cal kyr BP, sedimentation  
119 in the lake was marked by higher amounts of organic material accumulation, but was punctuated by  
120 multi-centennial-length periods with relatively lower organic matter content. A return to minerogenic  
121 sedimentation at 1 cal kyr BP is interpreted to represent the re-advance of Annabreen at that time.



122  
123 **Figure 1:** A) Map of Svalbard and surrounding surface currents as well as locations of marine sediment  
124 cores MSM5/5-723-2 and MSM5/5-712-2 (Werner et al., 2013; 2015); B) Aerial image of island of  
125 Amsterdamøya with Annabreen glacier and Gjøavatnet lake (this study) and Hakluytvatnet lake (Gjerde  
126 et al., *in press*) labeled. Blue polygons represent the approximate locations of marble outcrops (Ohta et  
127 al., 2007). Dashed white line denotes approximate catchment of Gjøavatnet.

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130 **2) Regional Setting**

131 Svalbard's climate is characterized by highly variable temperatures and low average annual  
132 precipitation. A meteorological station at Ny-Ålesund, ~95 km south of Amsterdamøya, recorded average  
133 annual temperatures of -5.2°C over the period 1981-2010, though average temperatures of individual  
134 years ranged from -12°C to 3.8°C (Førland et al., 2011). The average annual precipitation at Ny-Ålesund  
135 is 427 mm (Førland et al., 2011), with the majority of moisture sourced from the south/southwest and  
136 occurring during the fall and winter months (Førland et al., 2011). The dry nature of this environment

137 suggests that relatively small changes in the precipitation budget could play a large role in the mass  
138 balance of local glaciers. There is a positive relationship between temperature and precipitation in all  
139 seasons (Førland et al., 2011).

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## 141 **Study Site**

142 Lake Gjøavatnet (79°46'00"N, 10°51'45"E, 2 m a.s.l.) is a small proglacial lake on the eastern  
143 coast of Amsterdamøya, Svalbard (**Figure 1**). The lake catchment contains a steep-sided cirque (area of  
144 ~2.8 km<sup>2</sup>), marked by the presence of the Annabreen glacier (surface area of 0.87 km<sup>2</sup>), which currently  
145 terminates in the southwest corner of the lake. The limited vegetation in the catchment is characterized as  
146 northern Arctic tundra (Birks et al., 2004). Gjøavatnet itself has an area of ~0.52 km<sup>2</sup> and has a maximum  
147 depth of ~15m. The lake is separated from the ocean by a narrow (~30 m wide) strip of bedrock (Hjelle  
148 and Ohta, 1974; Ohta et al., 2007), which is currently covered with large numbers of smoothed boulders  
149 and driftwood (**Figure 1**). There has been no net postglacial uplift relative to sea level in the immediate  
150 region during the Holocene (Forman, 1990; Landvik et al., 1998; 2003). The surrounding bedrock is  
151 comprised mainly of Mesoproterozoic banded gneiss and migmatite of the Smeerenburgfjorden Complex  
152 containing small outcrops of marble with skarn mineralization in the northwestern part of the catchment  
153 (blue outlines in **Figure 1**) (Hjelle and Ohta, 1974; Ohta et al., 2007).

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## 156 **3) Methods**

### 157 **3.1) Fieldwork and Lake Coring**

158 Prior to coring, Gjøavatnet was surveyed using ground penetrating radar (GPR) as well as a  
159 Lowrance sonar bathymetric device to determine lake bottom bathymetry and soft sediment distribution  
160 (**Figure 2**). GPR profiles were collected using a Mala RAMAC GPR unit with a 50 MHz antenna.

161 Five sediment cores were collected from Gjøavatnet in the summer of 2014; 2 piston cores (GJP-  
162 01-14; 210 cm in length, and GJP-02-14; 82 cm), and 3 gravity/surface cores (GJD-01-14; 39 cm, GJD-  
163 02-14; 46 cm, and GJD-03-14; 42 cm) (**Table 1**) (see **Figure 2** for coring locations). Surface cores were

164 collected using a Uwitec surface corer. Piston cores were collected using a percussion piston coring  
165 device and were hammered until refusal to ensure maximum sediment recovery. GJP-01-14 was cut into  
166 two sections in the field (1 of 2: 135.5 cm long, 2 of 2: 71 cm long) to allow for transport. The sediment  
167 cores were then shipped back to the University of Bergen for splitting and analysis.

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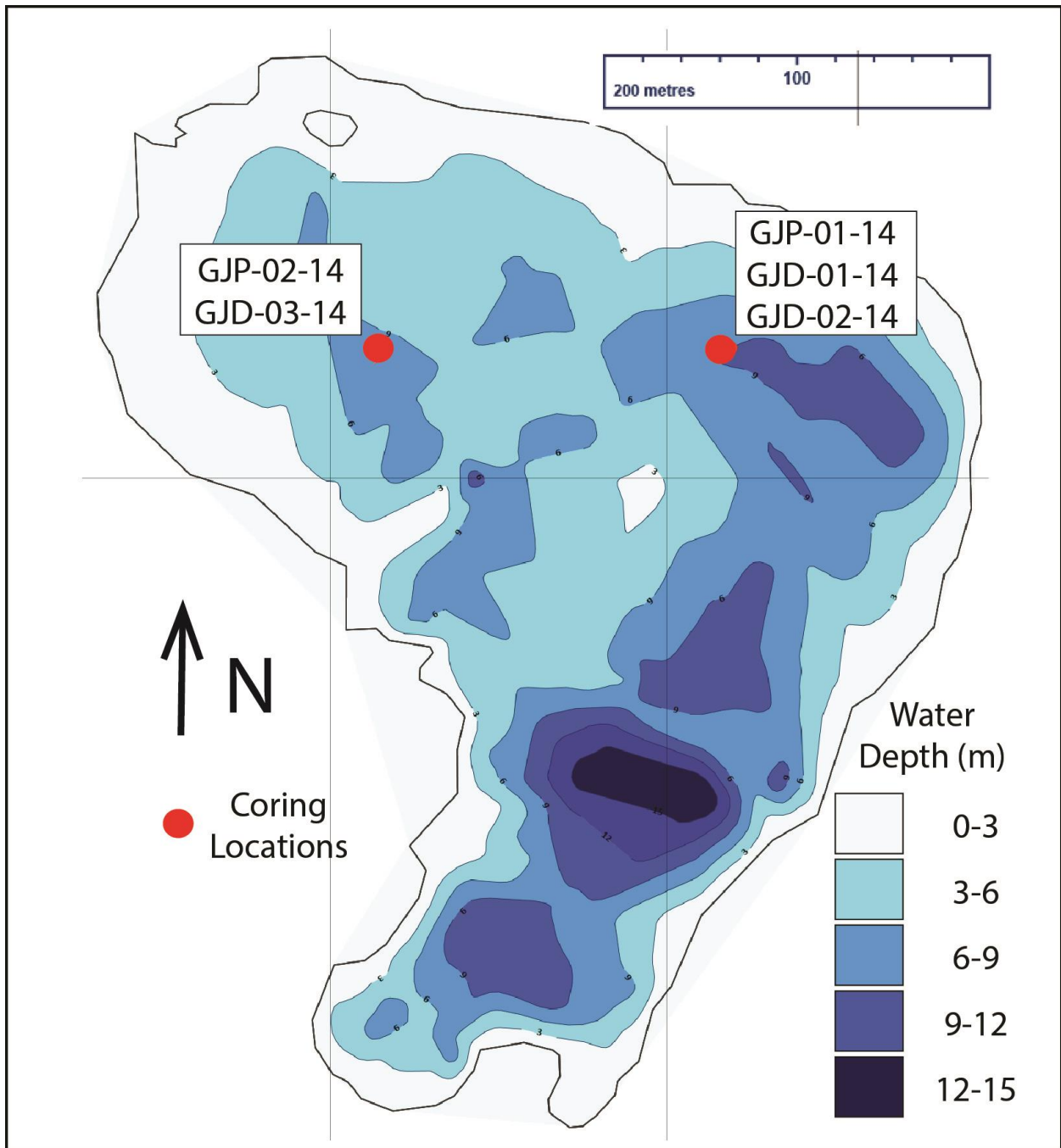
**Table 1: Sediment cores collected from Gjøvatnet**

Core Name	GJP-01-14	GJP-02-14	GJD-01-14	GJD-02-14	GJD-03-14
Core Type	Piston	Piston	Surface/Gravity	Surface/Gravity	Surface/Gravity
Core Length (cm)	206.5	95	30	43	42

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### 3.2) Composite sediment record

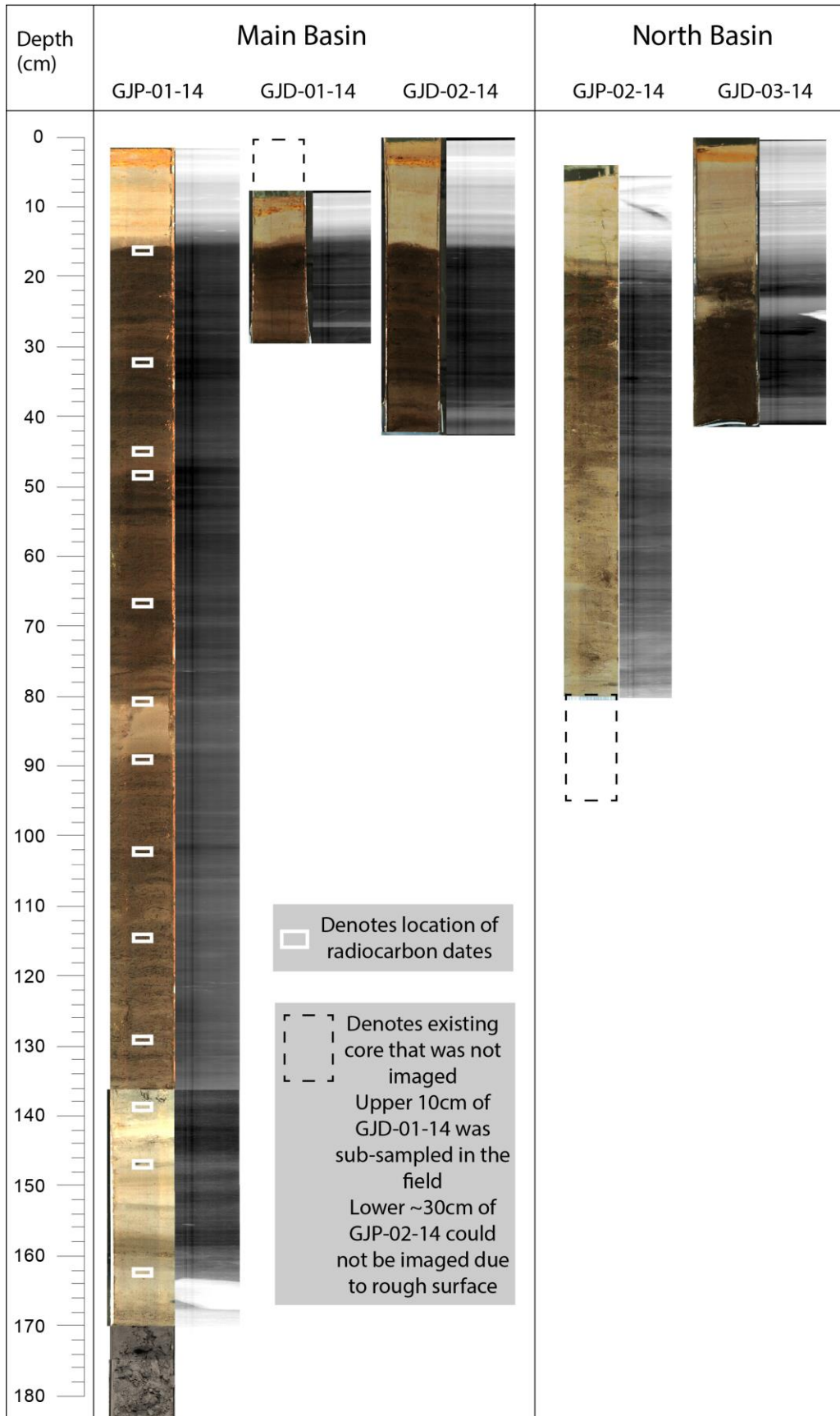
173 A composite sediment record was created based mainly on piston core GJP-01-14, which was  
174 collected from the deepest part of the lake (**Figure 2**) and is the longest core retrieved from Gjøvatnet  
175 (**Figure 3**). Comparison of both visual stratigraphy and proxy data between piston core GJP-01-14 and  
176 surface cores GJD-01-14 and GJD-02-14 (recovered from the same basin) reveal that only 2.5 cm of  
177 sediment were lost from the upper part of the GJP-01-14 during piston coring. The basal sediment in both  
178 piston cores (GJP-01-14 and GJP-02-14) is comprised of diamict, interpreted as glacial till, suggesting the  
179 entire Holocene record was recovered (**Figure 3**). To achieve the highest resolution record possible we  
180 focused our investigation on the GJP-01-14 core (206.5 cm length). Results from this core form the basis  
181 of most of our interpretations.



