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Molines

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A role for the ocean in the winter sea ice distribution north of Svalbard.

Christophe Herbaut<sup>1</sup>, Marie-Noëlle Houssais<sup>1</sup>, Anne-Cécile Blaizot<sup>1</sup> and Jean-Marc Molines<sup>2</sup>

1 : LOCEAN, Sorbonne Université, CNRS, IRD, MNHN

2 : Institut des Géosciences de l'Environnement, Université Grenoble-Alpes, CNRS

22 **Key Points:**

23 Identification of large ice melt rates in pack ice along the pathway of the Atlantic Water.

24 Two main types of large melt events: a) ice advection over warm SST and b) enhanced vertical  
25 heat fluxes due to wind driven mixing.

26 Boundary current strengthening accompanied by shoaling of the Atlantic Water layer is a  
27 secondary factor leading to large melt rates.

28

29 **Abstract**

30 The ocean is suggested to play a major role in the ongoing winter decay of the sea ice cover in  
31 the western Eurasian Basin. Using a high-resolution sea ice-ocean model, we investigate the  
32 processes influencing the ice-ocean interactions in winter in the waters north of Svalbard, with  
33 a particular focus on those contributing to sea ice melt events of large amplitude. These short  
34 term events, lasting 5-10 days, are associated with locally large melt rates mostly found along  
35 the pathway of the Atlantic Water. The sum of all these events over the simulation period is  
36 found to contribute 40% of the total winter melt. Episodes of strong surface winds, occasionally  
37 associated with enhanced velocity shear at the mixed layer base can trigger enhanced  
38 entrainment of Atlantic Water through the relatively shallow upper thermocline in the Atlantic  
39 Water boundary current, leading to substantial ocean heat transfer to the sea ice. In some cases,  
40 strengthening of the boundary current also contributes to fueling the heat transfer to the ice.  
41 Another type of large melt events, not linked to increased ocean vertical heat flux but due to ice  
42 being advected over warm surface waters is also identified, sometimes associated with episodes  
43 of ice close-up. Sea ice budget calculations show that, overall, large melt events contribute  
44 significantly to the eastward retreat of the winter marginal ice zone on the upper slope east of  
45 Svalbard while episodes of northward advection of ice largely dominate the ice edge retreat  
46 over the shelf north of Svalbard.

47 **Plain text summary.**

48 Model simulations show that large sea-ice melt events can be found in winter within the Arctic  
49 pack ice. These events are generated by a transfer of heat from the underlying ocean to the sea  
50 ice. The melt is mainly confined above the Atlantic Water boundary current which is known to  
51 carry warm water from the Atlantic Ocean to the Arctic. During periods of strong winds, part of  
52 the heat stored in the subsurface Atlantic Water reservoir can be transferred to the surface layer  
53 to help melt the ice. On some occasions, a strengthened current contributes to bringing warm  
54 water closer to the ocean surface, reinforcing the wind effect. Large melt rates are also found  
55 when ice is being moved over warm surface waters.

56

57

## 58 **1. Introduction**

59 The negative trend in the Arctic sea ice extent is characterized by a regional dependency and  
60 acceleration in the recent years (Close et al., 2015; Onarheim et al., 2018). In the Atlantic sector  
61 of the Arctic, the negative trend is larger in winter and is marked by a northeast retreat of the  
62 sea ice edge north of Svalbard (Onarheim et al., 2014). This decadal trend has been shown to  
63 be linked to oceanic influence and the Atlantic Water (AW) heat transport (e.g., Årthun et al.,  
64 2012; Ricker et al., 2021), which can lead to significant skill in decadal Arctic sea ice prediction  
65 in the Atlantic sector (Yeager et al., 2015, Årthun et al., 2017). The regions of skillful prediction  
66 could extend farther east into the Nansen Basin in the near future (Årthun et al., 2019).

67 The AW flows into the Arctic via the Barents Sea and Fram Strait branches. The former one  
68 loses a large part of the transported heat before entering the Arctic through St Anna Trough  
69 (Skagseth et al., 2008). The second branch is indeed the main heat source for the Arctic Ocean.  
70 The AW is carried from Fram Strait into the Arctic Ocean via different pathways which have  
71 their common source in the West Spitsbergen Current (Aagaard et al., 1987; Cokelet et al.,  
72 2008; Crews et al., 2019; Meyer et al., 2017; Koenig et al., 2017; Menze et al., 2019) and then  
73 merge east of Svalbard to form a 30-50 km wide eastward boundary current along the southern  
74 slope of the Nansen Basin (Våge et al., 2016; Pérez-Hernández et al., 2017).

