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Diatom composition and fluxes over the Northwind Ridge, western Arctic Ocean: impact of marine surface circulation and sea ice distribution

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1 **Diatom composition and fluxes over the Northwind Ridge, western**
2 **Arctic Ocean: impact of marine surface circulation and sea ice**
3 **distribution**

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47 **Abstract**

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49 Over the last decades the western Arctic Ocean has undergone unprecedented
50 environmental changes. However, long-term marine phytoplankton *in situ* observations are still
51 rare and therefore insufficient to fully characterize evolutionary trends. This study investigate
52 diatom flux and composition in sediment trap material collected in the Northwind Ridge,
53 western Arctic Ocean from August 2008 to September 2009. Our data show that *Chaetoceros*
54 resting spores are the predominant species accounting for >40% of the diatom composition.
55 The sea ice diatom group, which includes *Fossula arctica*, *Fragilariopsis cylindrus* and *F.*
56 *oceanica*, dominates the rest of the assemblage throughout the observation period. While the
57 diatom fluxes in winter are extremely low, higher values are found in summer, with summer
58 2009 flux values being twice as high as in 2008. High total mass and diatom fluxes in summer
59 2009 are attributed to the intertwined effect of a weakened Beaufort Gyre, strengthened Pacific
60 Water Inflow (PWI) and distribution pattern of the sea ice. Enhanced values of coastal diatoms
61 and terrigenous proxies in summer 2009 are in agreement with intensified PWI. Sea ice diatoms
62 and sea ice biomarker IP₂₅ fluxes are both high during the sea ice melting season and
63 significantly correlated ($r^2 = 0.64$, $p < 0.01$). Our data also suggest that sea ice diatoms are
64 prone to selective dissolution in the water column and sediments, implying biases on diatom
65 assemblages and subsequently on paleoceanographic reconstructions.

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79 **Key words:** Diatoms, Sediment trap, Chukchi Sea, Northwind Ridge, Sea ice, Pacific water
80 inflow, Beaufort Gyre

81 1. Introduction

82 Over the past decades, the Arctic Ocean has undergone unprecedented environmental
83 changes with drastic impacts on the terrestrial (Post et al., 2013) and marine ecosystems (Carroll
84 and Carroll, 2003; Grebmeier et al., 2006, 2012; Arrigo et al., 2008; Wassmann, 2011;
85 Wassmann et al., 2011). Among other factors, the decline of sea ice has been shown to affect
86 marine phytoplankton production and distribution patterns (e.g. Arrigo et al., 2008; McLaughlin
87 et al., 2011; Coupel et al., 2012; Arrigo and van Dijken, 2015; Renaut et al., 2018). Blooms in
88 autumn are now reported during the prolonged ice-free period in the Arctic (Ardyna et al., 2014)
89 as well as significant changes in phytoplankton community due to enhanced freshening caused
90 by global warming (Li et al., 2009; He et al., 2012; Coupel et al., 2015; Zhuang et al., 2018;
91 Lee et al., 2019).

92 The highly productive Chukchi Sea in the western Arctic has experienced rapid sea ice
93 retreat (Comiso, 2012; Serreze et al., 2016) leading to increased primary production and
94 enhanced biological carbon uptake (Harada, 2016). The strengthening of the Pacific Water
95 Inflow (PWI) bringing nutrients into the Chukchi Sea during the last decade (Woodgate et al.,
96 2012; Woodgate, 2018) is another factor that may have altered the phytoplankton structure
97 (Zhuang et al., 2016). The Chukchi Sea is thus a critical region for understanding oceanographic
98 on-going and future changes in the Arctic Ocean and their consequences on polar ecosystems.

99 Apart from flagellates, diatoms appear to dominate the phytoplankton and bottom sea ice
100 community in the Chukchi Sea (Booth and Horner, 1997; Sukhanova et al., 2009; Poulin et al.,
101 2011; Joo et al., 2012). Because of diatom sensitivity to environment changes, their composition,
102 abundance, distribution as well as size have been used as indicators of marine environmental
103 conditions (Zernova et al., 2000; Smol and Stoermer, 2010; Romero and Armand, 2010). For
104 instance, recent studies speculated that the occurrence of the endemic North Pacific diatom
105 *Neodenticula seminae* in the North Atlantic might reflect stronger PWI driven by sea ice loss
106 (Reid et al., 2007; Poulin et al., 2010; Bluhm et al., 2011). Better understanding of the
107 relationship between diatoms and environmental variables today is also essential to produce
108 robust paleoceanographic reconstructions based on fossil diatoms in Arctic sediments (e.g. Ren
109 et al., 2009; Sha et al., 2016, 2017; Miettinen, 2018). Paleo-records are still very sparse in the
110 western Arctic in part because of our poor knowledge of modern diatom ecology and the
111 complex deposition pathways in this environment.

112 Sediment traps have been extensively deployed throughout the world ocean to study
113 biological processes and carbon export (Romero and Armand, 2010). In the Arctic, sediment
114 trap deployments have focused on the continental shelves and slopes (e.g. Wassmann et al.,
115 2004 and references therein; O'Brien et al., 2006, 2011; Fahl and Nöthig, 2007; Forest et al.,

116 2007, 2008, 2011; Lalande et al., 2009; Fahl and Stein, 2012) as well as the deep basins (Honjo
117 et al., 2010; Hwang et al., 2015; Lalande et al., 2019). Few studies have been undertaken in the
118 Chukchi Sea. Analyses of the ice-tethered sediment traps over the Chukchi Rise revealed the
119 heterogeneous patterns of biological processes and variety of particulate organic carbon sources
120 (Honjo et al., 2010). Recent studies of the first multi-years mooring observations in the
121 Northwind Abyssal Plain have documented seasonal variations of phyto- and zooplankton
122 (Matsuno et al., 2014; Ikenoue et al., 2015; Onodera et al., 2015, 2016; Tokuhiko et al., 2019)
123 and also evidenced early winter high flux induced by cold eddies from the shelf-break
124 (Watanabe et al., 2014, 2015).

125 In this study, we report on seasonal variations of diatom vertical flux obtained from the
126 sediment trap mooring deployed over the Northwind Ridge, Chukchi Borderland from August
127 2008 to September 2009 as part of the Chinese National Arctic Research Expedition
128 (CHINARE) program. These data are compared to the biomarker data of Bai et al. (2019)
129 measured in the same samples and discussed in light of the overlying water mass characteristics,
130 sea ice concentrations and atmospheric circulation in the western Arctic Ocean.

131 **2. Oceanographic settings**

132 Station DM is located at the southern Northwind Ridge, a steep ledge extending from the
133 shallow Chukchi Shelf into the Arctic deep basin as a part of the Chukchi Borderland (Fig. 1).
134 The surface hydrology of the region is influenced by the dynamic Beaufort Gyre (BG)
135 circulation, the PWI entering the Arctic Ocean through the Bering Strait and the seasonal sea
136 ice coverage.

137 The anticyclonic BG circulation is driven by the Beaufort High (Proshutinsky and Johnson,
138 1997). Over the last two decades, the stronger BG circulation has resulted in enhanced
139 freshening in the Canadian Basin (Giles et al., 2012) with significant impact on the
140 phytoplankton productivity and polar ecosystem (He et al., 2012; Coupel et al., 2015).

141 The PWI is primarily driven by a sea level difference between the Bering and Chukchi Seas
142 (Coachman and Aagaard, 1966), leading to an average northward water transport of ~ 0.8 Sv
143 (Roach et al., 1995). The PWI splits into three water masses in the Chukchi Sea, i.e. the saline
144 and nutrient-rich Anadyr Water (AW) as the western branch, the fresher and nutrient-depleted
145 Alaska Coastal Water (ACW) to the East, and in between the Bering Shelf Water (BSW) of
146 intermediate salinity (Fig. 1; Woodgate et al., 2005a; Grebmeier et al., 2006). In the past two
147 decades, long-term mooring observations in the Bering Strait have shown a gradual increase of
148 the PWI volume transport ($\sim 70\%$ from 2001 to 2014; Woodgate et al., 2012; Woodgate, 2018),
149 entraining more heat and nutrients into the Arctic Ocean. In the western Chukchi Sea, the
150 fresher Siberian Coastal Current (SCC) periodically interacts with the AW.

151 Sea ice occurs in the study area from November to July while it reaches a minimum over
152 the northern Chukchi Sea in September (Fig. 1). Several studies have shown that sea ice
153 reduction induced by PWI impacts on marine ecosystem (e.g. Shimada et al., 2006; Harada,
154 2016; Serreze et al., 2016). It has also been hypothesized that sea ice motion is also correlated,
155 at least partially, with the Arctic Oscillation (AO) (Rigor et al, 2002; Shimada et al., 2006).

156 **3. Material and methods**

157 *3.1 Sediment trap samples*

158 A one year-round mooring was deployed at Station DM (DM hereafter) on the southern
159 Northwind Ridge, in the vicinity of the Canada Basin, from August 2008 to September 2009
160 (Fig. 1; 74°24.0' N, 158°14.0' W, 1650 m water depth). This conical sediment trap (McLane
161 PARFLUX Mark 78H-21) was equipped with 21 sampling cups and installed at ~870 m water
162 depth. Samples were generally collected every two weeks (15 or 16 days, depending on the
163 month) during the diatom production season (July to November) while during the low flux
164 period in winter (December to June) sampling intervals were longer (28 to 31 days, depending
165 on the month; Bai et al., 2019). The 21 sampling cups were filled with artificial seawater
166 (salinity~35) and antiseptic HgCl₂ to preserve trap material from degradation.

167 After recovery, the wet samples were sieved using a 1-mm mesh nylon sieve to remove
168 swimmers. The fine fractions (<1 mm) were split into aliquots with a McLane wet sample
169 divider (WSD-10). An aliquot (1/4 or 1/8) was filtered on a polycarbonate filter (0.45 µm pore
170 size, 47 mm diameter) for diatom and biochemical analyses. Samples were then dried in an
171 oven at 45 °C for 72 h (Bai et al., 2019).