182

183 **Figure 2:** Bathymetric map of Gjøvatnet with coring sites noted by red circles. Our investigation  
 184 focused mainly on core GJP-01-14 from eastern basin.  
 185





187 **Figure 3:** Line-scan images and x-radiographs of all cores collected from Gjøavatnet. GJP-01-14 was the  
188 focus of this investigation. Depth of radiocarbon dates are denoted with white rectangles. Dashed black  
189 lines represent existing core that could not be imaged. Lighter shades in x-radiographs represent denser  
190 material. Note that color difference in lowest portion of GJP-01-14 (section from ~170-180 cm) is due to  
191 simply to different lighting during imaging.

### 192 **3.3) Laboratory Analyses**

193 Cores were split and imaged at the University of Bergen prior to analyses. All five cores were  
194 analyzed for surface magnetic susceptibility (MS) at 0.5 cm resolution using a Bartington MS2E point  
195 sensor. The cores were also analyzed using an ITRAX X-ray fluorescence (XRF) core scanner located at  
196 EARTHLAB, University of Bergen, to determine elemental concentrations. Scans were carried out using  
197 a molybdenum (Mo) tube with a downcore resolution of 200  $\mu\text{m}$ . The voltage and current were set to  
198 30kV and 45mA respectively, with an XRF count time of 10 seconds. For GJP-01-14, Itrax core scanning  
199 data were not collected below 167cm depth because the sediment surface was too uneven for the  
200 instrument to accommodate.

201 GJP-01-14 was sub-sampled at 0.5 cm intervals for weight loss-on-ignition (LOI), dry bulk  
202 density (DBD), and water content (WC) ( $n=335$ ) (after Dean, 1974; Heiri et al., 2001). A syringe was  
203 used to ensure a constant 1  $\text{cm}^3$  of sediment was removed. The sediment below ~167cm in GJP-01-14  
204 was either too stiff to properly remove the necessary volume of sediment for accurate LOI and DBD  
205 analysis or was simply composed of large clasts, making the analyses impossible. Accordingly, the  
206 majority of our proxy data (and associated figures) do not include data from the bottom ~40 cm of the  
207 GJP-01-14 core.

208 Macrofossils for radiocarbon dating were removed and sent to the Poznan Radiocarbon  
209 Laboratory in Poland for analysis (**Table 2**).  $\delta^{13}\text{C}$  values of bulk sediment ( $n=50$ ) were measured at the  
210 Lamont Doherty Earth Observatory using a Costech elemental combustion system (EA) coupled to a  
211 Delta V Plus IRMS (Thermo). A two-point isotope calibration curve was constructed using standards  
212 USGS40 and USGS41 to place measured values on the VPDB scale. A third standard (USGS24) was run  
213 during the period of data acquisition to evaluate the accuracy of measurement.

214 Eight samples were chosen for diatom taxonomic analysis from 121.5, 135.5, 137, 149.5, 161.5,  
215 163.5, 164.5, and 165.5 cm composite depth in GJP-01-14 to evaluate the possibility of early Holocene  
216 marine incursions and to characterize lacustrine conditions. Diatoms were isolated from the sediments  
217 using standard oxidative techniques modified from Renberg (1990) and mounted on glass coverslips  
218 using Naphrax mounting medium. At least 300 diatom samples were identified from each slide at 1000x  
219 under oil immersion and identified using predominantly Arctic diatom floras (e.g. Antoniades et al.,  
220 2008).

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### 222 **3.4) Statistical and Multivariate Analyses**

223 Principal component analysis (PCA) and computation of correlation coefficients were carried out  
224 on 10 measured proxies using Matlab software for Windows. This included 9 geochemical element counts  
225 (Ti, K, Ca, Rb, Sr, Fe, Mn, Si, Al) from the Itrax core scanner as well as %LOI. These elements were  
226 selected based on their high signal response on the Itrax (counts per second generally >100), their  
227 prevalence in siliclastic sediments, and previous studies that have identified them as useful for  
228 reconstructing minerogenic input from bedrock erosion (Bakke et al., 2013; Balascio et al., 2015; Røthe et  
229 al., 2015). Itrax data were smoothed using a 24pt running mean and resampled at 0.5cm intervals to  
230 achieve comparable resolution to the LOI data. All datasets were also log-transformed prior to analysis.

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## 232 **4) Results and Interpretations**

### 233 **4.1) Chronology**

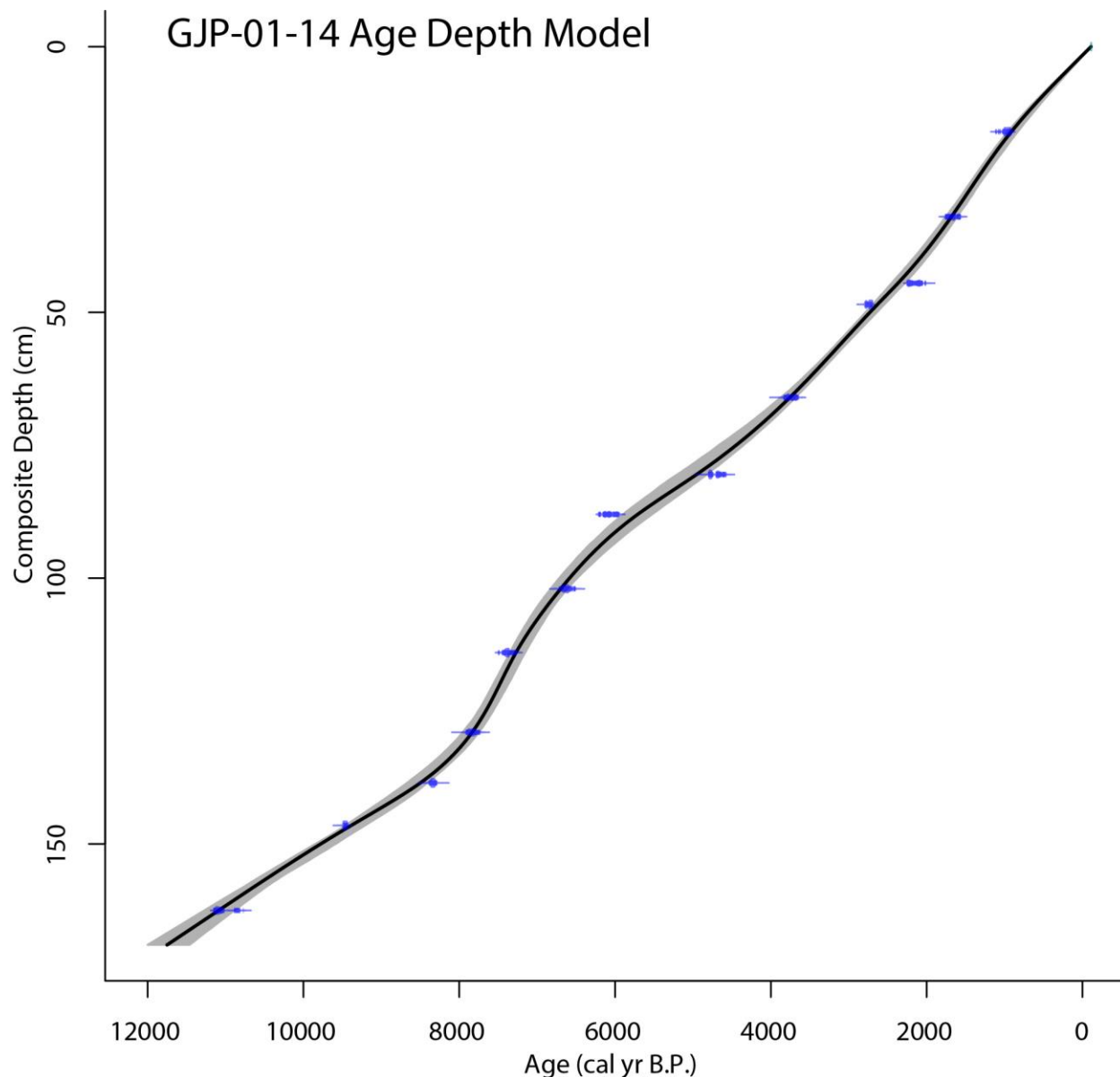
234 The chronology of the composite sedimentary record is based on 13 radiocarbon dates of organic  
235 macrofossils taken from the GJP-01-14 core (**Table 2**). An age model was created using the Clam age  
236 modeling package (Blaauw, 2010) for the open-source software R (v. 3.0.1; R Development Core Team,  
237 2013) (**Figure 4**). A smooth spline function was used to create the age model, with a default smoothing  
238 value of 0.3 applied. Radiocarbon dates were calibrated using the terrestrial northern hemisphere  
239 Intcal13.14C curve (Reimer et al., 2013). Calculated sedimentation rates in Gjøvatnet using this age

240 model vary from a maximum of ~29.8 cm/kyr to a minimum of ~8.6 cm/kyr with an average value of  
 241 14.7 cm/kyr. We also created an age model using the same parameters but with a linear interpolation  
 242 between data points instead of a smooth spline. We only use the linear interpolation-based age model as  
 243 an illustrative tool to highlight differences in sedimentation rate throughout the core, and do not use it for  
 244 any of our paleoclimate interpretations. We note that age estimates below a depth of 162.5 cm (11,140 cal  
 245 yr BP calibrated age) are based on extrapolation of the age model and therefore have unconstrained  
 246 uncertainty. This impacts only a small portion of our proxy data, which extends to ~167 cm, but does  
 247 affect our ability to accurately date the onset of lacustrine sedimentation in Gjøavatnet.