75 Several studies have indeed pointed out the role of the AW in the decline of the winter sea ice  
76 cover in the Eurasian Basin during the recent decades (Polyakov et al., 2010; 2017; 2020b).  
77 The growing role of the AW in this decline is attributed to AW warming and weakened  
78 stratification in the upper halocline known as the “Atlantification” of the Arctic Ocean. First  
79 detected in the Barents Sea and western Nansen Basin, as patterns of enhanced sea ice thickness  
80 decrease along the AW pathway (Ivanov et al., 2012), the impact of the ocean is now detected  
81 in the eastern Eurasian Basin where weakening of the stratification in early winter would

82 precondition the region to enhanced winter ventilation and substantial heat flux from the AW  
83 layer to the sea ice (Ivanov et al., 2016; Polyakov et al., 2020b). The resulting decrease in winter  
84 sea ice growth would have contributed to the recent negative trend in the winter sea ice  
85 thickness in this region in similar proportions to the thermodynamic atmospheric forcing  
86 (Polyakov et al., 2017). Both thermal and haline convection would play a major role in this  
87 mechanism, but the increase of the velocity shear, coupled with a weakening of the stratification  
88 could also potentially increase the vertical heat fluxes through enhanced vertical mixing  
89 (Polyakov et al., 2020c).

90 The region north of the Svalbard archipelago is the place where the AW encounters the sea ice  
91 and contributes to significant ice melt through the release of large amounts of heat to the surface.  
92 On average, in the region extending immediately north and east of Svalbard (hereafter referred  
93 to as the North Svalbard region) the sea ice budget is dictated by a balance between the  
94 southward ice transport into the region and bottom melt fed by the convergence of warm AW  
95 (e.g. Duarte et al., 2020). The AW layer is found just below the surface mixed layer, and large  
96 vertical heat fluxes in the range 240-300 W/m<sup>2</sup> have been measured in the upper thermocline  
97 on the shelf north of Svalbard in winter and, with same order of magnitude, in the under-ice  
98 boundary layer (Sirevaag and Fer, 2009). These values contrast with the central Arctic Ocean  
99 where weak vertical mixing at intermediate depth is responsible for small vertical heat fluxes,  
100 on order of 1 W/m<sup>2</sup> or less at the top of the AW layer (Guthrie et al., 2013; Carmack et al.,  
101 2015).

102 The winter sea ice extent in the North Svalbard region is subject to large interannual variability  
103 being the result of both ocean and atmosphere variability. While, on the long-term, these  
104 variations correlate significantly with the AW temperature anomalies (Onarheim et al., 2014),  
105 ice advection was found to explain most anomalous winter sea ice seasons of the period 2012-  
106 2019 (Lundesgaard et al., 2021). On shorter time scales, occurrences of large ice-free areas in

107 mid-winter are suspected to be associated with enhanced ocean influence along the AW  
108 pathway over the recent years (Ivanov et al., 2016). AW influence is effective despite the  
109 existence of a strong halocline which normally isolates the AW from the cold, fresh Polar Water  
110 filling the upper 20-30 meters of the water column as a result of annual sea ice melt (Rudels et  
111 al., 2004). Indications of deep reaching winter mixing were indeed reported for winter 2013  
112 over the Eurasian slope at 30°E, when winter convection reached into the AW layer down to  
113 400-500 m (Pérez-Hernández et al., 2019). In winter 2015, during the N-ICE2015 drift  
114 experiment in the waters north of Svalbard, entrainment of AW was observed to be enhanced  
115 during storm events, leading to ocean-ice heat flux of several tens of  $\text{W/m}^2$  on average when,  
116 at the same time, the AW layer depth was particularly shallow (Meyer et al., 2017). Composite  
117 estimates of the sea ice melt rates focusing on the major storms in the same period would even  
118 lead to higher values on order of 200-300  $\text{W m}^{-2}$  (once converted into ice-ocean heat flux),  
119 peaking at up to 600  $\text{W m}^{-2}$  over the AW pathway (Graham et al., 2019).

120 The above results were obtained from a limited number of events, and for a particular winter  
121 (2015) with low sea ice conditions and would deserve a robust assessment over a longer time  
122 span using a consistent evaluation of the ice-ocean interactions. Such an approach was used  
123 recently by Duarte et al. (2020) with a one-dimensional sea ice model coupled to a slab ocean.  
124 While their model results show consistent order of magnitude between observed sea ice melt  
125 rates and the model predictions in a number of case studies, the idealized ocean model physics,  
126 especially water column restoring to observations to account for the missing physics (e.g. ocean  
127 advection), call for a more detailed representation of the ocean in order to assess the relative  
128 importance of the various processes at play. Moreover, sea ice melt rate estimates rely on sea  
129 ice budgets which are subject to large uncertainties due in particular to the low spatial resolution  
130 of the sea ice drift satellite products and the largely unknown sea ice thickness.