172 *3.2 Diatom preparation and analysis*

173 Quantitative diatom slides were prepared according to the standard method developed at
174 the Alfred Wegener Institute, Germany (Gersonde and Zielinski, 2000). For each cup, 30 mg or
175 15 mg sub-sample was used. Although only 1 mg was available for cup 14# (May 2009), the
176 sample was analyzed to keep the entire record intact. The subsamples were then treated with
177 HCl (~36% w/w concentration) and H₂O₂ (~30% w/w concentration) to remove calcareous and
178 organic material. After rinsing, a known aliquot of the residue was evenly dropped on a cover
179 glass (24 mm × 24 mm). Permanent slides were mounted with Naphrax for diatom identification
180 (n_D ~1.7; Fleming, 1954).

181 Around 400 diatom valves (max. 604.5, min. 372) were counted for each sample with a
182 Motic BA400E microscope at ×1000 magnification. For cup 14# only 215 diatom valves could
183 be counted. Diatom valves were counted according to the method of Schrader and Gersonde
184 (1978). Diatoms were identified to species level following Medlin and Priddle (1990), Hasle

185 and Syvertsen (1997), Suto et al. (2004), Katsuki et al. (2009), Obrezkova et al. (2014) and
186 Tsoy and Obrezkova (2017). Because resting spores of genus *Chaetoceros* can hardly be
187 identified at a species level by light microscopy, they were combined and labeled as
188 *Chaetoceros* resting spores (RS).

189 The diatom counts were converted into relative abundances with respect to the total diatom
190 assemblage. The diatom concentration (valves per gram dry sample) was estimated using the
191 following formula (Esper et al., 2010):

$$192 \quad \text{Diatom concentration} = (1/dw) \times (csa/ta) \times (sv/split) \times (vn/tn) \quad (1)$$

193 in which, dw is the dried sample weight in milligrams, csa is the area of the cover slide (576
194 mm²), ta is the area of one counted traverse (6 mm²), sv is the volume of processed diatom
195 suspension (6-15 ml), $split$ is the volume of aliquot dropped onto the cover slide (from 27 to 81
196 $\times 10^{-2}$ ml), vn is the total counted diatom valves and the tn is the number of fully counted
197 traverses.

198 The diatom fluxes were calculated by the following equation:

$$199 \quad \text{Diatom flux [valves } m^{-2} d^{-1}] \\ 200 \quad = \text{Total mass flux [mg } m^{-2} d^{-1}] \times \text{Diatom concentration [valves mg}^{-1}] \quad (2)$$

201 where the diatom concentration is derived from equation (1).

202 Diatom taxonomic diversity was evaluated by the H-index based on the Shannon-Weaver
203 formula (Shannon and Weaver, 1949):

$$204 \quad H = -\sum P_i \log_2 P_i \quad (3)$$

205 where H is the diversity index and P_i is the relative abundance of species i .

206 3.3 Biomarker analysis

207 The concentrations of IP₂₅ and brassicasterol (24-methylcholesta-5,22E-dien-3 β -ol)
208 published by Bai et al. (2019) are used for comparison with our diatoms fluxes to discuss sea
209 ice conditions. Campesterol (24-methylcholest-5-en- β -ol) and 24-ethylcholest-5-en-3 β -ol are
210 also used to assess higher plant inputs. All sterols were silylated with 100 μ L BSTFA (bis-
211 trimethylsilyl-trifluoroacet-amide) (80 $^{\circ}$ C, 1 h) prior to the gas chromatography (GC) analysis.
212 The GC analyses were carried out on a Varian 3300 with a septum programmable injector and
213 a flame ionization detector (Bai et al., 2019).

214 3.4 Carbonate and lithogenic matter

215 The CaCO₃ content of sinking material was calculated from particulate inorganic carbon
216 (PIC) values determined by the difference between total carbon (TC) and particulate organic

217 carbon (POC; Bai et al., 2019). The lithogenic matter (LM) was estimated as the difference
218 between the total mass (TM; from Bai et al., 2019) and the biogenic material composed of
219 CaCO₃, opal (Wu et al., unpublished data) and organic matter (OM) that conventionally convert
220 from POC by a factor of 1.8 (Müller et al., 1986):

$$221 \quad LM = TM - (CaCO_3 + Opal + OM) \quad (4)$$

222 The lithogenic matter percentage was then calculated by the following equation:

$$223 \quad LM (\%) = LM/TM \quad (5)$$

224 3.5 Environmental data

225 Daily environmental data over the study period at DM, including shortwave radiation, snow
226 thickness, sea ice cover and sea ice thickness, were from the National Centers for
227 Environmental Prediction (NCEP)/Climate Forecast System Reanalysis (CFSR) 6-hourly
228 dataset (Saha et al., 2010, 2014).

229 Monthly Pacific inflow data, e.g. volume transport, temperature and salinity, were acquired
230 from the “Bering Strait: Pacific Gateway to the Arctic” project (Woodgate et al., 2015;
231 Woodgate, 2018).

232 Regional sea ice concentrations were from Nimbus-7 SMMR and DMSP SSM/I-SSMIS
233 Passive Microwave Data with a grid resolution of 25 × 25 km (Cavalieri et al., 1996).

234 Summer (July, August, September) monthly averaged sea surface chlorophyll-*a*
235 concentrations for a 2° × 2° area around DM were retrieved from the Moderate Resolution
236 Imaging Spectroradiometer (MODIS) with a 4 × 4 km resolution and calculated by algorithm
237 of Gohin et al (2002) collected by GlobColour data (<http://globcolour.info>).

238 4. Results

239 4.1 Environmental conditions

240 Station DM experiences polar night from November to mid-February and receives
241 maximum solar irradiance of ~369 W m⁻² in summer, from mi-April to mid-July (Fig. 2a).
242 Seasonal sea ice was present from late October 2008 to mid-August 2009. The ice-free season
243 (sea ice concentration <15%) is approximately one month longer in 2008 than in 2009 (mid-
244 July to late October vs. mid-August to early October, respectively; Fig. 2b). Winter sea ice
245 thickness over the station is 2.19 m on average with a maximum thickness of 2.79 m recorded
246 in March 2009 (Fig. 2c). Similarly, snow on ice appears in mid-October 2008 and nearly has
247 thawed out early June 2009 (Fig. 2d). The snow accumulation is over 25 cm in winter.

248 4.2 Total mass flux

249 The seasonal pattern of the total mass fluxes (TMF) earlier described by Bai et al. (2019),
250 show higher values in August and September, both in 2008 and 2009 (Fig. 2e). However, the
251 TMF in September 2009 is ca. eightfold larger than in 2008. Particulate organic carbon (POC)
252 fluxes exhibit similar fluctuations as TMF, with extremely low values in winter and early spring,
253 from November 2008 to June 2009 (Fig. 2f). One exception to this is the POC flux value of
254 $3.51 \text{ mg m}^{-2} \text{ d}^{-1}$ in April 2009, representing approximately 12% of the TMF as compared to an
255 annual mean value of 4% (Bai et al., 2019).

256 4.3 Lithogenic matter flux

257 In few samples, the PIC content is negative as a result of POC values exceeding the TC
258 values ($\text{PIC} = \text{TC} - \text{POC}$). These numbers reflect the very low PIC content of these samples
259 ($<1\%$), and uncertainties on TC and POC measurements. Thus the CaCO_3 contents based on
260 these PIC values became negative. In general, CaCO_3 is less than 10% of the total mass and has
261 thus no significant impact on the lithogenic matter content. Accordingly, we revised the
262 negative carbonate values to zero. The derived lithogenic matter fluxes reach $225 \text{ mg m}^{-2} \text{ d}^{-1}$
263 and $\sim 600 \text{ mg m}^{-2} \text{ d}^{-1}$ in summer of 2008 and 2009, respectively, while low values ($<50 \text{ mg m}^{-2}$
264 d^{-1}) are recorded in winter (Fig. 2e). On average, the lithogenic matter accounts for $>70\%$ of
265 the total mass with no clear seasonal pattern (Fig. 2e).

266 4.4 Terrigenous sterol flux

267 Fluxes of 24-ethylcholest-5-en-3 β -ol show similar variations as terrigenous campesterol
268 suggesting a primary terrestrial source of this sterol, thus most likely mainly represented by β -
269 sitosterol (Fig. 2g). Campesterol and β -sitosterol fluxes reach ~ 1 and $2.8 \text{ } \mu\text{g m}^{-2} \text{ d}^{-1}$ in August
270 2008, respectively. Both decrease to less than $0.5 \text{ } \mu\text{g m}^{-2} \text{ d}^{-1}$ in the following winter (Fig. 2g)
271 and increase to highest fluxes in summer 2009 peaking at $9.6 \text{ } \mu\text{g m}^{-2} \text{ d}^{-1}$ in July for β -sitosterol,
272 and to $8.4 \text{ } \mu\text{g m}^{-2} \text{ d}^{-1}$ in September for campesterol.

273 4.5 Diatom flux and species composition

274 Total diatom fluxes are strongly seasonal (Fig. 2h). Two peak values occur early August
275 2008 ($\sim 10 \times 10^6 \text{ valves m}^{-2} \text{ d}^{-1}$) and late July to September 2009 ($> 14 \times 10^6 \text{ valves m}^{-2} \text{ d}^{-1}$),
276 respectively. The diatom fluxes in winter are two orders of magnitude lower ($< 0.5 \times 10^6 \text{ valves}$
277 $\text{m}^{-2} \text{ d}^{-1}$). They rise abruptly late July 2009 when solar irradiance is nearly at maximum and sea
278 ice starts retreating (Fig. 2a, b, h).

279 In total 93 diatom species or species groups were identified in the sediment trap samples.
280 Apart from *Chaetoceros* RS, diatoms were categorized into 4 groups based on their ecological
281 preferences (Medlin and Priddle, 1990; Hasle and Syvertsen, 1997; von Quillfeldt, 1997, 2000;
282 Kohly, 1998; von Quillfeldt et al., 2003; Onodera et al., 2015). The sea ice group mainly

283 consists of *Fossula arctica*, *Fragilariopsis cylindrus*, *F. oceanica*, *Nitzschia frigida*, *Pauliella*
284 *taeniata* and certain species of genera *Diploneis*, *Navicula*, *Nitzschia*, *Pinnularia*. The cold-
285 water diatom group is mostly composed of *Thalassiosira antarctica* var. *borealis* RS, *T.*
286 *nordenskioldii* and, in minor amounts, *Bacterosira bathyomphala* RS and *T. bulbosa*. The
287 cosmopolitan diatom group includes *Chaetoceros atlanticus*, *Thalassionema nitzschioides* and
288 *Thalassiosira eccentrica*, while *Paralia sulcata* and species of genera *Cocconeis*, *Cyclotella*,
289 *Cymbella*, *Delphineis* and *Eunotia* form the coastal diatom group. The remaining diatom
290 species make up the other group.