248 **Table 2: Radiocarbon results from macrofossils taken from GJP-01-14 core**

Core	Composite Depth (cm)	Sample Material	Sample Mass (mg)	Mg Carbon	<sup>14</sup> C Age	Error +/- (1σ)	Calibrated Age ±2 sigma	Δ <sup>13</sup> C ‰ VPDB
GJP-01-14 1of2	16	Plant Remains	4.2	0.8	1125	30	960 - 1172	-23
GJP-01-14 1of2	32	Plant Remains	9.1	0.8	1795	35	1619 - 1819	-22.4
GJP-01-14 1of2	44.5	Plant Remains	4.4	0.8	2175	35	2066 - 2312	-20.5
GJP-01-14 1of2	48.5	Plant Remains	7.8	0.7	2685	30	2753 - 2846	-24
GJP-01-14 1of2	66	Plant Remains	11.6	1.02	3530	35	3702 - 3895	-22.8
GJP-01-14 1of2	80.5	Plant Remains	2.2	0.5	4230	40	4628 - 4861	-24.1
GJP-01-14 1of2	88	Plant Remains	6.9	0.5	5350	50	6002 - 6084	-24.4
GJP-01-14 1of2	102	Plant Remains	8.7	0.6	5860	40	6563 - 6778	-27.7
GJP-01-14 1of2	114	Plant Remains	6.2	1.01	6520	50	7323 - 7556	-22.4
GJP-01-14 1of2	129	Plant Remains	14.2	1.35	7060	50	7789 - 7976	-25.5
GJP-01-14 2of2	138.5	Plant Remains	3.9	0.51	7590	40	8343 - 8448	-26.8
GJP-01-14 2of2	146.5	Plant Remains	3.6	Not Reported	8550	30	9494 - 9547	-26.1
GJP-01-14 2of2	162.5	Plant Remains	2.9	Not Reported	9690	40	10827 - 11217	Not Reported

249



250 **Figure 4:** Age depth relationship for GJP-01-14 core created in using the Clam modeling package  
 251 (Blaauw, 2010) for software R (v. 3.0.1; R Development Core Team, 2013) with 95% confidence  
 252 intervals for individual radiocarbon dates in blue.  
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 256 **4.2) Multivariate Analysis**

257 Calculated correlation coefficients for each of XRF-based elemental abundance datasets reveal  
 258 that most of the geochemical elements are highly positively correlated with each other and negatively  
 259 correlated with LOI (**Table 3**). The exception to this pattern is Ca, which is only weakly positively  
 260 correlated with the other elements (but still negatively correlated with LOI). Principal component  
 261 analysis yielded 2 components responsible for 91% of the observed variance in the dataset (**Table 4**).

262 Most of the geochemical elements align with PC1, responsible for 80.6% of the variance (**Table 3,**  
 263 **Figure 5**). As suggested by their correlation coefficients, most of the geochemical elements are positively  
 264 correlated with PC1, while LOI is inversely correlated. Accordingly, the downcore scores for PC1 look  
 265 broadly similar to elemental counts from the Itrax (**Figure 6**). PC2 (10.7% of variance) shows weak  
 266 correlations with most elements and LOI, but is highly positively correlated with the element Ca. We  
 267 recognize that PC1 is an extension of the elemental data and we utilize the principal component analysis  
 268 to highlight the similar behavior of these elements over the length of our record, and as justification for  
 269 interpreting each as a record of bedrock erosion (and therefore glacial activity). In the ensuing discussion,  
 270 we will consider Ti elemental abundances as a proxy for glacially derived sediment input to Gjøvatnet;  
 271 however, the use of PC1 or of a different element (apart from Ca, see below) would not change any of our  
 272 interpretations or conclusions.

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**Table 3: Correlation coefficients for variables used in PCA analysis**

<b>Correlation Coefficients</b>	<b>Ti</b>	<b>Al</b>	<b>Si</b>	<b>K</b>	<b>Ca</b>	<b>Mn</b>	<b>Fe</b>	<b>Rb</b>	<b>Sr</b>	<b>LOI</b>
<b>Ti</b>	1.00									
<b>Al</b>	0.81	1.00								
<b>Si</b>	0.87	0.75	1.00							
<b>K</b>	0.97	0.81	0.95	1.00						
<b>Ca</b>	0.78	0.61	0.50	0.66	1.00					
<b>Mn</b>	0.83	0.77	0.85	0.88	0.39	1.00				
<b>Fe</b>	0.84	0.72	0.90	0.90	0.34	0.94	1.00			
<b>Rb</b>	0.95	0.71	0.88	0.95	0.69	0.77	0.83	1.00		
<b>Sr</b>	0.91	0.69	0.82	0.90	0.80	0.65	0.68	0.94	1.00	
<b>LOI</b>	-0.91	-0.67	-0.69	-0.82	-0.85	-0.63	-0.64	-0.86	-0.85	1.00

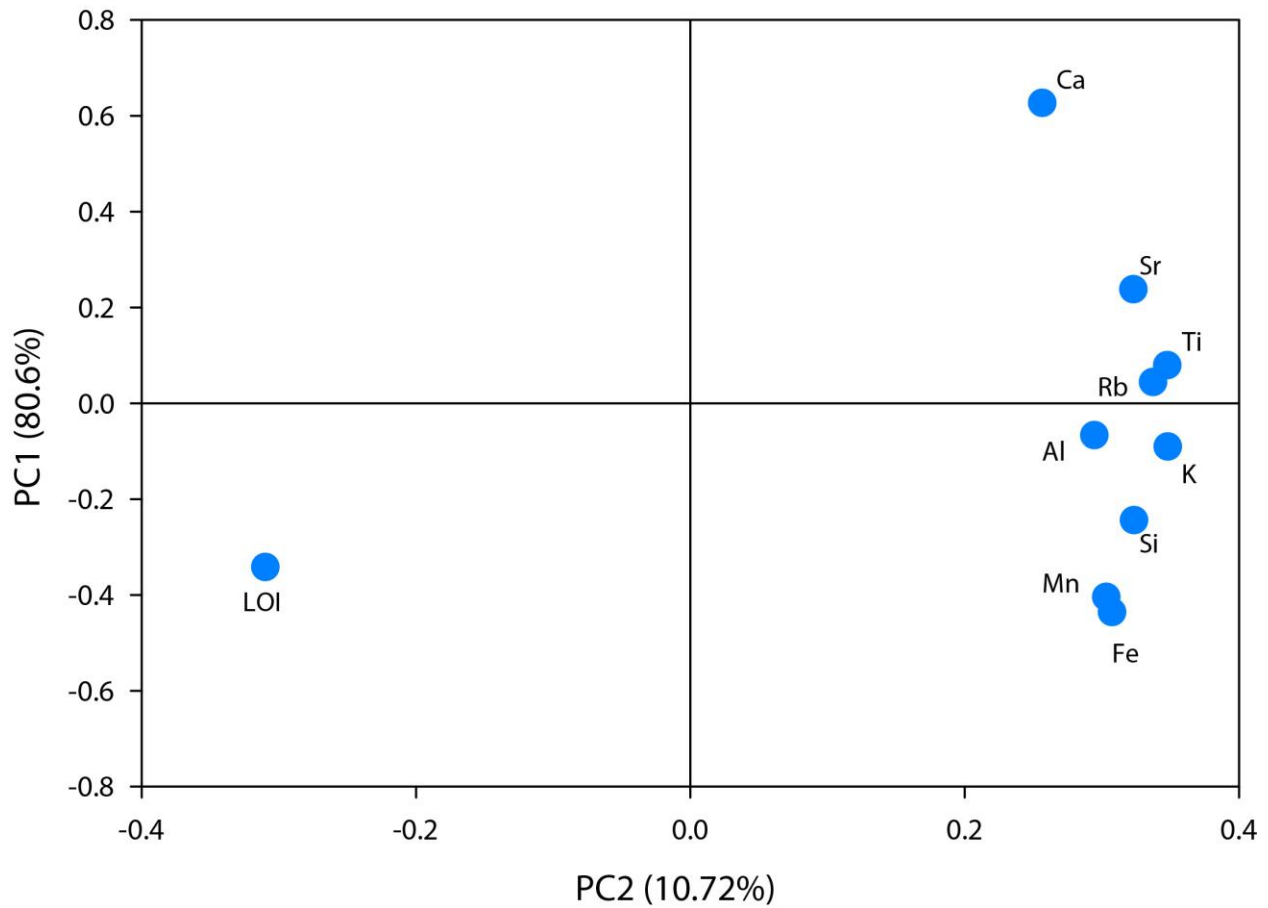
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**Table 4: Results from PCA analysis with 8 principal components**

<b>Principal Component #</b>	<b>Eigenvalue of Cov (x)</b>	<b>% Variance Captured this PC</b>	<b>Cumulative Variance Captured</b>
<b>1</b>	8.08	80.6	80.6
<b>2</b>	1.08	10.72	91.32
<b>3</b>	0.39	3.87	95.19
<b>4</b>	0.21	2.09	97.28
<b>5</b>	0.11	1.05	98.33
<b>6</b>	0.08	0.84	99.17

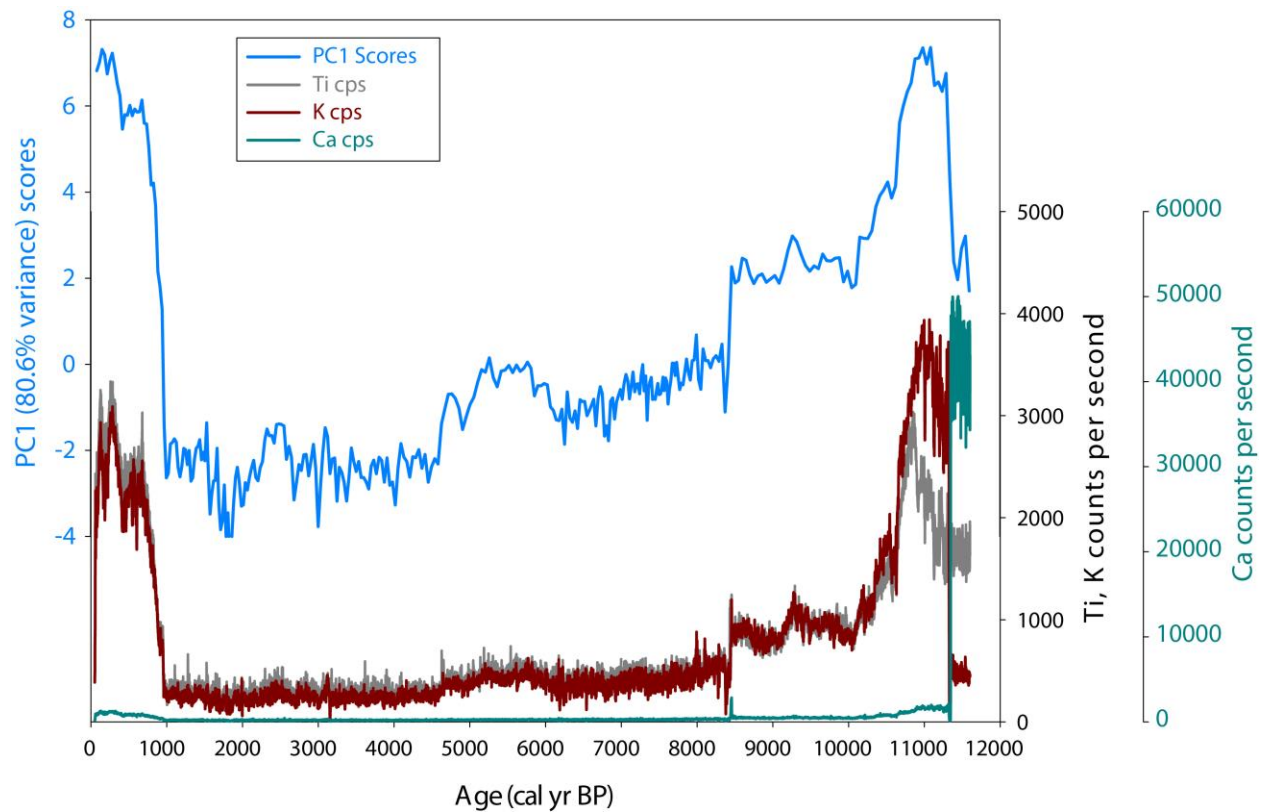
7	0.05	0.48	99.64
8	0.02	0.19	99.83

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**Figure 5:** Ordination diagram showing 1<sup>st</sup> (80.6% of variance) and 2<sup>nd</sup> (10.72% of variance) principal components of PCA.