131 An interesting question is to evaluate the separate contributions of forcing and ocean  
132 preconditioning to the variability of the ocean-ice heat fluxes. In all the cases mentioned above,  
133 vertical mixing appears to be largely influenced by the three-dimensional distribution of the  
134 AW and halocline layer, in particular, the actual depth of the upper AW layer interface. The  
135 latter controls the reachability of the subsurface heat reservoir for vertical mixing. In the eastern  
136 Eurasian Basin, recent shoaling of the AW layer was indeed found to contribute to enhancing  
137 winter convection (Polyakov et al., 2020b). According to the simplified dynamic framework of  
138 an idealized slope current, the equilibrium depth of the AW layer is set by a balance between  
139 lateral mean and eddy transport and vertical diffusion, with the halocline rising toward the basin  
140 periphery, over the core of the Atlantic boundary current (Spall, 2013). However, the  
141 mechanisms underlying the variability of the halocline depth are still debated. Over the upper  
142 slope at 30°E, the upper part of the halocline was observed to rise to depths as shallow as 25 m  
143 (Pérez-Hernández et al., 2019; Renner et al., 2018). Shelf break upwelling (e.g. Lind and  
144 Ingvaldsen, 2012) may also help occasionally bring the AW layer closer to the surface, a  
145 mechanism which has been already documented by Pickart et al., (2009) off northern Alaska,  
146 and mentioned by Falk-Petersen et al. (2015) to possibly explain the presence in winter of AW  
147 in surface north of Svalbard. Similarly, Ekman pumping in response to cyclonic surface wind  
148 stress curl can lead to significant (order of 20 m over one month) shoaling of the AW layer on  
149 the northern Barents Sea slope (Renner et al., 2018). Despite these valuable observations, the  
150 variability of the AW layer geometry and how this influences the distribution of the sea ice melt  
151 in the North Svalbard region is still largely unknown.

152 To identify the dominant factors controlling the upward heat transfer to the under-ice ocean  
153 layer and ultimately the sea ice in the North Svalbard region, in this study we investigate the  
154 large-scale ocean conditions that could lead to enhanced influence of the AW on the winter sea  
155 ice and evaluate the contribution of the associated sea ice melt to the winter sea ice distribution.



156 By focusing on the intra-seasonal variability of winter sea ice melt, we aim to identify the  
157 preferred areas and the conditions of occurrence of the most significant short-term melt events  
158 and to assess their contribution to the mean winter sea ice budget. The analysis is primarily  
159 based on results of a sea ice-ocean simulation covering the period 1997-2017, therefore  
160 providing a temporal context to earlier studies based on shorter-term observations.

## 161 **2. Data and model set-up**

162 The outputs of a simulation with a regional coupled sea ice-ocean model are used to analyze  
163 the relationship between the sea ice variability and the AW properties and circulation. The  
164 model is based on NEMO (Nucleus for European Modelling of the Ocean) version 3.6 (Madec,  
165 2008) coupled to the LIM3 (Rousset et al., 2015) sea ice model. The equations are discretized  
166 on 75 vertical levels with thickness varying from 1 m in the top layer to roughly 250 m at the  
167 deepest model level. Partial steps are used to better represent the bottom topography. A  
168 turbulent kinetic energy closure scheme that represents the evolution of the turbulent kinetic  
169 energy is used for vertical mixing of momentum and tracers (Blanke and Delecluse, 1993).  
170 The model grid has a horizontal resolution varying from  $\sim 2.5$  km in the Arctic Ocean to 4 km  
171 in the subtropics. The domain encompasses the Arctic Ocean and the North Atlantic Ocean,  
172 with open boundaries at  $28^{\circ}\text{N}$  in the Atlantic and at Bering Strait, along which the velocity and  
173 tracer distributions are prescribed from the 5-day outputs of a global  $\frac{1}{4}^{\circ}$  resolution hindcast  
174 simulation which has been run by the Drakkar group (<https://www.drakkar-ocean.eu/>) over the  
175 same period and using the DFS5 surface forcing (Dussin et al., 2018).

176 The model is initialized from rest with initial temperature and salinity distributions from the  
177 PHC 3.0 global ocean climatology, updated from Steele et al. (2001). The model surface forcing  
178 is based on 6-hour surface atmospheric fields from the ERA-I reanalysis (Dee et al., 2011).  
179 Turbulent heat fluxes are computed using the Large and Yeager (2004) bulk formulae with