291 *Chaetoceros* RS is the predominant group over the entire observation period (Fig. 3a). It
292 increases from 40% in August 2008 to ~80% in late September 2009. In contrast, the sea ice
293 diatom group shows a broad decline since the deployment of the trap (Fig. 3b-e). Within this
294 group, the relative abundances of *F. arctica*, *F. cylindrus* and *F. oceanica* are two- to three-fold
295 higher in summer 2008 than in 2009, whereas *P. taeniata* peaks in November 2008. The cold-
296 water diatom group, however, remains low throughout the study period (Fig. 3f-g). Unlike the
297 sea ice diatom species, the cold-water *T. antarctica* var. *borealis* RS, *T. bulbosa* and *T.*
298 *nordenskioldii* are more abundant in summer 2009 than in summer 2008. Interestingly, *T.*
299 *nitzschioides*, the major representative of the cosmopolitan diatom group, is more frequently
300 observed in winter than in summer (Fig. 3h). Similarly, the coastal diatom group, dominated by
301 *P. sulcata*, is also found in low relative abundances with slightly higher values from October
302 2008 to April 2009 when sea ice prevails (Fig. 3i).

303 Among the sea ice diatoms, *Haslea crucigeroides* and *Pleurosigma stuxbergii* var.
304 *rhomboides*, that are known producers of sea ice biomarker C₂₅ monounsaturated hydrocarbon
305 IP₂₅ (Brown et al., 2014), are detected though in low relative abundance (<1%), in both August
306 2008 after sea ice just retreated, and in July - early August 2009 at the onset of sea ice thaw
307 (not shown). Although widely observed in the Arctic Ocean, the sea ice bottom tethered diatom
308 *Melosira arctica* is only found sporadically over this one-year experiment (Abelmann, 1992;
309 Zernova et al., 2000; von Quillfeldt et al., 2003; Boetius et al., 2013; Lalande et al., 2014, 2019).

310 The North Pacific predominant and endemic taxa *Neodenticula seminae* (Jousé et al., 1971;
311 Sancetta, 1982; Onodera and Takahashi, 2009; Ren et al., 2014), which has been recently
312 reported in the North Atlantic water column and surface sediment (Starr et al., 2002; Reid et al.,
313 2007; Miettinen et al., 2013), is also found once in sinking particles at DM, in line with the data
314 obtained at the nearby sediment trap NAP site (Onodera et al., 2015).

315 4.6 Diatom taxonomic diversity

316 The diatom diversity index, H-index, decreases from a mean of ~2.6 in 2008 to ~2.3 in 2009
317 with an outlier value of ~3.9 in early July 2009 (Fig. 3j). The summer 2009 diversity minimum

318 is reached at maximum TMF and diatom fluxes (Fig. 2e, h, 3j). The H-index of summer 2009
319 is lower than found in summer 2008 (H-index: 1.8 vs 2.7) most probably due to the high
320 abundance of *Chaetoceros* RS and decrease of other species.

321 **5. Discussion**

322 *5.1 Interannual variability of diatom fluxes: link to surface circulation and sea ice*

323 Our sediment trap time-series shows high TMF and diatom flux values during the summer
324 season. However, in summer 2009 and particularly late September, fluxes are two- to three-fold
325 higher than measured in 2008 (Fig. 2e, h; Bai et al., 2019). Diatom fluxes in summer 2008
326 ($\sim 10 \times 10^6$ valves $\text{m}^{-2} \text{d}^{-1}$) are comparable to those found during the same season in the nearby
327 station NAP at 180/260 m depth in 2011 ($\sim 11 \times 10^6$ valves $\text{m}^{-2} \text{d}^{-1}$), regardless of the mooring
328 depths and collection years of each site (Onodera et al., 2015). Interestingly, the high diatom
329 flux values in summer 2009 at DM (14 to 27×10^6 valves $\text{m}^{-2} \text{d}^{-1}$) are of similar magnitude as
330 winter high flux events at NAP attributed to shelf-break eddies (Watanabe et al., 2014; Onodera
331 et al., 2015).

332 Lower diatom fluxes in 2008 than 2009 may suggest that the summer bloom was not
333 captured (Fig. 2-3) and that these low values reflect post-bloom production. Indeed, an early
334 bloom was reported in May 2008 in the Amundsen Gulf, East of the Beaufort Sea (Brown et
335 al., 2011; Belt et al., 2013) although this does not imply that the Northwind Ridge experienced
336 similar production growth. By contrast, a deep sediment trap in the southwestern Canadian
337 Basin indicates high POC values in August and September 2008, which based on a sinking
338 velocity of 85 m d^{-1} , corresponds to high surface production and export in July - August (Station
339 A in Fig. 1; Hwang et al., 2015; Onodera et al., 2015). Furthermore, ice-ocean-ecosystem
340 coupled model simulations point out peaking chlorophyll-*a* values in the Chukchi Sea in August
341 2008 (Wang et al., 2013). Therefore, we can reasonably infer that the diatom flux measured in
342 August 2008 represents the blooming season locally and that it was thus lower than in 2009.
343 This assumption is further supported by MODIS summer sea surface chlorophyll-*a*
344 concentrations of the area indicating $\sim 36\%$ higher values in 2009 ($0.382 \mu\text{g l}^{-1}$) than 2008 (0.282
345 $\mu\text{g l}^{-1}$).

346 Biogenic production at DM is influenced by the PWI, the oligotrophic waters of the BG and
347 the sea ice distribution during the blooming season (Fig. 1). The Arctic Oscillation (AO) plays
348 an important role on sea ice distribution and surface ocean circulation (Rigor et al., 2002;
349 Serreze et al., 2003; Steele et al., 2004).

350 Since 1997, the BG circulation is driven by an anticyclonic wind circulation resulting in
351 enhanced freshening of the BG waters (Giles et al., 2012; Rabe et al., 2014; Proshutinsky et al.,
352 2015; Wang et al., 2018). The subsequent stratification prevents nutrient to replenish the upper

353 ocean and thus stalls phytoplankton growth (McLaughlin and Carmack, 2010). The strength of
354 the BG is linked to the Arctic Ocean Oscillation (AOO) index, which varied annually and
355 notably shifted from a strong BG (AOO index = 2.9) in 2008 to a weaker BG in 2009 (AOO
356 index = 1.9; Fig. 4a) (Proshutinsky et al., 1999, 2015). We can thus hypothesize that
357 phytoplankton productivity at DM in 2008 was limited by stratified conditions maintaining
358 oligotrophic conditions in the upper ocean due to freshening under intensified BG. Likewise,
359 weakened and reduced BG in 2009 provided more favorable conditions for refueling surface
360 waters in nutrients to sustain higher production. This is consistent with fresher surface waters
361 reported near DM in summer 2008 than summer 2009 (Zhang, 2009; Kikuchi, 2009). Model
362 results also show prevailing west-northwestward surface flow in the Chukchi Borderland in
363 summer 2008 carrying nutrient-depleted waters from the Beaufort Sea to DM, whereas in
364 summer 2009 the surface ocean circulation was nearly northward bringing nutrient-rich PWI
365 waters to our site (Onodera et al., 2015). These findings highlight the key role of atmospheric
366 conditions in driving the surface current and nutrients transport at the BG boundary and their
367 impact on diatom flux.

368 The PWI is regarded as a pivotal nutrient source for the Arctic Ocean ecosystem (Walsh et al., 1997).
369 The mean annual transport of the PWI through the Bering Strait (~ 0.8 Sv) fluctuates
370 seasonally as well as annually (Roach et al., 1995; Woodgate et al., 2015) around values that
371 regularly increased since 1990 (Woodgate et al., 2012; Woodgate, 2018). The PWI in summer
372 2009 (~ 1.7 Sv) was stronger than in 2008 (~ 1.35 Sv) (Fig. 4b). Furthermore, during its
373 maximum transport period (April to July) this flow was ~ 0.2 Sv higher in 2009 than 2008.
374 Based on the distance between the Bering Strait and DM and current velocities in the Chukchi
375 Sea (Weingartner et al., 2005; Woodgate et al., 2005b), we can estimate that ~ 4 months is
376 needed for the PWI to reach our mooring site, which is in agreement with the time lag between
377 the PWI peak and high biogenic production in summer 2009 (Fig. 4). It is thus very reasonable
378 to conclude that stronger PWI by entraining more nutrients into the Arctic Ocean likely
379 triggered enhanced phytoplankton production in 2009.

380 The sea ice distribution during summer also shows different patterns in 2008 and 2009 (Fig.
381 5). As expected from AO and subsequent ocean circulation, in 2008 sea ice started retreating
382 westwards from the East, favoring the inflow of freshened and oligotrophic BG waters.
383 Meanwhile, the presence of sea ice in the South likely prevented nutrient-rich PWI waters from
384 advecting to our site. In contrast, sea ice melting in 2009 began from the South allowing the
385 PWI to flow to DM and influence primary production. This is further supported by surface wind
386 fields shown in Fig. S1. Easterly winds in July 2008 favored transport of freshened and
387 oligotrophic waters from the BG towards to the northwest or North of DM thereby limiting
388 biogenic production. In August 2008, southward winds further contributed to counteract the

389 northward advection of PWI, whereas in summer 2009 (July and August) Northerly winds
390 enhanced transport of nutrient-rich PWI to our mooring site, favoring diatom blooming (Fig.
391 S1)

392 In summary, our results provide consistent evidences on the impact of contrasting
393 atmospheric and oceanic circulation, notably the BG circulation, PWI nutrient transport as well
394 as sea ice retreat pattern on diatoms fluxes in the Chukchi Sea, between 2008 and 2009 (Fig.
395 5). They also underline that the southern Northwind Ridge is an area of strong interactions
396 between water masses from the Canadian Basin, the Pacific and even coastal Siberia, where
397 seasonal sea ice drift and motion makes it a sensitive region to ongoing climate change.