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 285  
 286 **Figure 6:** Downcore scores of 1<sup>st</sup> principal component from PCA analysis (blue line) plotted on age scale  
 287 with Itrax elemental data for Ti (gray), K (dark red) and Ca (cyan) for core GJP-01-14. Note different  
 288 scale for Ca data relative to Ti and K.  
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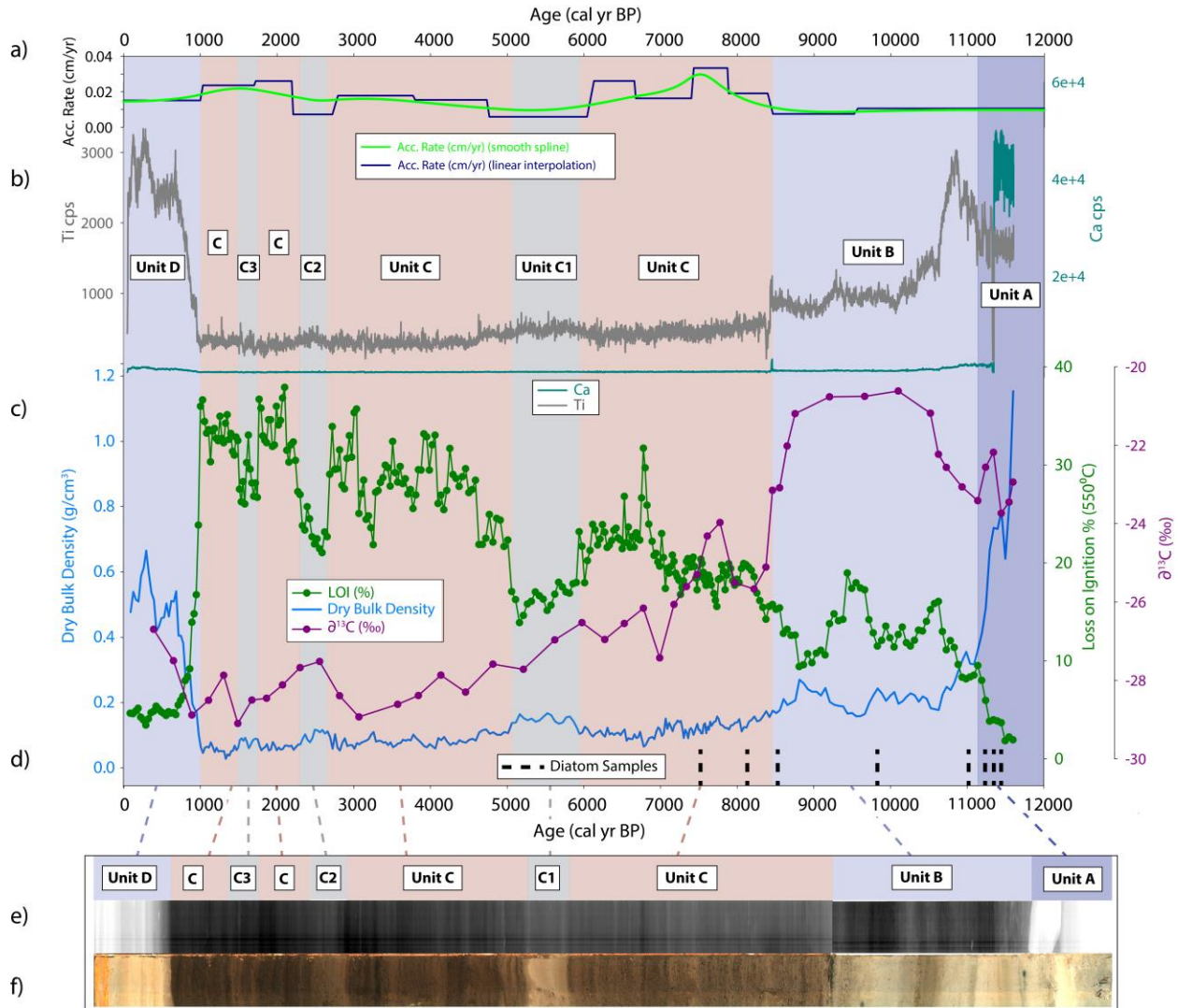
301 **4.3) Stratigraphy and Interpretation of Stratigraphic Units**

302 The composite Gjøvatnet sedimentary sequence was separated into 4 major units: A, B, C, and  
 303 D, and Unit C was further divided into 3 distinct subunits (**Figure 7**). This determination was made based  
 304 on major density transitions apparent in x-radiograph data and corresponding visual transitions between  
 305 gray, silty sediment and brown, organic rich sediment (**Figure 3**). Proxy data (Itrax elemental data,  
 306 diatom analyses, and LOI and DBD values) were then used to inform/confirm these definitions (**Table 5**)  
 307 (see below).

308 **Table 5: Relevant proxy data for sedimentary units identified in Gjøvatnet record**

Unit	Depth Range (cm)	Approx. Age Range (cal kyr BP)	Avg. Ca cps	Avg. Ti cps	Avg. DBD (g/cm <sup>3</sup> )	Avg. LOI	Avg. Acc. Rate (cm/kyr)	Avg. $\delta^{13}\text{C}$ (‰)
Unit A	209.5 - 162.5	Base – 11.1	24466 <sup>†</sup>	1623 <sup>†</sup>	0.71 <sup>†</sup> <i>n</i> = 9	3.9 <sup>†</sup> <i>n</i> = 9	9.8	-23.0 <i>n</i> = 5
Unit B	162.5 – 136.5	11.1– 8.4	655	1174	0.22 <i>n</i> = 56	13.0 <i>n</i> = 56	9.7	-22.2 <i>n</i> = 13
Unit C	136.5 – 16.5	8.4 – 1.0	231	372	0.09 <i>n</i> = 198	25.5 <i>n</i> = 198	18.4	-27.1 <i>n</i> = 24
Unit C1	89.5 – 81	5.9 – 5.0	253	489	0.15 <i>n</i> = 18	16.5 <i>n</i> = 18	10.1	-27.3 <i>n</i> = 2
Unit C2	48.5 – 42	2.7 – 2.2	199	341	0.10 <i>n</i> = 14	24.0 <i>n</i> = 14	15.7	-27.6 <i>n</i> = 2
Unit C3	32 – 27	1.7 – 1.5	174	279	0.08 <i>n</i> = 11	28.2 <i>n</i> = 11	21.6	-28.5 <i>n</i> = 1
Unit D	16.5 – 0	1.0 - present	785	2157	0.43 <i>n</i> = 29	6.9 <i>n</i> = 29	15.7	-27.7 <i>n</i> = 3

309 <sup>†</sup> Data collected down to a depth of ~167 cm



310

311 **Figure 7:** Downcore proxy results for GJP-01-14. Shaded background colors denote different  
 312 sedimentary units (Units A-D). a) Sediment accumulation rates based on smooth spline interpolation  
 313 (light green) and linear interpolation (dark blue) between radiocarbon dates; b) Itrax elemental data  
 314 showing Ti (gray) and Ca (cyan) (note different scales); c) Loss-on-ignition (dark green), bulk  $\delta^{13}\text{C}$   
 315 (purple) and dry bulk density values (dark blue); d) location of diatom taxonomy samples (black dashed  
 316 lines); e) normal image and f) X-ray image of GJP-01-14 core. Light/white shades in f) denote denser  
 317 material.

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324 **Unit A: Base of core (206.5 cm) – 11.1 cal kyr BP (162.5 cm)**

325 **Glacial Till**

326 The basal sedimentary unit, Unit A, consists of dark to light gray diamict (206.5 cm - ~170 cm),  
327 which transitions to dense silty sand (170 – 162.5 cm). The diamict in the lower portion of this unit is  
328 massive and poorly sorted, with individual clasts up to 5 cm in diameter. The x-radiograph of the upper  
329 section shows some faint evidence for horizontal bedding structures, but lacks the distinct laminations of  
330 later units. The unit broadly is characterized by low LOI values, ranging from 1.8% at ~11.6 cal kyr BP to  
331 ~8% at 11.2 cal kyr BP with a mean value of 3.9%. The highest dry bulk density values of the entire  
332 record are found in Unit A, with a mean value of 0.71 g/cm<sup>3</sup>. Values decrease steadily from a maximum  
333 of 1.15 g/cm<sup>3</sup> at a depth of 167 cm (lowest sample) to 0.41 g/cm<sup>3</sup> at the top of the unit (**Table 5, Figure**  
334 **7**). Unit A has the 2<sup>nd</sup> highest average Ti counts of the entire record (1622 cps), with relatively little  
335 variation about the mean. Ca counts during the majority of this interval are the highest of any unit by  
336 nearly two orders of magnitude. They reach a maximum value of 50,014 cps and are above 30,000 cps for  
337 most of the unit (mean of 24,466 cps) (**Table 5**) before declining to ~1500 cps before the end of Unit A.  
338  $\delta^{13}\text{C}$  values range from -22.2‰ to -23.7‰, with an average of -23.1‰ ( $n = 5$ ). Diatom species were  
339 analyzed in 3 samples from Unit A (dashed lines in **Figure 7**). Assemblages were dominated by  
340 *Pinnularia lenticular* and *Stauroneis anceps* family (*cf. gracilis, vandevijveri*), which are characteristic of  
341 a silty, shallow freshwater environment (Perren et al., 2012; Wojtal et al., 2014) (**Figure 8**).

342 One of the most interesting aspects of Unit A is the high Ca abundance during in this interval.  
343 High Ca counts are also observed in the other piston core collected from Gjøvatnet (GJP-02-14),  
344 confirming that it is a persistent feature of the sediments across the lake basin. One possible interpretation  
345 of the calcium signal is that it represents a period when the lake basin was subject to marine influence.  
346 Diatom analysis, however, has revealed that all species present at the time were freshwater-dwelling,  
347 ruling out the possibility of a marine influence. We note here that despite the abundant driftwood on the  
348 narrow strip of land separating Gjøvatnet from the ocean today, we see no evidence in our proxy data to  
349 suggest sustained or meaningful marine influence on the lacustrine sediment record.