180 daily dew point temperature, surface air temperature, wind speed and sea level pressure (SLP)  
181 as the main atmospheric inputs. Other forcing fields from the reanalysis include the 10-meter  
182 wind and the downward radiative fluxes. Surface albedos and temperatures simulated by the  
183 model are used to calculate the upward components of the radiative fluxes. Regional corrections  
184 have been made to improve the radiation fluxes over the North Atlantic and the surface air  
185 temperature over the Arctic. All corrections have been applied to the annual climatological  
186 cycle (obtained by averaging the daily fields between the 34 years of the reanalysis), while the  
187 field anomalies (obtained by subtracting this climatology from the original daily fields) are kept  
188 unchanged. Over the Atlantic Ocean, the ISCCP (International Satellite Cloud Climatology  
189 Project) radiation climatology (Rossow and Schiffer, 1999) is used instead of ERA-I while in  
190 the Arctic Ocean (north of Fram Strait) the ERA-I air temperature climatology is replaced by  
191 the IABP (International Arctic Buoy Programme) climatology (Rigor et al., 2000) in order to  
192 remove unrealistic positive summer values. Precipitation is extracted from the DFS5 forcing  
193 fields (Dussin et al., 2018), which are derived from ERA-I. Continental runoff is prescribed  
194 from the monthly climatology of Dai and Trenberth (2002). The simulation is performed over  
195 the period 1995-2017, and the analysis is based on the period 1997-2017. Restoring of the  
196 surface salinity to the PHC climatology is applied with a time scale of 30 days in the first 9  
197 years but not applied for the remaining 14 years of the simulation.

198 To assess the simulated sea ice variability against observations we use daily sea ice  
199 concentrations (SIC), which are estimated from satellite borne passive microwave radiometer  
200 (the Special Sensor Microwave / Imager (SSM/I) and Special Sensor Microwave / Imager  
201 Sounder (SSM/IS) ) observations by applying the ARTIST Sea Ice (ASI) algorithm (Kaleschke  
202 et al., 2001) and are provided on a 12.5x12.5 km grid.

203 The present analysis is based on 5 day-average model outputs. Winter is defined as the period  
204 extending from December to March, November and April being excluded as transition months

205 between seasons (the year attributed to each winter corresponds to that of January). The  
206 variability is characterized based on 5-day averaged anomalies which are constructed by  
207 removing separately for each year, the average of the winter period. The reconstructed  
208 anomalies thus characterize the intra seasonal variability over the winter period (hereafter  
209 referred to as the *winter variability*) without incorporating the interannual and longer-term  
210 variability.

211

### 212 **3. Results**

#### 213 3.1 Simulated mean AW pathways and transports

214 Figure 1 shows the simulated mean depth of the 0°C isotherm (which can be viewed as the  
215 upper limit of the AW layer) and velocity at 60 m, in the upper part of the AW core, averaged  
216 over the period 1997-2017. North of Fram Strait, the current splits into three branches, in  
217 agreement with earlier descriptions of the circulation in this region (e.g. Aagaard et al., 1987;  
218 Koenig et al., 2017; Crews et al., 2019): the Svalbard Branch which stays south of the Yermak  
219 Plateau on the upper slope north of Svalbard, the Yermak Pass Branch which flows over the  
220 plateau and the Outer Yermak Branch which tends to follow the slope around the plateau.  
221 Northeast of Svalbard, the different branches merge into a boundary current which is localized  
222 between the 250 and 1000 m isobaths. The width of the current (30-50 km) and the seasonal  
223 evolution of the AW T S properties, with shallower, warmer core in the fall-winter season (Fig.  
224 S1 and S2) are consistent with earlier observations, as well as the spatial distribution of the  
225 properties in late summer (Fig. S3) (Våge et al., 2016; Pérez-Hernández et al., 2017). Yet, the  
226 onset of the AW warming occurs later in the model as compared with the observations, leading  
227 to overestimated AW temperature in winter. When computed with the same criteria as those  
228 used by Perez-Hernandez et al., 2019, the model AW transport averaged over the period 2012-  
229 2013 agrees well with observations over the same period (Perez-Hernandez et al., 2019) with

230 an annual mean of 2 Sv and a similar seasonal evolution (Fig. S4). A model-data comparison  
231 performed upstream in Fram Strait shows reasonable agreement between the model and the  
232 observations also in the West Spitsbergen Current, yet with a current velocity core slightly  
233 shifted offshore in the model (Fig. S5-S6). More details on the model skill in representing the  
234 Atlantic Water properties and transports can be found in the Supporting Information.

### 235 3.2 Winter variability of the sea ice concentration

236 The mean winter distributions of sea ice concentration and thickness simulated by the model  
237 are in reasonable agreement with available observations, although systematic bias toward a  
238 more compact, thinner ice cover can be identified in our region of interest (see supporting  
239 information, Fig. S7 for further details). To determine the spatial and temporal structure of the  
240 *winter variability* of the ice cover, an empirical orthogonal function (EOF) analysis has been  
241 performed on the SIC observations over the period 1997-2017 and the domain 79°N-84°N, 0°E-  
242 50°E (Fig. 2). The EOF analysis shows occurrences of short-term (duration of 5-20 days) low  
243 sea ice concentration events extending all along the northern Svalbard and Barents Sea slope  
244 area (Fig. 2b), including large amplitude events in winter 2000, 2006, 2013 and 2017 (Fig. 2c).