398 5.2 Lateral transport of diatoms

399 Coastal diatoms, mostly dwelling in the continental shelf area, are indicative of lateral
400 advection from the nearby Chukchi Sea shelf waters. Previous studies have shown that while
401 *Paralia sulcata* prevails in the sediments of the Chukchi Sea Shelf, this species is rarely found
402 in the Chukchi Borderland (Ran et al., 2013). At DM, the coastal diatom fluxes are also
403 dominated by *P. sulcata* peaking in September 2009, pointing out a significant influence of
404 water masses from the shelf area under stronger PWI conditions (Fig. 6b; See 5.1; Woodgate,
405 2018). Apart from this episode, values were low and stable (Fig. 6b).

406 *Chaetoceros* RS is also considered as an indicator of allochthonous material (Onodera et al.,
407 2015). In our dataset *Chaetoceros* RS displays similar seasonal pattern as terrigenous biomarker
408 campesterol and β -sitosterol (Fig. 6c-d). The flux of these biomarkers produced by higher plants
409 (Huang and Meinschein, 1976) were found to also parallel lithogenic matter fluxes in both
410 shallow and deep sediment traps of the central Arctic Ocean (Fahl and Nöthig, 2007). The co-
411 eval temporal evolution of terrigenous sterols and lithogenic material at DM is consistent with
412 lateral advection from the shelf in summer of 2009 (Fig. 6d-e). In contrast, the lithogenic matter
413 content (in %) drops to its lowest values in summer 2009 witnessing a reduced contribution of
414 allochthonous material (Fig. 6b-d, f). This discrepancy cannot be explained by sea ice transport
415 of coastal diatoms and biomarkers in winter and later settling under ice-free conditions in
416 September 2009 (Fig. 6a). Besides, this mechanism cannot account either for the concomitant
417 decrease of lithogenic matter content in summer 2009. Another explanation is that coastal
418 diatoms and terrigenous material might be entrained by PWI. Enhanced PWI would thus be
419 responsible for higher phytoplankton production in summer 2009 and the advection of
420 terrigenous matter (See 5.1). This is further supported by the decrease of the C/N ratio (bulk
421 organic carbon over bulk organic nitrogen mole ratio; modified from Bai et al, 2019) to 6.4 in
422 summer 2009. This value, close to the Redfield ratio of 6.6, points out a major marine origin of
423 settling material (Fig. 6g). Boosted blooming production and export of organic matter likely

424 explain enhanced contribution of marine constituents to the overall flux and the subsequent
425 decline of lithogenic content (Fig. 6f).

426 The abundance and flux of lithogenic matter, mostly composed of silt and clay, are higher
427 at our site than at nearby mooring station NAP and LOMO2 (Fig. 1) in the central Arctic, despite
428 a difference in water depths. Lithogenic particles might be derived from a proximate branch of
429 the Northwind Ridge, lying at ~900 m water depth (Poore et al., 1994), close to the depth of
430 our sediment trap (~870 m). Benthic nepheloid layer developing hundreds meters above the
431 seafloor is another possible mechanism able to remobilize sediments from the Northwind Ridge
432 and transport them to the sediment trap location (Rutgers van der Loeff et al., 2002).

433 5.3 Sea ice diatoms and IP_{25}

434 Fluxes of sea ice diatoms and IP_{25} at DM display quite similar seasonal patterns (Fig. 7a).
435 Since IP_{25} is produced by specific sea ice diatoms (Brown et al., 2014), some correlation is
436 expected between the two fluxes ($n = 21$; $r^2 = 0.64$; $p < 0.01$; Fig. 7b; Table 2). This correlation
437 is statistically unchanged after discarding low flux values ($IP_{25} < 0.5 \text{ ng m}^{-2}\text{d}^{-1}$; sea ice diatom
438 $< 1 \times 10^5 \text{ valves m}^{-2} \text{ d}^{-1}$) ($n = 13$; $r^2 = 0.61$; $p < 0.01$). Both fluxes are high during the sea ice
439 melting season for both years. Although sea ice reconstructions based on micropaleontological
440 fossils or biomarkers preserved in the sediment cores have been successfully carried out in the
441 Atlantic and Arctic Ocean (e.g. diatoms: Sha et al., 2016, 2017; IP_{25} : Massé et al., 2008; Müller
442 et al., 2009; Vare et al., 2009; Müller and Stein, 2014; Belt et al., 2015; Xiao et al., 2015;
443 Cabedo-Sanz et al., 2016; Hörner et al., 2016, 2017; Stein et al., 2016, 2017; Kolling et al.,
444 2017; Clotten et al., 2018; Kremer et al., 2018), they have rarely been produced and compared
445 in the same core. Few available studies show poor agreement between sea ice diatoms and IP_{25} ,
446 a result that might be attributed to the specificity of IP_{25} producers as opposed to the broader
447 habitat of sea ice diatoms (Weckström et al., 2013; Sha et al., 2015). Our sediment trap record,
448 however, indicates consistent and synchronous fluctuations of the sea ice diatom group and IP_{25}
449 fluxes with respect to sea ice distribution (Fig. 7a). Departures of diatom versus biomarker
450 signals in sediments can, to some extent, be attributed to dissolution and degradation (Belt and
451 Müller, 2013). Previous studies in the Arctic have shown that IP_{25} is biosynthesized by sea ice
452 diatoms *Haslea crucigeroides* and *Pleurosigma stuxbergii* var. *rhomboides* (Belt et al., 2013;
453 Brown et al., 2014). These species are present in our sediment trap in summer mainly during
454 high IP_{25} fluxes (Fig. 7a). Although these species have been frequently reported at the bottom
455 sea ice and, to a lesser extent, in the water column in the Arctic (e.g. von Quillfeldt, 1997; von
456 Quillfeldt et al., 2003; Katsuki et al., 2009), they were identified in only a few summer samples
457 at ~870 m, possibly because of dissolution during settling. These species have seldomly been
458 found in Arctic sediments (e.g. Ran et al., 2013; Obrezkova et al., 2014; Astakhov et al., 2015;

459 Limoges et al., 2018). The non-IP₂₅ producing sea ice diatoms, e.g. *Fossula arctica*,
460 *Fragilariopsis cylindrus*, *F. oceanica*, also suffer from dissolution. Sediment trap data from the
461 Nordic Seas have shown that weakly silicified diatoms strongly dissolve in the sediments while
462 the robust ones are preserved (Kohly, 1998). Selective degradation of diatom frustules was also
463 demonstrated by comparing diatoms from shallow and deep traps at the nearby mooring site
464 NAP (Onodera et al., 2015). The relative abundance of *Rhizosolenia* spp., for instance,
465 decreased remarkably in the deep trap, whereas some diatom taxa increased in the abyssal
466 material (Onodera et al., 2015). In another study in the Chukchi Sea, the diatom assemblages
467 in sea ice and underlying sediments showed significant differences (von Quillfeldt et al., 2003).
468 Despite the fact that the diatom composition and species diversity of sediment trap at DM are
469 comparable to those of the sediments in the northern Chukchi Sea, weakly silicified diatoms,
470 *Chaetoceros* RS and sea ice diatoms among others, became less abundant in the sediments
471 while diatoms with stronger frustules such as *Thalassiosira antarctica* var. *borealis* RS
472 increased (Table 1; Ran et al., 2013). This finding supports the idea of selective diatom
473 dissolution during export to deep sediments, which pleas for cautious interpretation of paleo-
474 reconstructions. In contrast, given its low photo-reactivity and wide distribution in the Arctic
475 sediments, IP₂₅ should introduce less bias due to diagenesis (Brown and Belt, 2012; Belt and
476 Müller, 2013). Although a possible biogeochemical degradation of IP₂₅ was suggested in the
477 central Arctic based on shallow and deep sediment traps, the IP₂₅ seasonal patterns remained
478 similar from 150 to 1550 m water depth (Fahl and Stein, 2012). Nevertheless, investigations
479 are needed to explore IP₂₅ degradation and its possible impact on past sea ice reconstruction.

480 Major phytosterols, namely brassicasterol and dinosterol, show a high correlation with cold-
481 water diatoms ($r^2 = 0.83$ and 0.77 , respectively; $p < 0.01$; Table 2). From this result we can infer
482 that both phytosterol producers and cold-water diatoms thrive in ice-free conditions (Bai et al.,
483 2019). Cold-water diatoms, however, have been also reported underneath consolidated sea ice
484 (Arrigo et al, 2012, 2014). In this case, it is expected that both cold-water and sea ice diatoms
485 sink synchronously and result in a significant correlation between their respective fluxes ($r^2 =$
486 0.64 ; $p < 0.01$; Table 2). Interestingly, the tri-unsaturated highly branched isoprenoid alkene
487 (HBI-III), presumably associated with marginal ice zone (Belt et al., 2015; Smik et al., 2016;
488 Smik and Belt, 2017; Bai et al., 2019), shows a weak correlation with the sea ice diatom flux
489 ($r^2 = 0.48$; $p < 0.01$; Table 2). Although HBI-III producers have not yet been fully recognized,
490 the hypothesized producers are within diatom genera *Pleurosigma* and *Rhizosolenia* (Belt, 2018;
491 Belt et al., 2000, 2015; Rowland et al., 2001). In this area diatom species belonging to
492 *Pleurosigma* are mainly related to sea ice while *Rhizosolenia* is found in the ice-free waters
493 (von Quillfeldt et al., 2003; Onodera et al., 2015). The weak correlation between HBI-III and
494 sea ice diatoms rather hints a sea ice source or sea ice edge conditions, but more investigation

495 is needed to be more conclusive.