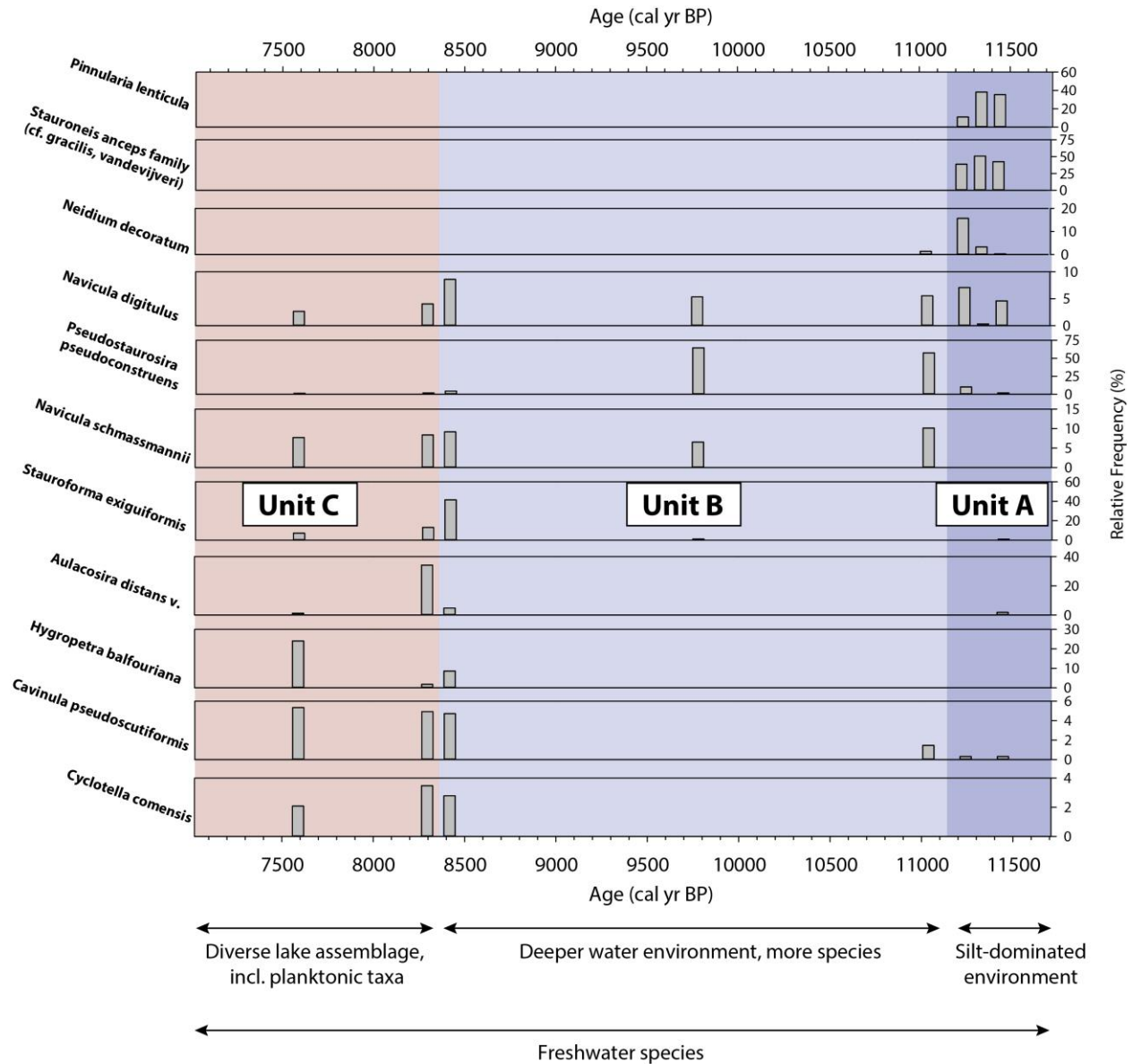
350 We interpret the high Ca abundance as a signal of bedrock erosion from the marble units within  
351 the Gjøavatnet catchment. Although the majority of the underlying bedrock consists of banded gneiss of  
352 the Smeerenburgfjorden Complex (Hjelle and Ohta, 1974; Ohta et al., 2007).), there are two small  
353 outcrops of marble that would have resulted in glacial flour with elevated Ca content in the northwest part  
354 of the catchment (blue outlines in **Figure 1**). The exposed marble units are not being eroded by the  
355 Annabreen glacier today, but would have been subject to glacial erosion if the glacier advanced across the  
356 outcrops. We suggest that the large decrease in XRF-inferred Ca deposition to the lake c. 11.5 cal kyr BP  
357 represents the retreat of Annabreen up-valley from the marble outcrops. If this interpretation is correct,  
358 the abrupt decline in sedimentary Ca abundance ~11.4 cal kyr BP represents a threshold response as the  
359 glaciers retreated beyond the marble outcrops, and as such, we do not consider the large change in Ca  
360 abundance in our assignment of stratigraphic units.

361 The carbon isotopic signature of this section is also intriguing, suggesting at face value a variable  
362 but perhaps predominantly marine source (values ranging from -22.2 to -23.7‰) (Meyers, 1997). Again,  
363 however, diatom analysis has ruled out this possibility. An alternative explanation may be found in the  
364 weathering of silicate rocks and limited recycling of carbon within the catchment immediately following  
365 deglaciation (Hammarlund, 1993). Glacial activity would have resulted in a large amount of freshly  
366 weathered siliciclastic material on the landscape following deglaciation. The weathering of this material  
367 could have resulted in bicarbonate delivery to the lake water, leading to  $^{13}\text{C}$  enrichment (i.e. more positive  
368  $\delta^{13}\text{C}$  values) of dissolved inorganic carbon and, therefore, of autochthonous organic material in the lake  
369 (Hammarlund, 1993).

370 We interpret Unit A to represent a period when Annabreen was terminating at or near the coring  
371 site during deglaciation of the catchment. Due to the extrapolated nature of our age model, we cannot  
372 place a definitive date on the onset of sedimentation in Gjøavatnet. The highest dry DBD and lowest LOI  
373 percentages are recorded during this interval, along with the presence of large individual clasts in the  
374 sediment, all of which suggest substantial glacial presence proximal to the coring site. The lowermost  
375 ~30cm of GJP-01-14 is comprised of glacial till/diamict, which was likely deposited when the glacier was

376 directly adjacent to (or overriding) the coring site. We speculate that the remainder of Unit A (~167 –  
377 162.5 cm), when DBD values were still relatively high, represents a period when Annabreen was likely  
378 terminating within the lake.

379           The boundary between Unit A and Unit B at ~11.1 cal kyr BP is primarily defined by DBD  
380 values, a shift in diatom species, and a major density change seen in the x-radiograph data (**Figure 3**).  
381 The diatom samples below the transition represent a silt-dominated environment (e.g Perren et al., 2012),  
382 as would be expected if Annabreen was terminating within the lake or was contributing a significant  
383 amount of meltwater to the lake system, whereas the samples above (in Unit B) are more diverse and  
384 point to a reduction of suspended silt in the water column. Together, these proxies suggest that while  
385 Annabreen was still active during the deposition of Unit B (see below), it was reduced in size and/or  
386 influence compared to Unit A.



387

388 **Figure 8:** Percent abundance of diatom taxa throughout the early evolutionary history of Gjøvatnet. Note  
 389 sedimentary unit delineations (labels and background shaded colors) that correspond with **Figure 7**.

390

391

392 **Unit B: 11.1cal kyr BP (162.5 cm) – 8.4 cal kyr BP (136.5 cm)**

393 **Deglaciation of Gjøvatnet Catchment**

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Unit B consists of a mixture of gray, laminated, clayey silt interbedded with relatively organic rich brown material. It is characterized by generally low but variable LOI values ranging from 8.1 – 19% (mean of 13.0%) (**Figure 7**). DBD values (0.14 – 0.36 g/cm<sup>3</sup>; mean of 0.22 g/cm<sup>3</sup>) are relatively high during this unit, though much lower than in preceding Unit A. Ti counts are higher in the lower portion of

398 this unit than Unit A, increasing to a maximum of 3034 counts at ~10.8 cal kyr BP before decreasing to  
399 ~800-1,000 cps for the majority of the Unit B. Ca counts in Unit B are significantly lower than in Unit A,  
400 decreasing from ~1,500 counts at the base of Unit B to ~300 cps (average value of 655 cps).  $\delta^{13}\text{C}$  values  
401 are the most enriched in this section of the core, rising from ~ -23.5‰ at 162.5 cm to a value of ~ -21‰  
402 (average of -22.2‰,  $n = 13$ ).

403 Three samples from Unit B were analyzed for diatom taxonomy. Assemblages in the lowermost  
404 two samples are dominated by *Navicula (Genkalia) digitulus* and *Pseudostaurosira pseudoconstruens* and  
405 also contain the first appearance of *Navicula schmassmannii* in the sediment record (**Figure 8**). These  
406 species are characteristic of a deeper lake with less suspended silt than the species from Unit A (Perren et  
407 al., 2012). The third sample, taken at a depth of 137 cm (approx. 8.6 cal kyr BP), is characterized by a  
408 more diverse assemblage including *Stauroforma exiguiformis*, *Hyropetra balfouriana*, and *Aulacoseira*  
409 *distans*.

410 We interpret Unit B to represent an interval when Annabreen was still present within the  
411 catchment, but likely not terminating within the lake. The two oldest diatom samples from this interval  
412 (~11 and 9.7 cal kyr BP) (**Figure 7, Figure 8**) are characteristic of a deeper lake with lesser suspended  
413 silt load than during the deposition of Unit A. Elemental abundance data (**Figures 6, 7**) suggest a broad  
414 decrease in the influence of Annabreen on the sediment record across this interval, although reductions in  
415 elemental counts occur stepwise (cf., Ti counts). The LOI trend across Unit B is also non-linear and is  
416 quite variable at multi-centennial timescales (**Figure 7**), suggesting the glacier may have been fluctuating  
417 dynamically during this time and/or that sedimentation was influenced by glaciofluvial dynamics in the  
418 glacier forefield as ice retreated. The third diatom sample from this section (just prior to 8.4 cal kyr BP)  
419 (**Figure 8**) reveals a more diverse assemblage than the two older samples from Unit B, including  
420 planktonic taxa. Again, this assemblage suggests Annabreen's influence on sedimentation in the lake  
421 waned throughout this period.

422 Bulk organic carbon isotope values in Unit B are the most positive of the record, with a mean of -  
423 22.2‰. Such a value is generally associated with marine algae (Meyers, 1997), however, only freshwater

424 diatom species are found in the Gjøavatnet sediment record, precluding a marine source for the relatively  
425  $^{13}\text{C}$ -enriched carbon isotope values. As discussed above, a possible explanation is the weathering of  
426 glacial flour derived from silicate rocks during this time that could have increased the  $\delta^{13}\text{C}$  value of DIC  
427 (dissolved inorganic carbon) in Gjøavatnet (Hammarlund, 1993).

428 The boundary between Unit B and Unit C at ~8.4 cal kyr BP is marked by an increase in LOI, an  
429 abrupt decrease in Ti abundance, and a concomitant shift in  $\delta^{13}\text{C}$  values (**Figures 6, 7, 8**). It is apparent  
430 that the nature of sedimentation in Gjøavatnet changed dramatically at this point. The most likely  
431 explanation is the disappearance, or the dramatic reduction in size, of the Annabreen glacier at ~8.4 cal  
432 kyr BP.

433

434 **Unit C: 8.4 cal kyr BP (136.5 cm) – 1.0 cal kyr BP (16.5 cm)**

435 **Non-glacial sedimentation in Gjøavatnet**

436 Unit C comprises the majority of the sediment record from Gjøavatnet (120 cm of ~206 total) and  
437 is composed of laminated brown organic rich sediment with interbedded gray minerogenic layers. The  
438 highest LOI values of the entire core are recorded in Unit C, and generally follow a linear increasing trend  
439 from ~18% to 35% (average of 25.5%) (**Figure 7**). This linear trend is interrupted by at least three distinct  
440 multicentennial-scale intervals characterized by abrupt shifts to relatively lower LOI values, higher Ti and  
441 DBD values, and decreases in sedimentation rate. These subunits are defined as C1 (81-89.5 cm, ~5.0 –  
442 5.9 cal kyr BP); C2 (42-48.5cm, 2.2 – 2.7 cal kyr BP); and C3 (27-32cm, 1.5-1.7 cal kyr BP). Proxy data  
443 for these three subunits are compared to average values from the remainder of Unit C, and presented in  
444 **Table 5**, to examine the differences between these subunits and Unit C in general. DBD values for Unit C  
445 generally follow the inverse trend of LOI, beginning at ~0.15 g/cm<sup>3</sup> and declining to a minimum of 0.03  
446 g/cm<sup>3</sup> near the top of the section. Ti counts are generally low throughout this portion of the record,  
447 ranging from 88 – 661 cps with an average value of 372 cps. Ca counts are extremely low in Unit C,  
448 averaging 231 counts (max of 305 cps, min of 69 cps), with little variation. During the deposition of this



449 sedimentary unit we interpret Ti abundances to reflect catchment dynamics unrelated to glacier activity  
450 (e.g., such as changes in runoff), or by changes in dilution by organic matter deposition.

451 Carbon isotope values in Unit C increased from ~ -25.5‰ to -24‰ at 7.6 cal kyr BP and then  
452 steadily decreased to -28.5‰ (average value of -27.1‰). Two samples analyzed for diatoms from the  
453 base of this unit, at ~ 8.1 (135.5 cm depth) and 7.6 cal kyr BP (121.5 cm depth), show similar  
454 assemblages to the uppermost sample from Unit B (**Figure 8**), revealing a diverse community including  
455 planktonic diatoms living in the upper water column

456 We interpret Unit C to represent a phase during the Holocene when the Annabreen glacier either  
457 completely melted or was too small to influence sedimentation in the lake. Ti counts and DBD are at their  
458 lowest during this period (**Table 5, Figure 7**). Bulk  $\delta^{13}\text{C}$  values during this period average -27.1‰, likely  
459 reflecting a lacustrine algal source (-25 to -30‰) (Meyers, 1997). LOI values increase from ~20% at ~8.0  
460 cal kyr BP to nearly 40% near the transition to Unit D at 1.0 cal kyr BP. An increasing trend in % organic  
461 matter during the Holocene has also been observed in other Svalbard lakes (e.g. Gjerde et al., *in press*;  
462 van der Bilt et al., 2015) and is attributed to lake and catchment ontogeny and greater nutrient recycling.  
463 The trend may have also have been influenced by increasing preservation of organic matter, as declining  
464 summer insolation potentially shortened the ice-free season resulting in greater bottom water anoxia or  
465 hypoxia (Laskar et al., 2004).