245 A similar analysis has been performed on the simulated SIC over the same period (Fig. 2a,c).  
246 Similar major low concentration events are found in the simulation, and the correlation between  
247 the first principal component of the observed and simulated SIC is 0.74, suggesting that the  
248 model reproduces rather well the *winter variability* of the SIC (Fig. 2c). However, the loading  
249 pattern of the simulated SIC does not extend as far northward and eastward. The origin of the  
250 discrepancy is a persistent overestimation of the winter sea ice extent in this region in the  
251 simulation, with the marginal ice zone (MIZ, defined as the region where the sea ice  
252 concentration is between 15 and 80% and is expected to be the most variable), being located  
253 more persistently over the shelf than over the slope region like in the observations. The first

254 EOF of the simulated sea ice thickness (not shown) displays a similar loading pattern as the  
255 SIC, and the correlation of 0.75 between the respective principal components suggests similar  
256 time evolution for the two properties.

### 257 3.3 Contribution of the melt to the sea ice volume budget.

258 As shown in Figure 3a, which displays the mean winter sea ice volume change over our study  
259 area (see the geographical extent of the study area in Fig, 2a), sea ice on average tends to grow  
260 throughout the winter season over the entire area (here and in the following the ice volume  
261 stands for ice volume per unit area). While no particular pattern of this winter growth is found  
262 over the upper slope area (the region encompassed between the 250 m and 1000 m isobaths),  
263 both its dynamic (transport convergence) and thermodynamic contributions show a pattern of  
264 large convergence and melt aligning with these isobaths (Fig. 3c,e). Enhanced convergence  
265 there coincides with weaker thermodynamic growth or even net melt, which extends eastward  
266 into the inner pack (the pattern is almost identical if one selects only melt occurring in locations  
267 where the SIC is larger than 80%). The signature of enhanced melt, concentrated over the AW  
268 current path (compare Fig. 1 with Fig. 3e), is consistent with a response to higher ocean heat  
269 flux in this area (Fig. 3b). The mean winter melt is in part due to short term (5-10 days) large  
270 melt events (Fig. 2d, see next section for an explanation of the index), which can lead to major  
271 ice volume decrease as observed, for instance, in January or in February 2006 (Fig 4a, green  
272 line). The standard deviation of the thermodynamic contribution (Fig. 3f) to the mean winter  
273 ice volume change averaged over the study area remains, however, 3 times smaller than that of  
274 the dynamic contribution (Fig. 3d) over the full period of the simulation. Actually, large ice  
275 volume changes driven by variations of the ice transport convergence (Fig. 3d), although  
276 affecting primarily the MIZ over the shelf north of Svalbard, also extend into the pack ice  
277 interior, immediately east of Svalbard. By contrast, the thermodynamic contribution affects the  
278 ice volume over the AW current pathway (Fig. 3f).

### 279 3.4 Large melt events.

280 To characterize the episodes of large melt and identify their likely causes, a time series of the  
281 area-weighted ice melt rate has been built by averaging only negative values of the net growth  
282 rate (or equivalently setting growth rate to zero where positive) over the study area. Doing so  
283 we aim to better represent the episodes of large melt, without incorporating the compensating  
284 effects of areas of positive net growth (note that changes in the melt rate amplitude were found  
285 to dominate over changes in the area of melt in the index variability). We then define a melt  
286 index by removing, separately for each year, the mean of the winter period to construct winter  
287 melt anomalies (Fig. 2d). Large melt events have been defined as those for which the index is  
288 larger than 1 standard deviation (similarly occurrences of low melt are defined as those for  
289 which the index is smaller than 1 standard deviation). Over the 1998-2017 winter period 46 5-  
290 day episodes of large melt are found, of which 12 contribute to pairs of consecutive episodes,  
291 leading to a total of 40 large melt events with duration ranging between 5 and 10 days. The  
292 amplitude and number of large melt events per winter are quite variable (the number ranges  
293 from a maximum of 5 events in 2001 to none in 2014) but, in total, they account for about 40%  
294 of the cumulated winter melt over the whole period. In the next section, we analyze three large  
295 melt events which occurred in 2006 and were associated with different driving processes. Then,  
296 we will evaluate the relevance of the conclusions drawn from these events to the other large  
297 melt events.

#### 298 3.4.1 Case study: large melt events in 2006.

299 During winter 2006, the ice cover starts to decrease on January 10 in response to large  
300 divergence of the ice volume transport (Fig. 4a). The decrease accelerates on January 15, due  
301 in particular to a large melt event, before the SIC reaches a minimum on January 25. From  
302 January 30 the ice cover starts to reconstruct due to ice being transported into the study area,

303 but the ice close-up is slowed down due to a second large melt event between February 24 and  
304 March 1 which tends to oppose a concomitant large convergence event.