496 Diatom sinking velocity of $>85 \text{ m d}^{-1}$ in summer has been recently estimated based on the
497 time lag between the shallow and deep trap diatom fluxes at NAP (Onodera et al., 2015). Using
498 this value, the summer diatom flux at $\sim 870 \text{ m}$ water depth at our site should represent the export
499 of material from surface waters or the sea ice produced about 10 days earlier. Hence, the diatom
500 bloom of summer 2009 might have started early July when the sea ice was about to retreat but
501 still compact (Fig. 2, 5). Incidentally, massive under sea ice phytoplankton blooms have been
502 observed in the Beaufort Sea in summer 2008 (Mundy et al., 2009) and in the Chukchi Sea in
503 summer 2011, maximizing at $>100 \text{ km}$ into the sea ice cover boundary (Arrigo et al., 2012,
504 2014). This phenomenon was attributable to intensified light penetration through sea ice into
505 the water column due to more melt ponds (Arrigo et al., 2014; Mundy et al., 2014). Thawing
506 snow over sea ice as well as leads in the ice pack might also have played a role in promoting
507 under ice bloom (Ambrose et al., 2005; Róžańska et al., 2009; Arrigo, 2017; Assmy et al., 2017;
508 Lalande et al., 2019). Although no melt ponds data exist at DM for summer 2009, on-site model
509 estimates indicate that the snow cover had almost vanished and the sea ice thickness decreased
510 from $\sim 2.8 \text{ m}$ to $\sim 1.5 \text{ m}$, at that time of maximum solar irradiance (Fig. 2a-d), leading to similar
511 sea surface state in summer 2009 as in 2011, allowing light penetration to trigger under sea ice
512 bloom.

513 **6. Conclusions**

514 The one-year long sediment trap time-series (August 2008 - September 2009, $\sim 870 \text{ m}$ water
515 depths) from the southern Northwind Ridge in the western Arctic Ocean highlights the strong
516 seasonality of diatom abundances and fluxes. *Chaetoceros* RS and, to a lesser extent, the sea
517 ice diatoms, e.g. *Fossula arctica*, *Fragilariopsis cylindrus*, *F. oceanica*, dominate the diatom
518 composition.

519 Diatom fluxes are two- to three-fold higher in summer 2009 than in summer 2008, a result
520 that is explained by a weaker BG circulation, an intensified PWI bringing nutrients into the
521 Arctic Ocean and the earlier retreat of sea ice from the South in 2009. Our findings underline
522 the sensitivity of the southern Northwind Ridge export fluxes to the spatio-temporal pattern of
523 sea ice melting and upper ocean circulation regime.

524 High fluxes of coastal diatoms, terrigenous biomarkers and lithogenic matter show indicate
525 lateral advection in summer 2009 likely triggered by enhanced PWI rather than by drifting sea
526 ice.

527 Sea ice diatom fluxes are coherent with IP_{25} fluxes, both showing high values during the
528 sea ice melting period reflecting marginal ice zone conditions. However, sea ice diatoms might
529 be prone to selective dissolution in the water column that is likely to alter diatom assemblage

530 in sediments and introduce bias in sea ice reconstructions. This result reinforces the need for
531 multi-proxy approaches to assess proxy biases and improve past sea ice estimates.

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548 **Author contributions**

549 J. R. designed the study and wrote the manuscript with contribution of M.-A. S., J. C., H.
550 L., J. Z., Y. B., Y. Z and B. W. J. R. carried out the diatom identification. Y. B. and Y. L.
551 contributed the biomarker analyses (campesterol and β -sitosterol), J. Z. performed the
552 determination of particulate inorganic carbon (PIC) and calculated the CaCO₃ content. Z. J.
553 and H. J. contributed the particulate organic carbon (POC) and B. W. the opal flux. Z. Y. and L.
554 L. retrieved the environmental data from different database. All authors contributed to the final
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557 **References**

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Diatom composition and fluxes over the Northwind Ridge, western Arctic Ocean: impact of marine surface circulation and sea ice distribution

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Figures captions

Figure 1. a) Location of Station DM (red circle) and surface currents in the study area (grey arrows). The white dash line represents the multi-year sea ice minimum for 1979-2017 (20% of sea ice concentration) (from Cavalieri et al., 1996). Two other sites for discussion are also presented (yellow circles). Currents system: PWI- Pacific Water Inflow; ACW-Alaska Coastal Water; AW-Anadyr Water; BSW-Bering Slope Water; BG-Beaufort Gyre; SCC-Siberia Coastal Current. Topography: NR-Northwind Ridge; NAP-Northwind Abyssal Plain; CR-Chukchi Rise; CAP-Chukchi Abyssal Plain; MR-Mendeleev Ridge. b) Location of the sediment traps discussed in the text. The red dot shows Station DM (this study), the yellow dots show the other sites from previous studies.

Figure 2. Time series of daily environment variables, diatoms and biogeochemical tracers measured in the sediment trap moored at Station DM from August 2008 to September 2009. a) Shortwave radiation ($W m^{-2}$); b) Sea ice concentration (%); c) Sea ice thickness (m); d) Snow thickness (cm). a-d) are derived from the National Centers for Environmental Prediction (NCEP)/Climate Forecast System Reanalysis (CFSR) 6-hourly dataset (Saha et al., 2010, 2014). e) Total mass flux in $mg m^{-2} d^{-1}$ (dark bar; from Bai et al., 2019), lithogenic matter flux in $mg m^{-2} d^{-1}$ (grey bar) and percent weight of lithogenic matter (brown line); f) Particulate organic carbon (POC) flux in $mg m^{-2} d^{-1}$ (Bai et al., 2019); g) Campesterol and β -sitosterol flux in $\mu g m^{-2} d^{-1}$ represented by orange and brown bars, respectively; h) Stacked flux of different diatom groups in $10^6 valves m^{-2} d^{-1}$ (see text). Note that the fluxes in e) and g) are not stacked.

Figure 3. a-i) Diatom fluxes (in $10^5 valves m^{-2} d^{-1}$) and relative abundances (in %) of the main species and species groups (see text) and j) diatom species diversity (H-index). Main species of *Chaetoceros* RS group in green (a), sea ice group in blue (b-e), cold-water group in purple (f-g), cosmopolitan group in red (h) and coastal group in yellow (i).

Figure 4. Time series of environment data and fluxes measured at Station DM. a) Arctic Ocean Oscillation index (Proshutinsky et al., 2015) which indicates the strength of Beaufort Gyre; b) Transport in Sv of the Pacific water inflow (Woodgate et al., 2015; Woodgate 2018); c) Sea ice concentration in % (Saha et al., 2010, 2014); d) Total diatom flux in $10^6 valves m^{-2} d^{-1}$; e) Total mass flux in $mg m^{-2} d^{-1}$ (Bai et al., 2019).

Figure 5. Representation of the surface currents and sea ice conditions in the study area in July and August of 2008 and 2009. PWI: Pacific water inflow, BG: Beaufort Gyre. Blocked currents by sea ice are indicated by the dashed lines. Sea ice data were retrieved from Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data with a grid resolution of $25 \times 25 km$ (Cavalieri et al., 1996).

Figure 6. a) Sea ice concentration (in %) and terrigenous related fluxes: b) Coastal diatom flux in $10^5 valves m^{-2} d^{-1}$; c) *Chaetoceros* resting spore flux in $10^5 valves m^{-2} d^{-1}$; d) Total flux of campesterol and β -sitosterol in $\mu g m^{-2} d^{-1}$; e) Lithogenic matter flux in $mg m^{-2} d^{-1}$ (grey bars) and lithogenic matter in % weight (red line). g) Bulk organic C/N ratio is also shown (blue line; Bai et al., 2019) and compared to the Redfield ratio as a reference (grey dash line).

Figure 7. a) Fluxes of sea ice diatom in 10^5 valves $\text{m}^{-2} \text{d}^{-1}$ and IP_{25} in $\text{ng m}^{-2} \text{d}^{-1}$ (relative to the internal standard). The occurrence of sea ice diatoms that produces IP_{25} are indicated by squares. b) Correlation plot between sea ice diatom flux and IP_{25} flux ($r^2 = 0.64$; $p < 0.01$).

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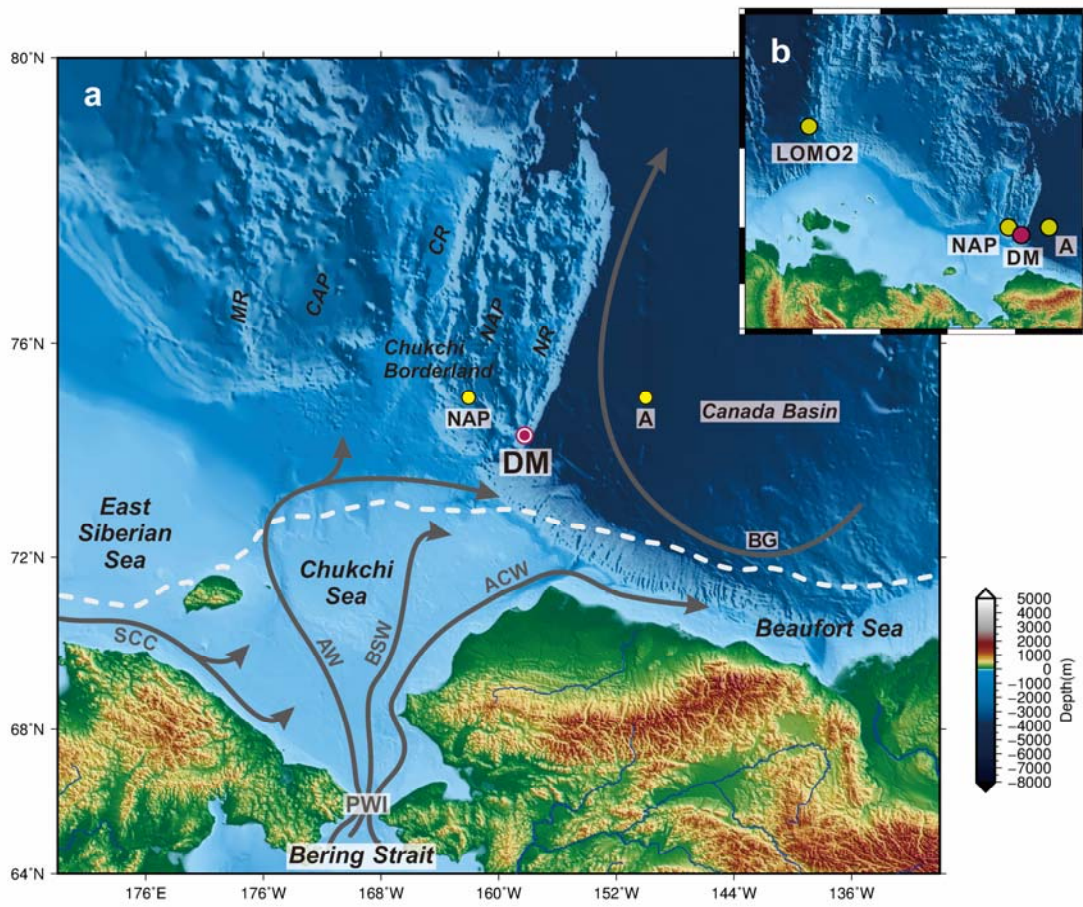
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9 Figures