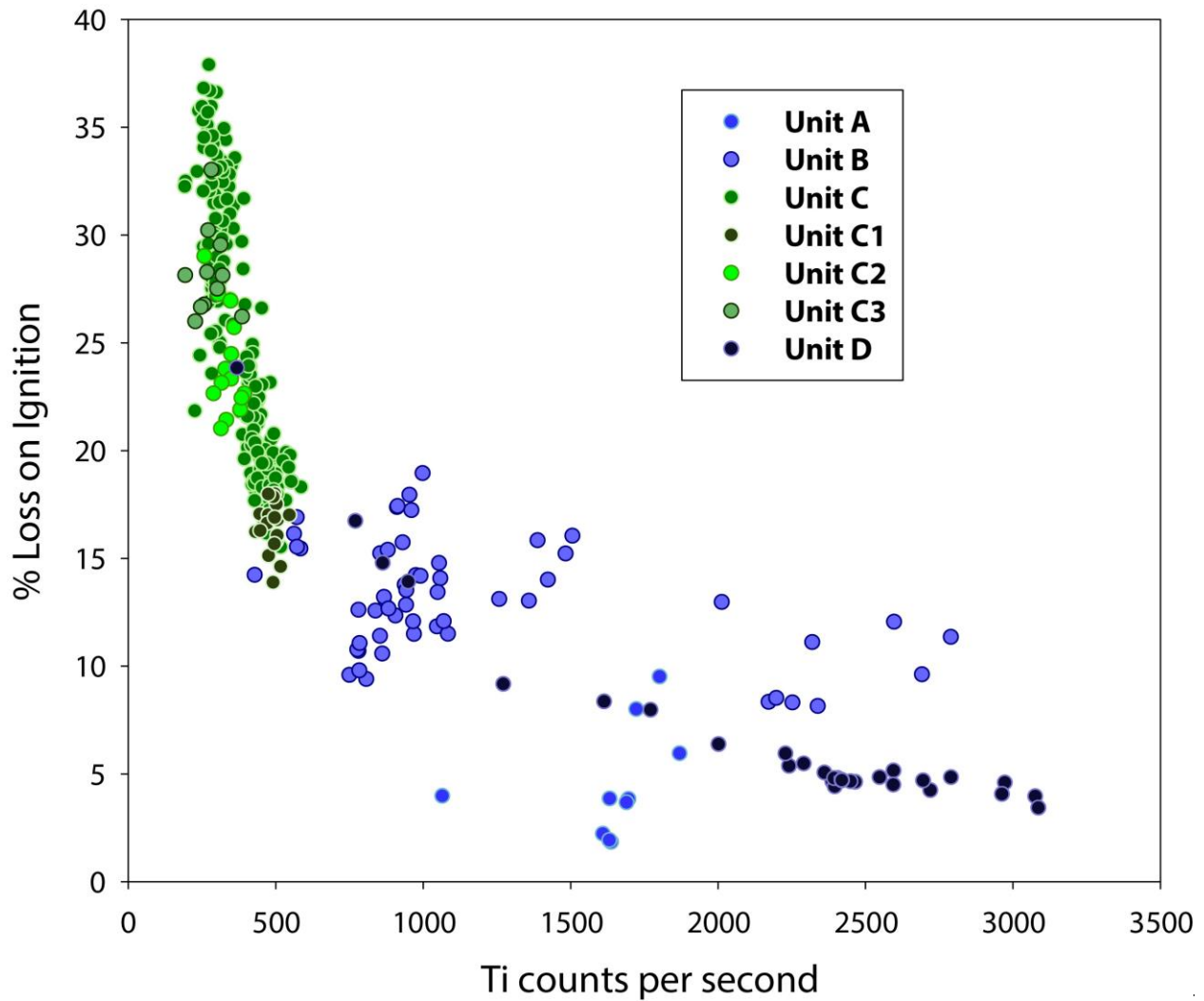
466

### 467 **Interpretation of Subunits C1, C2, and C3**

468 Unit C is punctuated by abrupt transitions between brown, organic-rich sediment and gray, more  
469 minerogenic sediments (the latter defined as subunits C1-C3; **Figure 7**). These transitions are apparent in  
470 LOI, visual stratigraphy, x-radiograph images, and DBD values. As evidenced by the visual stratigraphy,  
471 core images, and x-radiograph, these units appear massive, with few laminations. Subunits C1-C3 could  
472 represent: (i) short-lived advances of the Annabreen glacier, (ii) slump activity (e.g. turbidites), or (iii)  
473 periods of reduced organic matter accumulation. We suggest that advances of Annabreen are not likely to  
474 have caused these changes because sediment characteristics during subunits C1, C2, and C3 are not

475 consistent with other intervals associated with glacial erosion in the catchment (Units A, B, D). Ti counts  
476 are much lower in the Unit C subunits, for example (**Figure 7**). Additionally, the relationship between Ti  
477 and %LOI is broadly similar across Unit C and subunits C1-C3 relative to Units A, B and D, inferred to  
478 represent a glacial signature (**Figure 9**). Furthermore, the timing of Unit C1 corresponds with the interval  
479 when nearby lake Hakluytvatnet completely dried out, likely in response to dry conditions (c. 7.7-5.0 cal  
480 kyr BP), suggesting the precipitation regime was not favorable for the regrowth of Annabreen (Gjerde et  
481 al., *in press*; Balascio et al., *in press*). It is also unlikely that these subunits were the result of slump  
482 events. Radiocarbon dates from either side of both units C1 and C2 were used to quantify sedimentation  
483 rates across each interval using a linear interpolation between data points (**Figure 7**), and indicate that  
484 sedimentation rates slowed during deposition of these subunits. Mass wasting events, such as slumping,  
485 would lead to an increase, not a decrease in sedimentation rate.

486         We therefore interpret subunits C1, C2, and C3 as intervals of reduced organic productivity, most  
487 likely driven by periods of prolonged summer lake ice cover and/or drier and colder conditions on  
488 Amsterdamøya. Sediment was likely delivered to the lake during a short period of reduced ice cover  
489 during the summer, which have been potentially limited to a moat around the lake edge.



490

491 **Figure 9:** Relationship between % Loss-on-ignition (LOI) and Ti counts from Itrax. Data from  
 492 sedimentary units when Annabreen glacier is interpreted to be present are shown in blue shades (Unit A,  
 493 B, D); data from units when Annabreen interpreted to be absent are shown in green shades (Unit C, C1,  
 494 C2, C3).

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499 **Unit D: 1.0 cal kyr BP (16.5 cm) – present**

500 **Re-advance of Annabreen**

501 The uppermost stratigraphic unit, Unit D, is characterized by an abrupt shift to gray, minerogenic  
 502 sediment with faint laminations and a small number of sand sized grains (visually identified from the split

503 core). Unit D is characterized by low LOI values (mean of 6.9%) and high DBD (mean of 0.43 g/cm<sup>3</sup>)  
504 (**Table 5**). The dense nature of Unit D is also evident from the X-radiograph (**Figure 7**). Ca counts  
505 increase in this unit relative to Unit B (average of 785 vs. 231), but remain nearly two orders of  
506 magnitude lower than Unit A at the base of the core. Bulk  $\delta^{13}\text{C}$  values increase from -28.9‰ to -26.7‰  
507 across the three samples representing this section. Ti counts in Unit D are the highest of the entire record  
508 (mean of 2,127 cps).

509 We interpret Unit D to represent the reemergence of the Annabreen glacier, though this could also  
510 represent the transition from residual cold-based ice to a polythermal glacier. The youngest radiocarbon  
511 date from the Gjøavatnet record was taken just below this transition, allowing confident age assignment to  
512 the boundary. Although there are only three  $\delta^{13}\text{C}$  samples from this interval, we note that they increase  
513 from -28.9‰ to -26.7‰, a change similar in magnitude to the pattern seen at the beginning of Unit B,  
514 which is likely related to increased glacial erosion and delivery of relatively <sup>13</sup>C-enriched material to the  
515 lake (Hammarlund et al., 1993).

## 516 **Summary**

517 The sedimentary units described here represent the broad phases of environmental change within  
518 the Gjøavatnet catchment. Unit A represents the period when Annabreen was likely larger than today and  
519 much of the catchment covered by ice. We infer that the glacier was terminating within the lake during  
520 this time. The transition to Unit B at ~11.1 cal kyr BP likely marks the time when Annabreen retreated  
521 from the lake basin. We propose that the glacier was present in the Gjøavatnet catchment until ~8.4 cal  
522 kyr BP. From ~8.4-1.0 cal kyr BP the glacier was absent or had become dramatically diminished in size,  
523 and variations in sediment properties were likely controlled primarily by changes in summer temperature  
524 and/or the duration of the summer ice free season. At ~1 cal kyr BP the local ELA lowered enough to  
525 allow Annabreen to reform, and the glacier has been terminating in the lake since that time.

526  
527

## 528 **5) Regional Paleoclimate Context of Gjøavatnet Record**

529

### 530 **5.1) Early Holocene: Deglaciation of Amsterdamøya**

531  
532           During the Last Glacial Maximum, ice extended all the way to the shelf edge in NW Svalbard, ~8  
533 km offshore (Ingólfsson and Landvik, 2013). During this time it is likely that the majority of the  
534 Gjøavatnet catchment was covered by ice, although the >300 m high plateaus on the island have been ice  
535 free for at least 80 cal kyr BP (Landvik et al., 2003). Ice began to retreat from the shelf edge sometime  
536 prior to 14 cal kyr BP, reaching the coast of NW Svalbard by ~13.8 cal kyr BP (Ingólfsson and Landvik,  
537 2013). The deglaciation of the nearby Hakluyvatnet catchment at ~12.8 cal kyr BP suggests some small  
538 cirques on the island were largely ice-free by this time (Gjerde et al., *in press*). However, Annabreen was  
539 still terminating near the coring site in Gjøavatnet for another ~1.7 kyr until it retreated out of the lake  
540 basin at ~11.1 cal kyr BP and then disappeared, or at least greatly diminished in size, at 8.4 cal kyr BP.