305 During the first large melt event, the melt increases concomitantly to the occurrence of strong  
306 southeasterly wind (Fig. 4b and 5a), and melt rates larger than 6 cm/d can be observed within  
307 the pack ice over the path of the boundary current (and also in the Kvitoya Trough) (Fig. 5b).  
308 The melt pattern coincides with large values of ice-ocean heat flux which can reach more than  
309  $400 \text{ W/m}^2$  (Fig. 5c). A vertical section across the shelf-slope area at  $31^\circ\text{E}$  shows that the largest  
310 melt rates are found just above the region where the AW layer is closer to the surface (the depth  
311 of the  $0^\circ\text{C}$  isotherm is about 15-20 m) and the largest vertical diffusive heat flux and the  
312 warmest temperature can be observed (Fig. 6a,b). The vertical diffusive flux is responsible for  
313 the heat transfer from the upper part of the pycnocline to the surface layer. Between  $81^\circ\text{N}$  and  
314  $81.4^\circ\text{N}$ , the maximum of this flux is in the subsurface and the flux decreases towards the surface  
315 in the upper 10 meters suggesting a convergence of heat and a warming of this layer. Between  
316  $81.45^\circ\text{N}$  and  $81.6^\circ\text{N}$  the vertical gradient of the diffusive heat flux is of opposite sign,  
317 suggesting an additional heat source near the surface. However, when forming a heat budget  
318 for the upper 10 meters of the water column over the area where the melt rate is larger than 2  
319 cm/day during the large melt event (hereafter referred to as the *melt area*), the vertical diffusive  
320 heat flux appears as the major heat source for this layer on January 15 (Fig. S8), and all this  
321 heat is lost to the ocean surface with very minor contributions from both vertical and horizontal  
322 advection. On average over the *melt area*, horizontal advection is indeed negligible, which  
323 suggests that this process mainly contributes to horizontal redistribution of the heat within the  
324 area (being responsible, for instance, for local input to the surface layer as identified on the  
325 northern part of the  $31^\circ\text{E}$  section). When the heat budget is computed over a layer which extends  
326 further down, below the pycnocline (Fig. S8), the dominant heat source on January 15 is  
327 provided by vertical advection, suggesting that the heat provided to the surface by vertical

328 diffusion through the upper pycnocline is refueled by vertical advection. This is consistent with  
329 the positive vertical velocity found in the pycnocline (Fig. 6d). Actually, the time evolution of  
330 the vertical advection contribution between January 5 and 25 follows that of the Ekman suction  
331 velocity averaged over the *melt area* (later on, other processes are involved in the subsequent  
332 ice opening), suggesting a response to the surface stress curl.

333 The maximum of the vertical diffusive heat flux (Fig. 6b) is found in the area of substantial  
334 deepening of the mixed layer and where the mixed layer was the shallowest before the  
335 occurrence of the large melt event (Fig. 6a). Here, the mixed layer is defined as the depth at  
336 which the potential density exceeds the surface density by  $0.1 \text{ kg/m}^{-3}$ . The maximum of the  
337 diffusive heat flux also coincides with a narrow area where both the strongest near-surface  
338 stratification (maximum of the Brunt Väsälä frequency) (Fig. 6f), and the maximum vertical  
339 shear of velocity (Fig. 6e) are found. A vertical profile in the area of large heat flux (Fig. S9)  
340 shows that the Richardson number (computed using 5-day averaged variables) remains close to  
341 the critical values at the base of the mixed layer, suggesting that enhanced mixing is not only  
342 due to an increased input of energy at the surface by larger surface stress but might also result  
343 from increased shear instability at the mixed layer base. The January large melt event  
344 contributes to eastward expansion of the ice pack opening (compare SIC isopleths on January  
345 15 and 20 in figure 5b) that has started on January 10 (Fig. 4a).

346 Between February 24 and March 1 2006, a second large melt event occurs (Fig 4a) but the  
347 driving processes are distinct from those explaining the first event. On February 24, the MIZ  
348 north of Svalbard is fairly far north (Fig. 7c), as a consequence of the large ice opening which  
349 started at the beginning of January, and open water occupies a large area north of Svalbard,  
350 extending from the shelf northeastward to the slope region. Then, the large persistent northerly-  
351 northwesterly winds (Fig. 7a) during the next days drive a westward-southwestward ice drift,  
352 advecting the ice pack over the warm water present on the shelf-upper slope (Fig. 7g). The large



353 melt rate area (Fig. 7c) coincides indeed with the area of large ice transport convergence (Fig.  
354 7e). Despite it being a large melt event, this event is thus associated with a close-up of the ice  
355 pack (Fig 4a).

356 During the next winter, on December 26, 2006, a third large melt event begins, which is less  
357 intense than the two previous ones (Fig. 4a and 8g). The event, which is marked by moderate  
358 wind anomalies (Fig. 4b, 8a), lags by 1-2 weeks a strengthening of the boundary current  
359 between December 1 and December 16 (Fig. 4b and 8c). The increase of the current which can  
360 be traced back into the WSC, is associated with shoaling of the AW layer along the AW  
361 pathway in our study area (Fig. 8e and 9a,b). Despite moderate winds, these oceanic conditions  
362 may have favored enhanced vertical mixing on December 26, with a deepening and warming  
363 of the mixed layer, and subsequent ice melt (Fig. 9c).