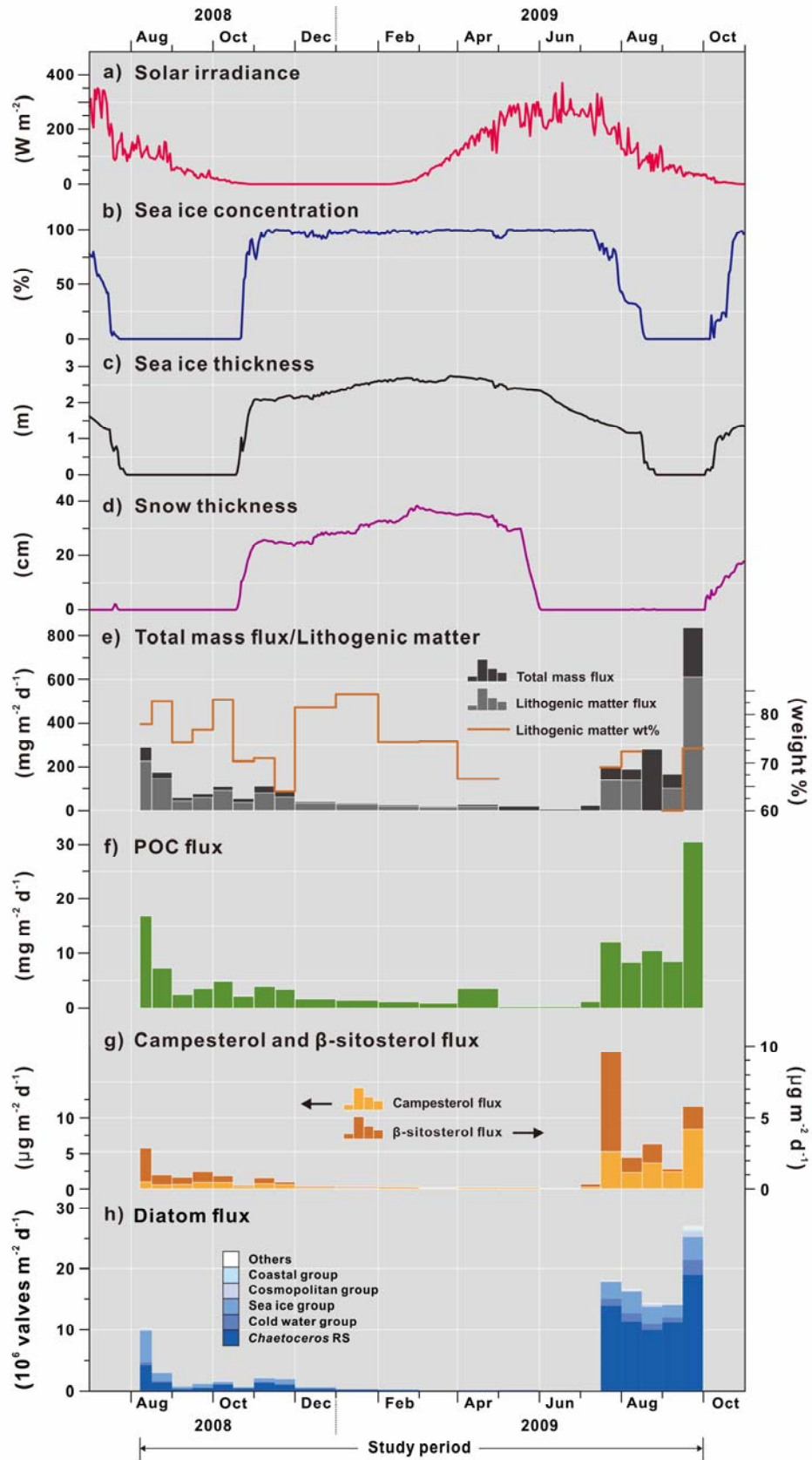
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11 Figure 1-7