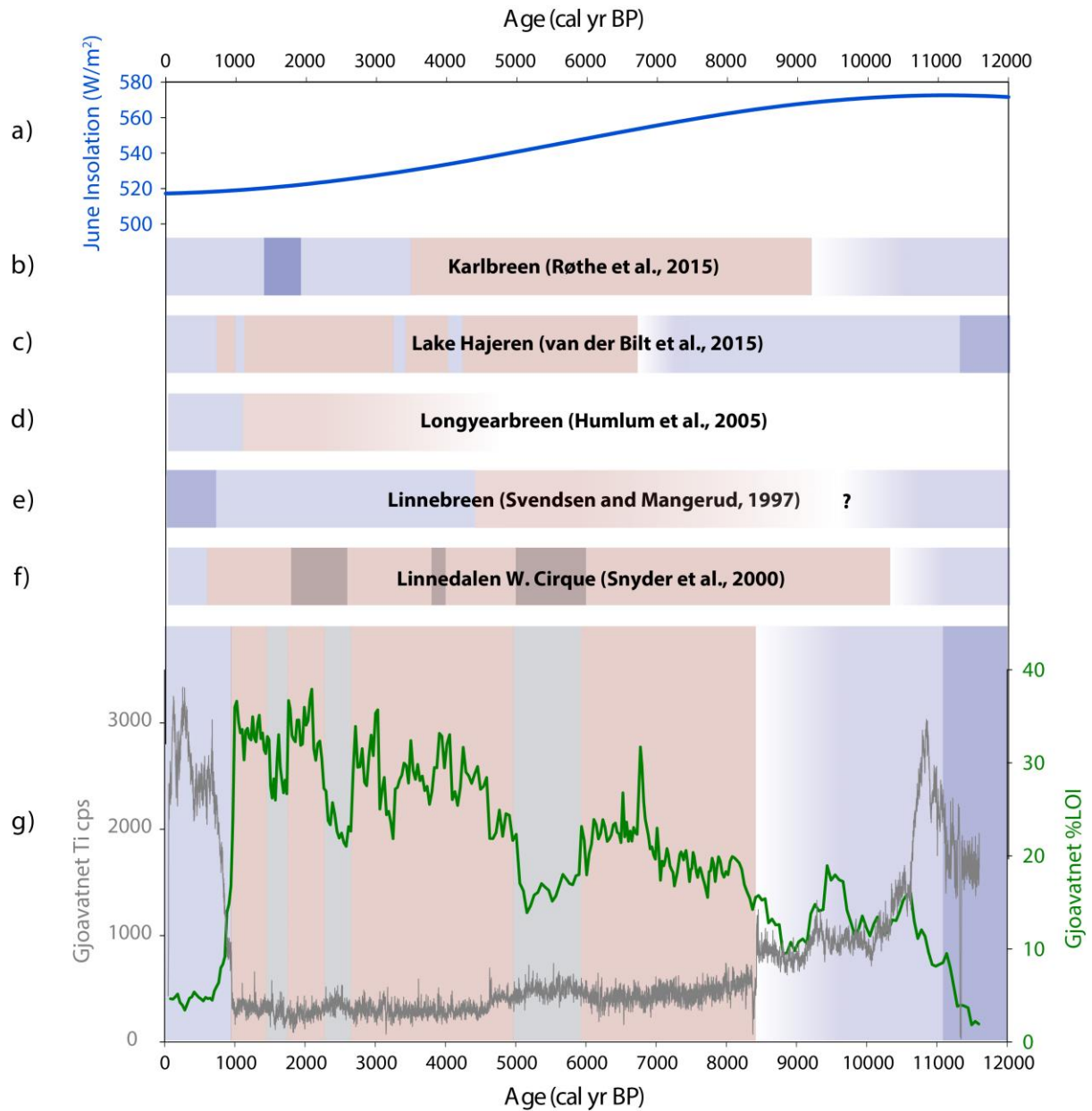
541           Proglacial lake records from the nearby Mitrahalvøya peninsula also point to a complex deglacial  
542 history in western Svalbard, with the catchment of Lake Kløsa deglaciating ~9.2 cal kyr BP (Røthe et al.,  
543 2015), while glacial ice persisted in the catchment of Lake Hajeren until ~7.4-6.7 cal kyr BP (van der Bilt  
544 et al., 2015) (**Figure 10**). Further south in the Linné valley, the Linnébreen glacier and another small  
545 cirque glacier in the same area are believed to have melted away during the early Holocene ( Svendsen  
546 and Mangerud, 1997; Snyder et al., 2000; Reusche et al., 2014). Lacustrine alkenone-based temperature  
547 reconstructions from Amsterdamøya and the Mitrahalvøya peninsula also point to warm conditions before  
548 ~8 kyr BP (van der Bilt et al., 2016).

549           The proposed final deglaciation of the Gjøavatnet catchment (~8.4 cal kyr BP) occurred during a  
550 prolonged period of warm surface water conditions in Fram Strait (Müller et al., 2012; Werner et al.,  
551 2013; 2015; Rasmussen et al., 2014). Aagaard-Sorensen et al. (2014) found the warmest Mg/Ca based  
552 temperatures in their record from core MSM5/5-712-2 from ~10.5-7.9 kyr BP. IP<sub>25</sub> concentrations (a  
553 biomarker indicative of diatoms associated with sea ice) (Belt et al., 2007) from the same core also  
554 suggest the region experienced “significantly reduced ice cover” between 8.2 and 7.8 cal kyr BP (Müller  
555 et al., 2012). IP<sub>25</sub> concentrations in the nearby MSM5/5-723-2 core also suggested warm temperatures and  
556 low concentrations of sea ice from the period ~11-7 kyr BP (Werner et al., 2015). Foram-based

557 temperatures also suggest increased advection of warm Atlantic water into this part of the Fram Strait  
558 during this time (Werner et al., 2013). While we acknowledge the possibility that Annabreen was  
559 dramatically reduced in size during the mid-Holocene (or became a cold-based glacier with little erosive  
560 power), we point to warming trends seen in marine records, along with evidence that many other glaciers  
561 in Svalbard melted away during this time (**Figure 10**), as strong evidence for our interpretations.

562           Warmer conditions following deglaciation generally define the early Holocene climate of this  
563 region, although periodic cooling events have been identified in the North Atlantic and attributed to  
564 freshwater forcing (Sejrup et al., 2016). The “8.2” event (Alley et al., 1997) is the most prominent of  
565 these, although we do not see it expressed in the Gjøavatnet record, a finding that echoes another similar  
566 study from Svalbard (van der Bilt et al., 2015). It may be that dry conditions during this time (Rohling  
567 and Pälike, 2005), coupled with its brief duration (Thomas et al., 2007), prevented Annabreen from re-  
568 growing sufficiently to impact the sediment in Gjøavatnet.

569



570

571 **Figure 10:** Holocene glacial activity reconstructions from Svalbard: a) June insolation values at 79°N  
 572 over the Holocene (Laskar et al., 2004); schematic depiction of glacier activity in Svalbard based on  
 573 published reconstructions: a) and b) proglacial lake reconstructions from Mitrahelvøya peninsula (Røthe  
 574 et al., 2015; van der Bilt et al., 2015); c) proglacial paleosol and vegetation study (Humlum et al., 2005);  
 575 d) & e) proglacial lake records from Linnévatnet (Svendsen and Mangerud, 1997; Snyder et al., 2000); f)  
 576 %LOI and titanium counts from Gjøavatnet (this study). Blue boxes represent periods of glacial activity  
 577 (dark blue = enhanced glacial activity), red boxes suggest no glacier was present in the catchment or  
 578 glacial activity was greatly reduced; gray boxes indicate periods of reduced organic matter accumulation  
 579 (Gjøavatnet) or increased laminations (Snyder et al., 2000) that are not interpreted as glacial activity.

580

581

## 582 5.2) Middle Holocene – Sea ice and freshwater influences on Amsterdamøya climate

583

584 During the middle Holocene period on Amsterdamøya (~8-1 cal kyr BP), we propose that  
585 Annabreen was absent from the catchment. This period is generally marked by the absence of a glacier in  
586 the Gjøavatnet catchment and punctuated by periodic decreases in organic matter accumulation in the lake  
587 (subunits C1-C3). We hypothesize that the most likely mechanism for these intervals is an abrupt change  
588 in temperature and/or precipitation (we note these two parameters are positively correlated today (Førland  
589 et al., 2011)), which at this maritime location would most likely be driven by offshore oceanographic  
590 conditions. Periodic increases in freshwater input and the presence of a surface freshwater layer (and  
591 likely accompanying sea ice) could dramatically lower both temperature and precipitation near  
592 Gjøavatnet, leading to the reduction in organic matter accumulation seen during the subunits.  
593 Interestingly, Unit C1 occurs during a period of increased deposition of discrete laminae in Linnévatnet  
594 (Snyder et al., 2000) (grey shaded area in **Figure 10**). The deposition of Unit C1 also occurs during a  
595 hiatus in sedimentation in nearby Haklutuyvatnet (Gjerde et al., *in press*), suggesting dry conditions on  
596 Amsterdamøya during this time. Numerous authors have suggested the presence of a freshwater cap or  
597 lens near Svalbard (Rasmussen et al., 2013; Werner et al., 2015, 2013), although the exact timing and  
598 magnitude of this oceanographic feature remains somewhat ambiguous. With respect to the most  
599 prominent of the subunits, C1, there does appear to be complimentary evidence for oceanographic  
600 changes during the period ~6-5 cal kyr BP.

601 Multiple marine records from near Svalbard and in the Fram Strait have found evidence for an  
602 increased flux of cold water from the Arctic around 6 cal kyr BP. Ślubowska et al. (2005, 2007) found  
603 increases in the concentrations of the benthic foraminifera *E. excavatum*, characteristic of Arctic ocean  
604 water, ~6.8 and 6 cal kyr BP north of Svalbard. Ebbesen et al. (2007) cite a shift in  $\delta^{18}\text{O}$  values of  
605 foraminiferal tests at 6 cal kyr BP as evidence for a change in the relative contributions of water masses  
606 off western Svalbard. Numerous studies have also shown that the thermophilous mollusk, *Mytilus edulis*,  
607 likely died out in northern Svalbard around 6-5 cal kyr BP (Blake, 2006; Salvigsen, 2002). Evidence from  
608 a proglacial fjord record in Nordauslandet points to the rapid deposition of a glacial diamict between 5.8



609 and 5.7 cal kyr BP (Kubischta et al., 2011). These reconstructions point to cooling conditions occurring  
610 during the hiatus in Haklutuyvatnet and the deposition of Unit C1 in Gjøavatnet.

611 Interestingly, several proxy reconstructions from the Fram Strait, just west of Amsterdamøya  
612 (**Figure 1**), suggest there was warm and saline water in the *subsurface* ocean (~depth of 100 m) during  
613 this time (Müller et al., 2012; Werner et al., 2013; 2015; Aagaard-Sorensen et al., 2014). Two proxies for  
614 sub-surface seawater temperature from core MSM5/5-712-2 depict warm temperatures from 6.1 to 5.2 ka  
615 (Werner et al., 2013; Aagaard-Sorensen et al., 2014) (**Figure 11**). The warm, saline water is presumed to  
616 be Atlantic-sourced water carried by the WSC. Werner et al. (2013) attribute the subsurface temperature  
617 increase to insulation by meltwater that limited heat loss to the atmosphere; if correct, these hydrographic  
618 conditions suggest an increased amount of cold, fresh meltwater (and also associated sea ice) near  
619 Amsterdamøya during this time. The subsurface warmth is not a feature of all temperature records from  
620 the Fram Strait, however. The foram-based subsurface temperature reconstruction from nearby core  
621 MSM5/5-723-2 (Werner et al., 2015) instead suggests that a broad cooling trend began around 6 kyr BP.  
622 6 kyr BP also marks the beginning of a decline in both the concentration of subpolar planktic forams and  
623 reconstructed subsurface (75m depth) temperatures in the Kongsfjorden Trough (Rasmussen et al., 2014).

624 Local sea ice reconstructions also do not support the notion of dramatic increases in sea ice in  
625 eastern Fram Strait from ~6-5 kyr BP. IP<sub>25</sub> concentrations from cores MSM5/5-723-2 and MSM5/5-712-2  
626 are broadly stable during this time, as are IRD concentrations (**Figure 11**) (Müller et al., 2012; Werner et  
627 al., 2013, 2015). Interestingly, however, relatively high IP<sub>25</sub> concentrations are found on the East  
628 Greenland Shelf from ~6.5-5.6 kyr BP, the only extended period of high IP<sub>25</sub> accumulation until ~1.0 cal  
629 kyr BP (Müller et al., 2012). Funder et al. (2011) also inferred increased export of multiyear sea ice out of  
630 the Fram Strait after 6 kyr BP, based on driftwood deposits on the coast of East Greenland.