#### 364 3.4.2 Large melt events over the simulation period

365 Building on the information gained from the 2006 case study, we pursue the analysis by  
366 performing a more systematic analysis of the large melt events over the full period of the  
367 simulation. Composites are constructed based on indices which best represent the different  
368 mechanisms that have been identified in the previous section in relation to the large melt events  
369 of year 2006.

##### 370 3.4.2.1 Large melt associated with ice advection over warm water.

371 In order to identify the large melt events which are associated with advection of ice over warm  
372 water, we define a sea surface temperature (SST) index as the average of the SST over the *melt*  
373 *area*. After analyzing the time evolution of the sea ice response, a time lag of 5 days was  
374 identified as the lag of maximum response and, therefore, the SST index is based on the SST  
375 averaged over the 5 days preceding the melt event. Composites of the ice melt rate averaged  
376 over the large melt events which are associated with a value of the SST index higher than -

377 1.5°C are constructed. 8 such large melt events are identified (Fig. 10), with two of them lasting  
378 10 days. The melt pattern associated with these events indicates enhanced melt occurring  
379 predominantly to the west of 27°E and mainly in the MIZ (Fig. 7d). As described for the second  
380 large melt event of the winter 2006, these large melt events are associated with high ice volume  
381 transport convergence (Fig. 7f), a process which is likely driven by strong northeasterly winds,  
382 as shown by the corresponding composites of the surface wind (Fig. 7b). In that case the  
383 convergence of the transport is due to advection of ice over open water, which is consistent  
384 with an ice edge being advected over the warm water pool located to the southwest of the ice  
385 edge (Fig 7h). The ice transport tends to overcompensate the ice melt, a balance which leads to  
386 an expansion of the ice cover (compare the black and blue lines in Fig. 7h). These close-up  
387 episodes can be consecutive to ice opening events (Fig. 10), which, as in 2006, can be sustained  
388 more than one month.

#### 389 3.4.2.2 Melt driven by large surface stress events

390 During the January 2006 event, wind driven mixing is the dominant mechanism. We then  
391 investigate if this process can drive other large melt events. Accordingly, we define a surface  
392 stress index by averaging, over the study area, the magnitude of the stress at the ocean surface  
393 (defined, for a given model grid cell, as a combination of the ice-ocean stress in the ice-covered  
394 part of the grid cell and the wind stress over the ice-free part). Large stress events have been  
395 defined as those which are associated with stress anomalies of amplitude larger than 1 standard  
396 deviation. 25 out of the 36 5-day periods of large melt which are not preceded by warm SSTs  
397 are associated with large stress (Fig. 10). The composite of the melt rate averaged over these  
398 events (Fig.11a) differs from the pattern in Figure 7d and displays a pattern with an enhanced  
399 signature above the boundary current pathway which extends eastwards into the pack ice as far  
400 as 40°E (a similar pattern is obtained when the composite is calculated over areas where the  
401 SIC exceeds 80%).

402 The pattern of the melt reflects that of the mean depth of the 0°C isotherm, with areas of larger  
403 melt located over shallow 0°C isotherm (compare Fig. 11a with Fig. 1). The largest melt is  
404 found around 24°E, where the boundary current narrows and the 0°C shoals due to the steep  
405 topography. A vertical section along 24°E shows the deepening of the mixed layer during the  
406 large melt events accompanied by the formation of negative temperature anomaly in the  
407 pycnocline just below the area of large melt (Fig. 11b). This would be consistent with an  
408 enhanced vertical heat flux from the pycnocline to the ocean surface, with a part being lost to  
409 slightly warm the mixed layer. North of the boundary current, where the mixed layer is deeper  
410 and the vertical temperature gradient is weaker, vertical mixing is not strong enough to affect  
411 significantly the ice melt. More generally, while the mean depth of the upper boundary of the  
412 AW layer is decisive for the pattern of sea ice melt, the composite differences of the AW depth  
413 based on the large melt events associated with large stress do not show significant anomalies,  
414 suggesting that Ekman suction anomalies would only contribute to sea ice melt in some specific  
415 cases as in January 2006.

416 These large melt events are not linked with a preferred wind direction (Fig. S10). This finding  
417 is consistent with correlation maps (not shown) between the melt index and the zonal and  
418 meridional components of the 10 m wind which do not display absolute values exceeding 0.2.  
419 In contrast, the correlation map of the melt index with the wind speed indicate absolute values  
420 of 0.4-0.5 north of Svalbard.

421 While wind driven mixing appears to be the dominant process to generate large melt events,  
422 not all the large melt events are associated with strong winds, and only slightly more than 50%  
423 of the large stress events can be linked to large melt events suggesting that other processes  
424 might modulate the melt.