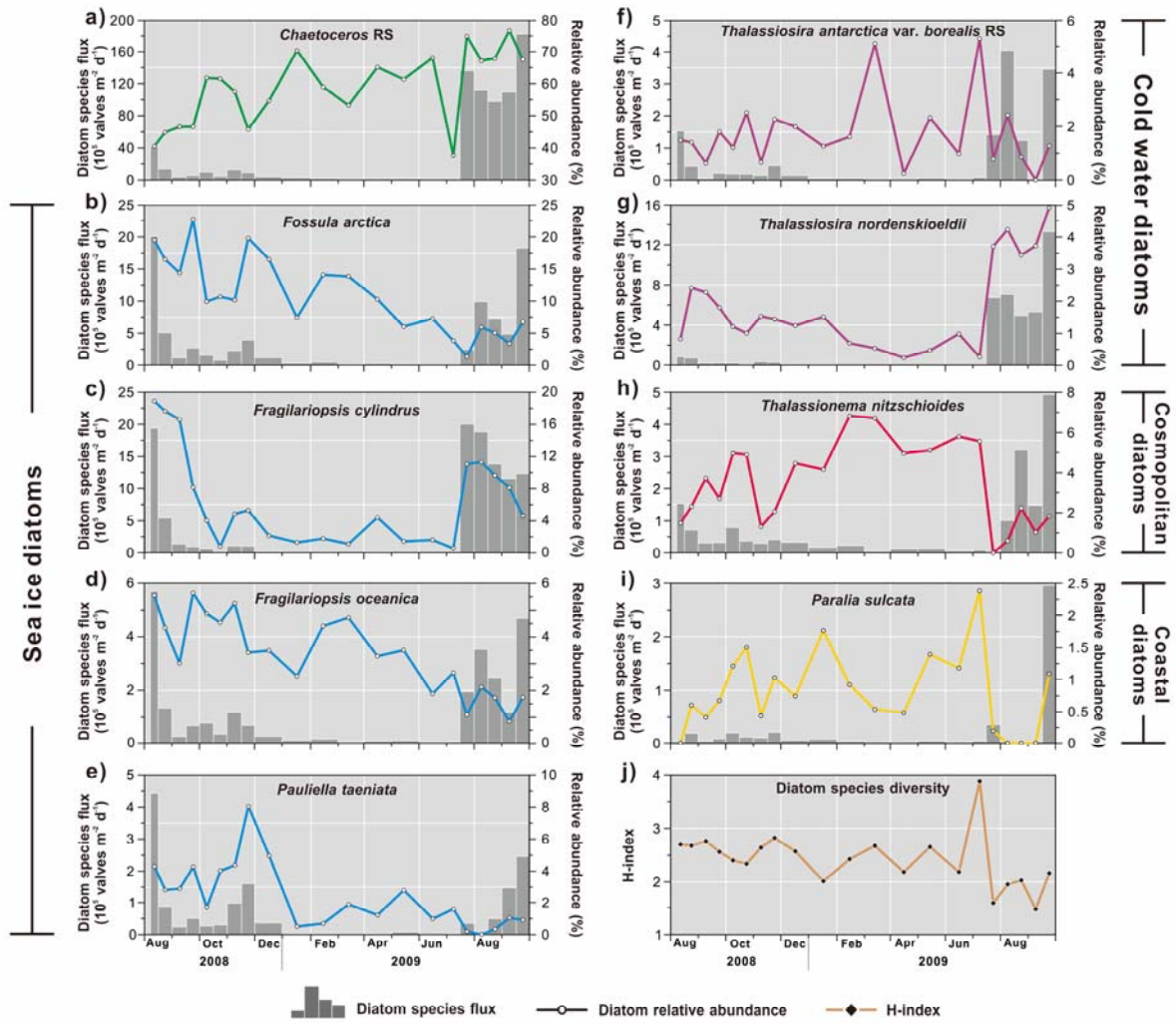
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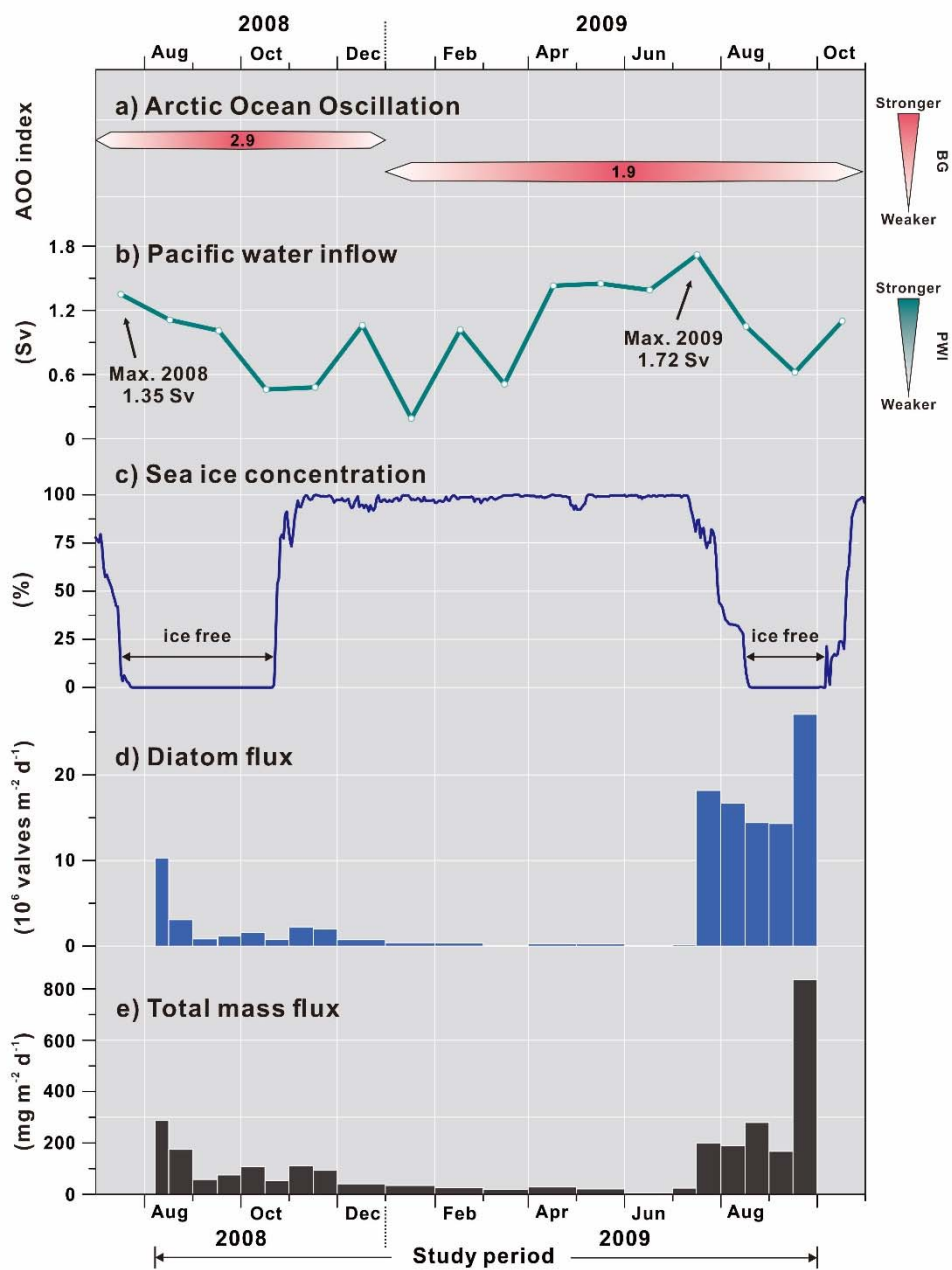
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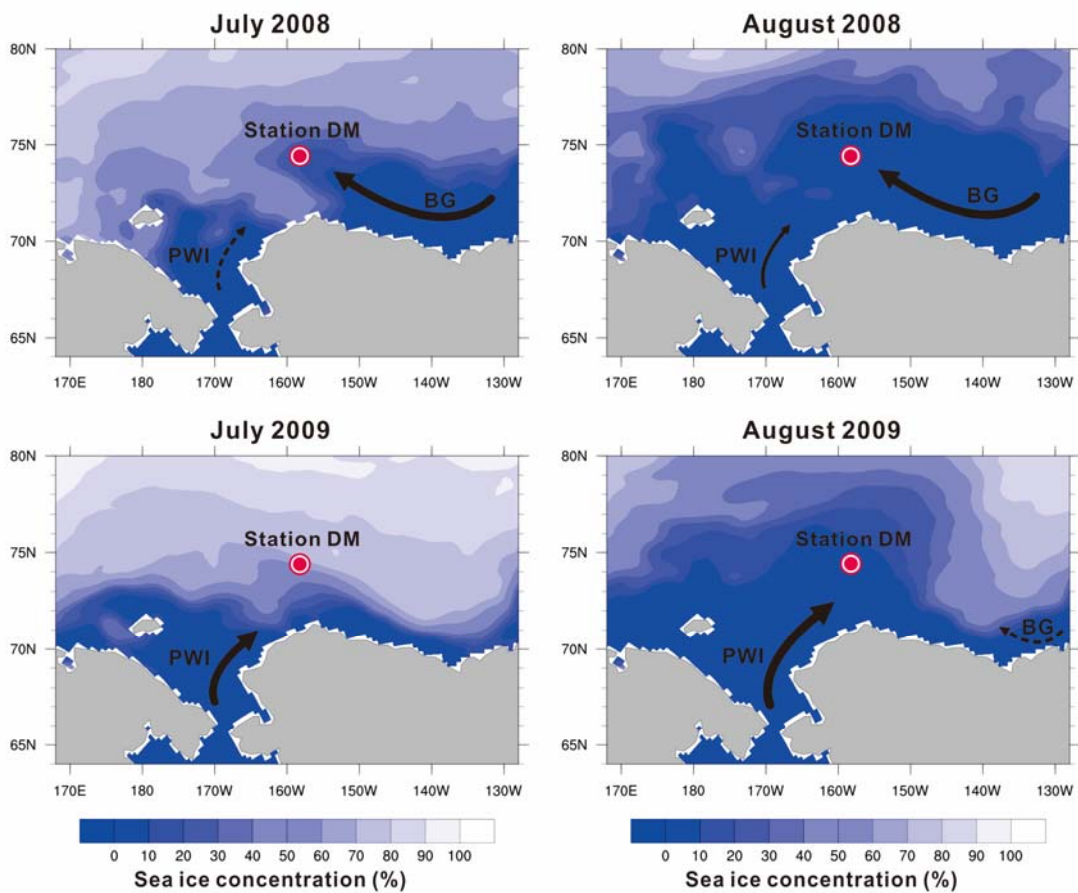
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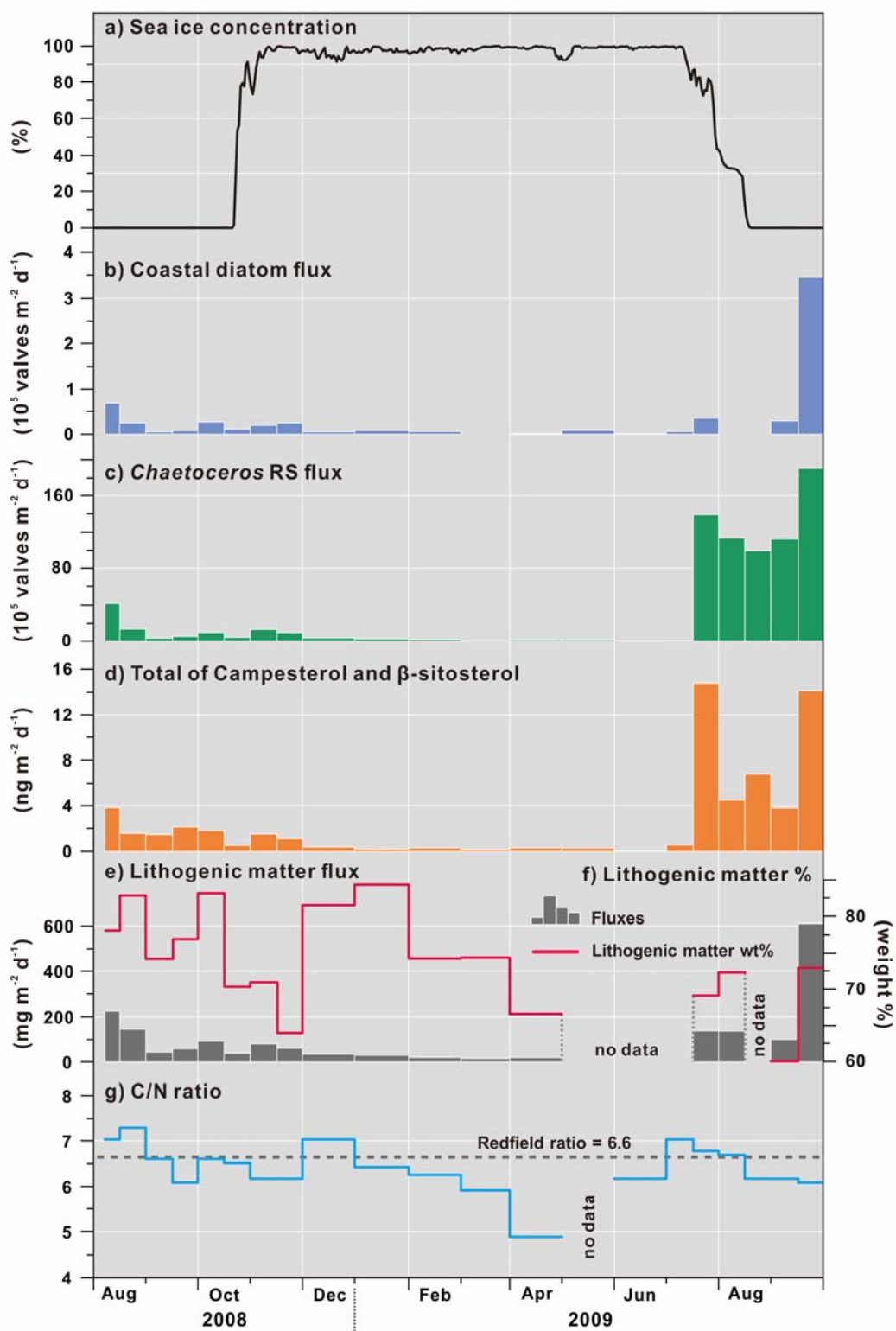
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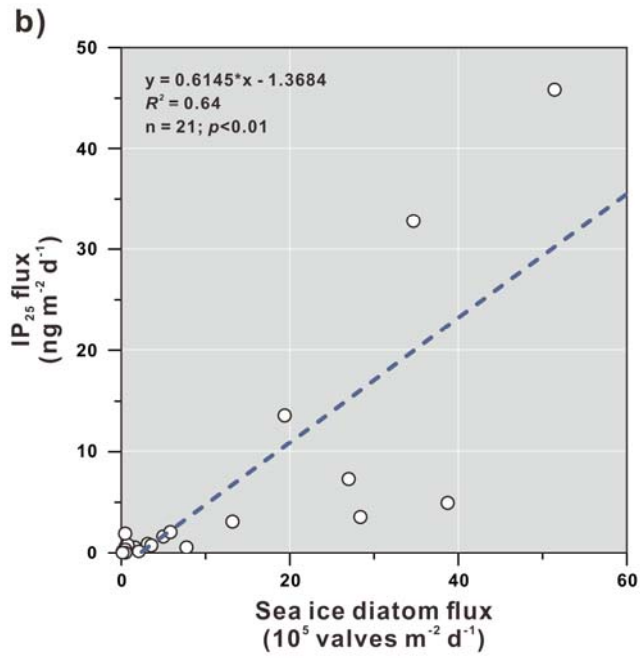
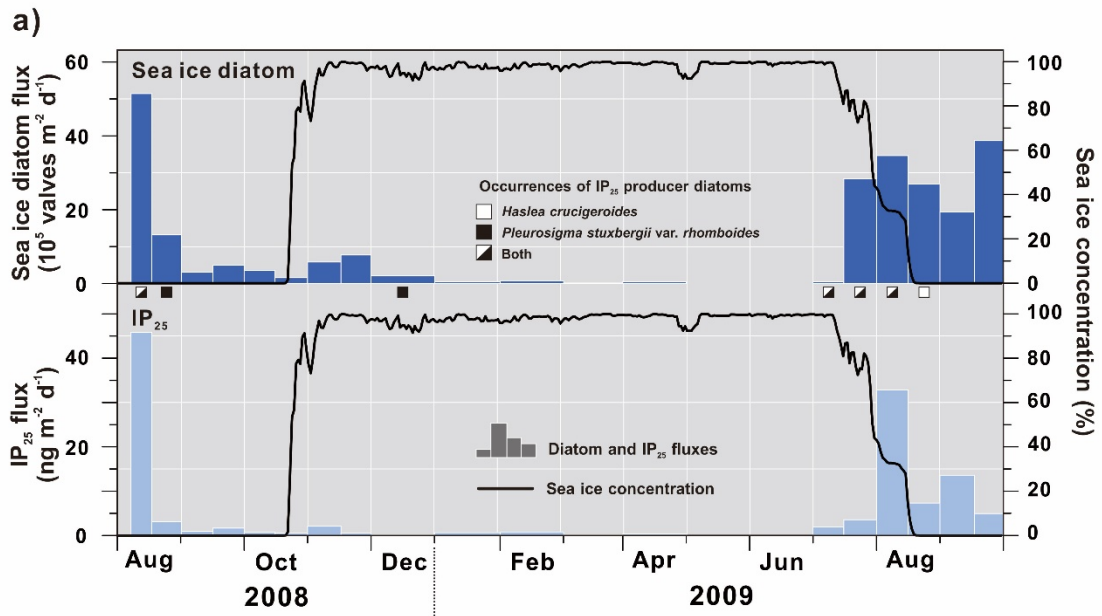
16 **Figure 5**