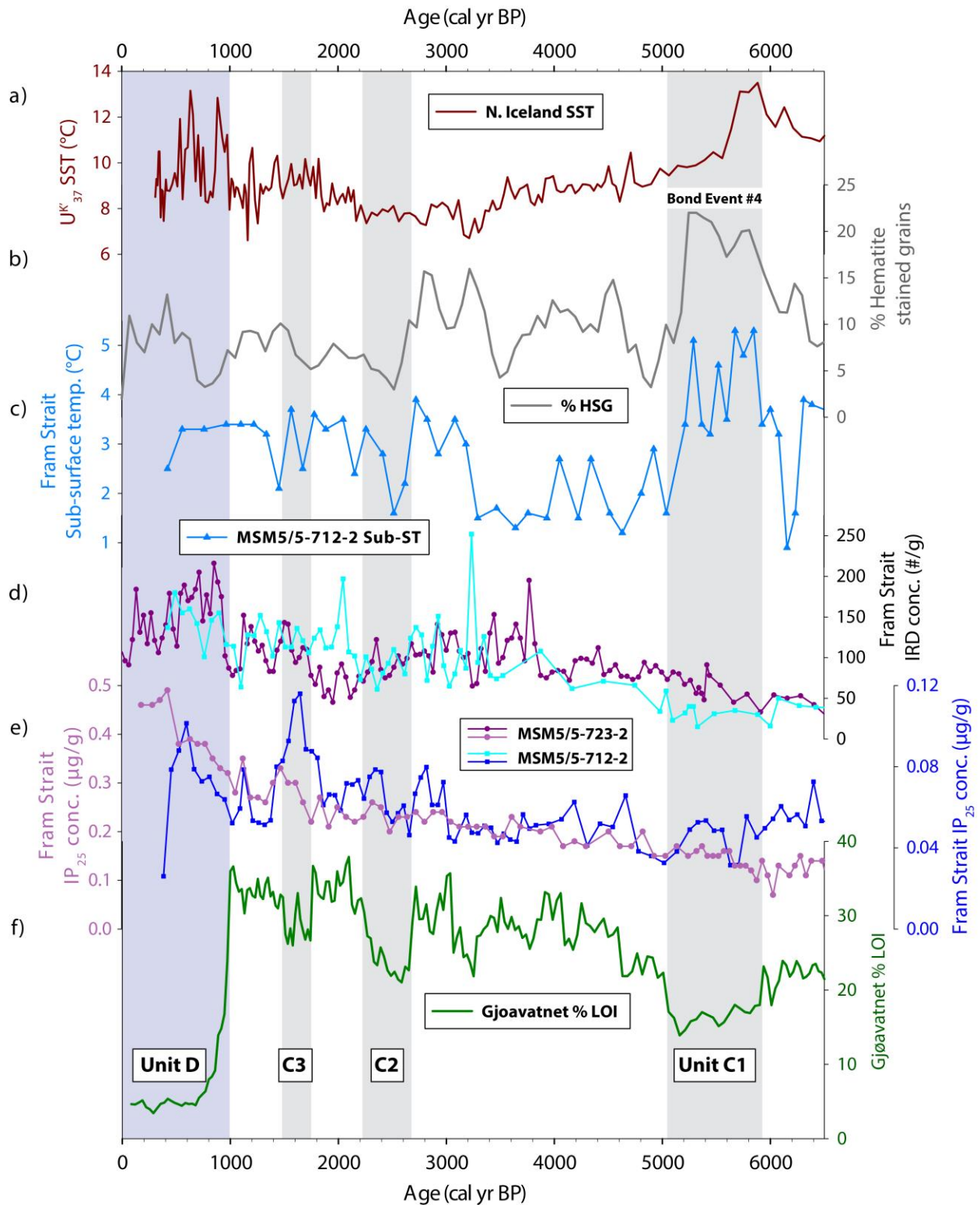
631 Further afield, numerous marine records from the north of Iceland also suggest dramatic changes  
632 were occurring ~6-5 cal kyr BP. Castañeda et al. (2004) and Knudsen et al. (2004) both cite cooling  
633 trends beginning at 6.2 and 6.0 cal kyr BP respectively, while the alkenone based SST record of Moosen  
634 et al. (2015) shows a similar cooling trend beginning at this time (**Figure 11**). There is also evidence for

635 an increase in sea ice concentrations north of Iceland beginning at 6.2 cal kyr BP (Cabedo-Sanz et al.,  
636 2016). Risebrobrakken et al. (2011) cite 6 cal kyr BP as the end of the Holocene Thermal Maximum in  
637 the North Atlantic, based on 6 cores from the Nordic and Barents Seas. Further south in the Atlantic basin  
638 Hoogakker et al. (2011) suggest 6.5 cal kyr BP was the start of a major reorganization of water masses  
639 associated with deep ocean circulation. Our C1 subunit also aligns with the largest so-called Bond Event  
640 of the Holocene (Bond Event #4) identified in a marine record in the North Atlantic (Bond, 2001) (**Figure**  
641 **11**).

642 Although there is not a clear relationship between these offshore records and the timing of  
643 subunits C2 (c. 2.7-2.2 kyr BP) and C3 (c. 1.7-1.5 kyr BP), the IP<sub>25</sub> concentration record of Müller et al.  
644 2012, shows the highest accumulation rates of IP<sub>25</sub> in the entire record during subunit C3 (**Figure 11**). A  
645 minor dip in LOI concentrations in Gjøavatnet at ~3.2 cal kyr BP, which is one of the most pronounced in  
646 the Hakluyvatnet record (Gjerde et al., *in press*), does appear contemporaneous with a spike in IRD in the  
647 record of Werner et al., (2013) as well. Similar to the 6-5 cal kyr BP interval, many of the marine records  
648 from the Fram Strait suggest a warming of subsurface waters and an increase in meridional overturning  
649 circulation after 3 ka (e.g. Berben et al., 2014; Rasmussen et al., 2013; Sarnthein et al., 2003; Werner et  
650 al., 2015) while simultaneously suggesting increased sea ice cover at the surface (Müller et al., 2009;  
651 2012; Werner et al., 2015).

652 In summary, it appears there is evidence for greater subsurface warming and increased Atlantic  
653 advection around 6-5 kyr BP (Werner et al., 2013; 2015; Aagaard-Sorensen et al., 2014), as well as  
654 increased sea ice concentrations in the western part of the Fram Strait (Müller et al., 2012, Funder et al.,  
655 2011), when we infer cold conditions on Amsterdamøya. The agreement between the end of unit C1 in  
656 Gjøavatnet and the resumption of sedimentation in nearby Hakluyvatnet also points to a wider forcing  
657 beyond the lake catchment itself. We propose similar oceanographic conditions may have existed during  
658 the deposition of subunits C2 and C3.

659



660

661 **Figure 11:** Compilation of relevant paleoclimate data from NW Svalbard, Fram Strait, and Davis Strait:  
 662 a) Alkenone based sea-surface temperatures from NW Icelandic shelf (Moossen et al., 2015); b) %  
 663 hematite stained grains from core VM29-191 as a proxy for drift ice, w/ Bond Event #4 labeled (Bond et  
 664 al., 2001); c) sub-surface (~100m depth) foram-based temperatures from core MSM5/5-712-2 (Werner et

665 al., 2013); d) IRD counts and e) IP<sub>25</sub> concentrations from core MSM5/5-712-2 (cyan/blue square data  
666 points) (Werner et al., 2013) and MSM5/5-723-2 (purple circle data points) (Werner et al., 2015) from  
667 Fram Strait; f) % loss-on-ignition data (green) from Gjøavatnet lake. Background shaded colors refer to  
668 sedimentary units from Gjøavatnet record (**Figure 7**).  
669

### 670 **5.3) Late Holocene advance of Annabreen glacier**

671 The sediment record from Gjøavatnet clearly indicates that input of minerogenic material, formed  
672 by bedrock erosion by the Annabreen glacier, abruptly began again ~1.0 cal kyr BP, and that this input  
673 has continued up to the present day. Other terrestrial records also suggest glaciers began to regrow around  
674 Svalbard around this time (Humlum et al., 2005; Snyder et al., 2000; van der Bilt et al., 2015), although  
675 Karlbreen on Mitrahelvøya appears to have begun to re-advance earlier (~3.5 cal kyr BP) (Røthe et al.,  
676 2015). Linnebreen also began to regrow much earlier in the Holocene (~4-5 cal kyr BP), but is believed to  
677 have reached its maximum Holocene extent during the 18-19<sup>th</sup> centuries (Svendsen and Mangerud, 1997).  
678 Offshore marine records also suggest broad cooling conditions ~1.0 cal kyr BP. Werner et al. (2015) posit  
679 that elevated advection of North Atlantic waters into the eastern Fram Strait region likely ended around  
680 1.0 kyr BP. Evidence for seasonal sea ice and unstable oceanographic conditions was found south of  
681 Svalbard in the western Barents Sea after 1.1 kyr BP (Berben et al., 2014) and Hald et al. (2004) point to  
682 a peak in IRD off western Svalbard at 0.8 cal kyr BP.

683 It remains unclear whether colder summer temperatures or increased wintertime precipitation led  
684 to the reemergence of Annabreen. Temperature reconstructions from Kongressvatnet (D'Andrea et al.,  
685 2012), as well as from Lake Skardtjørna (Velle et al., 2011), both on western Svalbard, suggest that  
686 summer temperatures were relatively stable during the past 1800 years, prior to recent anthropogenic  
687 warming.

688

### 689 **6) Conclusions**

690 This paper reports a Holocene reconstruction of the Annabreen glacier from Lake Gjøavatnet on  
691 Amsterdamøya, NW Svalbard. We show that sedimentation was dominated by glacial activity in the  
692 catchment during the early and late Holocene and likely responded to regional-scale oceanographic

693 changes in the intervening period. The early Holocene interval in Gjøavatnet is characterized by a two -  
694 phase sedimentation history, with Annabreen retreating out of the lake basin ~11.1 kyr BP and shrinking  
695 to its minimum Holocene extent by ~8.4 kyr BP. During the period from ~8.4-1.0 cal kyr BP it seems  
696 likely there was no glacial influence on sedimentation in Gjøavatnet. The organic-rich sediments  
697 deposited during this time are interrupted by at least three intervals of lower organic matter content (c.  
698 5.9-5.0 kyr BP, 2.6 – 2.2 kyr BP, and 1.7 – 1.5 kyr BP). We interpret the reductions in organic carbon  
699 content as periods of more extensive summer lake ice cover (due to colder summer conditions) related to  
700 changing oceanographic conditions in Fram Strait. At c. 1.0 cal kyr BP Annabreen re-advanced to the  
701 extent that it dominated sedimentation in the lake.

702 Maritime lakes in places like Svalbard provide the opportunity to capture both local glacier  
703 fluctuations along with offshore oceanographic conditions. Marine records are generally lower resolution  
704 than their lacustrine counterparts, and incorporate information from the entire water column, potentially  
705 complicating the detection of short-lived oceanographic events. Ocean based reconstructions over the  
706 Holocene near Svalbard suggest a complex and variable water mass and climate history, which we  
707 propose can be informed by the high-resolution lacustrine record from Gjøavatnet. While more research  
708 from both the terrestrial and marine realms would help confirm the climate patterns identified here, this  
709 study provides a unique record of both local glacial fluctuations and offshore oceanographic conditions  
710 spanning the Holocene.

711

712

### 713 **Acknowledgments**

714 Permission to perform field work in this region of Svalbard was granted by the Governor of  
715 Svalbard (RIS ID 5155, ref: 2012/00753-11 a.512). We thank Marthe Gjerde, Willem van der Bilt, and  
716 Sædis Ólafsdóttir for assistance with field work and Marthe and Willem for discussions on the  
717 manuscript. We also thank two anonymous reviewers for their constructive comments. Funding was  
718 provided by the Svalbard Science Forum (AFG project no. 235919) as well as the Norwegian Research

719 Council via the SHIFTS project. We thank Eivind Storen and Jordan Donn Holl for laboratory assistance  
720 at the University of Bergen, and Nicole DeRoberts for laboratory assistance at Lamont Doherty Earth  
721 Observatory.

722

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