17 **Figure 6**



18 Figure 7



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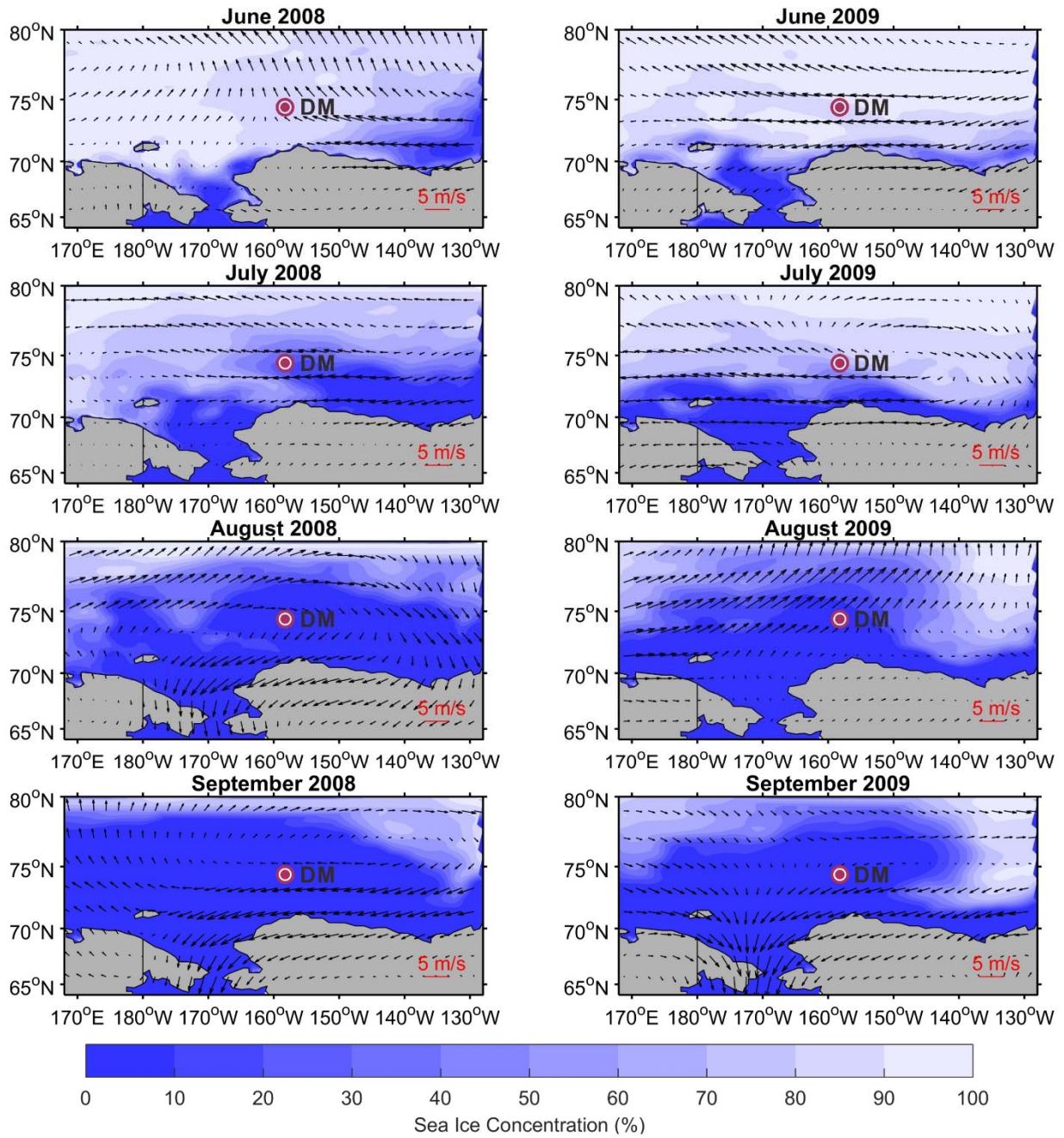
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9 Supplementary Figure

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11 Figure S1

12 **Figure S1**



13 **Figure S1.** Wind field of the study area in June, July, August and September of 2008 and 2009.

14 The wind data were retrieved from the NCEP/ NCAR atmospheric reanalysis data of daily 10
15 m-height wind speed (Kistler et al., 2001), with a horizontal resolution of 1.875°. Station DM
16 and sea ice concentration are also shown.

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9 Table 1-2

10 **Table 1**

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12 **Table 1. Comparison of the diatom diversity and relative abundances from the sediment trap sinking particles at Station DM and nearby surface**
 13 **sediments**

	Nr. of Species	H-index	<i>Chaetoceros</i> RS (%)	<i>Fossula arctica</i> (%)	<i>Fragilariopsis cylindrus</i> and <i>F. oceanica</i> (%)	<i>Thalassiosira nordenskiöldii</i> (%)	<i>Thalassiosira antarctica</i> var. <i>borealis</i> RS (%)
Station DM ^a	93	2.30	64.3 ± 11.3	8.2 ± 6.0	11.5 ± 6.0	2.9 ± 1.4	1.3 ± 1.4
Surface sediments ^b	78	2.46	63.3 ± 8.3	5.5 ± 2.9	11.3 ± 7.6	1.9 ± 1.1	3.3 ± 2.1

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^a Annual values based on merged seasonal samples

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^b Based on 10 surface sediments from northern Chukchi Sea (Ran et al., 2013)

16 **Table 2**

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 18 **Table 2.** Coefficient of determination (r^2) between fluxes and relative abundances (with and without *Chaetoceros* RS, respectively) of sea ice diatoms and cold water diatom
 19 group, fluxes of brassicasterol and dinosterol and fluxes of IP₂₅ and HBI-III from August 2008 to September 2009 at Station DM. Significant correlations are shown in bold and
 20 underlined, while weak correlations are in bold and italic.

	Sea ice diatom flux ^a	Cold-water diatom flux ^a	Sea ice diatom %	Cold-water diatom %	Sea ice diatom % (without <i>Chaetoceros</i> RS)	Cold-water diatom % (without <i>Chaetoceros</i> RS)	Brassicasterol flux ^b	Dinosterol flux ^b	IP ₂₅ Flux ^c	HBI-III flux ^c
Sea ice diatom flux ^a	1	-	-	-	-	-	-	-	-	-
Cold-water diatom flux ^a	<u>0.64</u> *	1	-	-	-	-	-	-	-	-
Sea ice diatom %	0.02	0.09	1	-	-	-	-	-	-	-
Cold-water diatom %	0.02	0.09	0.002	1	-	-	-	-	-	-
Sea ice diatom % (without <i>Chaetoceros</i> RS)	0.11	0.01	0.65 *	0.05	1	-	-	-	-	-
Cold-water diatom % (without <i>Chaetoceros</i> RS)	0.14	0.54 ^e	0.43*	0.281**	0.21**	1	-	-	-	-
Brassicasterol flux ^b	0.44*	0.83 *	0.04	0.09	0.01	0.35*	1	-	-	-
Dinosterol flux ^b	0.56 *	0.77 *	0.06	0.05	0.002	0.45*	0.67 *	1	-	-
IP ₂₅ flux ^c	0.64 *	0.16	0.09	<0.0001	0.17	0.002	0.03	0.08	1	-
HBI-III flux ^c	0.48 *	0.03	0.24**	0.02	0.21**	0.06	0.01	0.02	0.75 *	1

21 ^a in 10⁵ valves m⁻² d⁻¹

22 ^b in µg m⁻² d⁻¹

23 ^c in ng m⁻² d⁻¹

24 * p -value < 0.01

25 ** p -value < 0.05