425 3.4.2.3 Influence of the boundary current.

426 As aforementioned, the pathway of the boundary current is reflected in the pattern of the ice  
427 melt, suggesting that the current may play a role in the *winter variability* of the ice melt. A  
428 velocity index is thus defined based on the along-slope velocity distribution at 20°E,  
429 downstream of the merging of the AW Svalbard and Yermak branches. The index represents  
430 the mean velocity over the upper part (between 20 and 40 m depth) of the current core and does  
431 not consider the velocity variations in the Ekman layer. The composite of the mean velocity  
432 anomalies in the layer 20-40 m averaged over the strong current events (i.e events for which  
433 the velocity index exceeds 1 standard deviation) suggests a strengthening of the boundary  
434 current which occurs all along the current pathway, including in its upstream region in the WSC  
435 (see Fig. 8d). This anomalous circulation appears to be associated with anticyclonic wind  
436 anomalies centered on the northern Barents Sea (Fig. 8b). These strong current events are more  
437 persistent than the large melt events: their mean duration is about 15 days, but they can last up  
438 to 25 days. They are linked with a shoaling of the 0°C isotherm (Fig. 8f), and subsurface  
439 warming along the AW pathway, with the strongest values located east of 24°E (not shown).  
440 This warming remains significant when the temperature lags by 10 to 15 days. It is also  
441 interesting to notice the local impact of these strong current events on the surface stress, which  
442 depends, in ice covered areas, on the relative velocity of the ice with respect to the ocean surface  
443 current. During these events, a dipole pattern of the surface stress curl is observed along the  
444 path of the boundary current (Fig. 12), which could contribute to the shoaling of the AW layer.  
445 Based on the boundary current velocity index, we identify 17 5-day episodes of large melt  
446 events which are preceded by or in phase with strong current events (Fig. 10a). With the  
447 increase of the subsurface heat content, the strong current events generate favorable ocean  
448 conditions for increased vertical transfer of heat through mixing to the ice (Fig. 8h). This  
449 mechanism does not require strong winds, as demonstrated by the significant proportion (6 out

450 of 17) of events which occur during moderate surface stress conditions, and by the event  
451 documented on December 25, 2006.

#### 452 **4. Discussion**

453 Focusing on the *winter variability* of the coupled ice-ocean system north of Svalbard, we have  
454 shown that sea ice melt can occur as large amplitude, short term (5-10 days) events, which are  
455 mainly located over the AW pathway. Large melt rates are found in the MIZ but also in the  
456 interior pack ice where they can reach as far east as 40°E. Concomitant direct observations of  
457 sea ice melt rates and ice-ocean heat fluxes are too sparse to enable systematic validation of the  
458 model results. However, local observations of large ice-ocean heat fluxes during the N-ICE2015  
459 drift (Meyer et al., 2017; Graham et al., 2019), can be used to check if these are concomitant to  
460 simulated large melt events. The first of these large flux events which occurred during a storm  
461 between February 3 and 9 indeed emerges as a large melt event in our simulation, being  
462 identified as an event of advection of ice over warm waters (Fig. 10). An episode of strong  
463 northerly winds between February 6 and 9 (also mentioned by Graham et al., 2019), indeed  
464 followed strong southerly winds during February 4 to 5 which might have led to an ice cover  
465 opening. The second observed event also occurred during a storm and the model does show an  
466 increase of the 5-day averaged melt rate over the period including this storm (not shown), but  
467 not high enough for this event to be identified as large melt event. The reason for the low melt  
468 rate may be related to the spatial-temporal distribution of the wind anomalies. The strong winds  
469 were mostly confined to the northwestern part of our study region (not shown) and lasted only  
470 one day, and therefore might have not appeared as strong enough when averaged over the study  
471 area and a 5-day period.

472 An increase of the ice-ocean heat flux was indeed observed to coincide with storm activity and  
473 the presence of AW in the sub-surface ocean in earlier observations (Meyer et al. 2017; Graham

474 et al., 2019). In our simulation wind driven mixing, entraining warm AW into the upper layer,  
475 appears as the dominant mechanism generating large ice melt events. These episodes of  
476 enhanced mixing correspond to above-average ocean surface stress. Although 5-day averaged  
477 wind events are used in this analysis, the role of high frequency atmospheric forcing is certainly  
478 important. For instance, the largest melt event on March 11, 2004 (Fig 10a), is linked to a  
479 moderate 5-day average surface stress (Fig. 10b), while wind data at higher temporal resolution  
480 indeed reveal weak winds over this 5-day period except for a short (1 day) wind burst. To  
481 investigate if some of the melt events may be mis-classified with regards to the strength of the  
482 wind stress forcing, we compare two wind speed indices. One is based on the maximum of the  
483 5-day averaged wind speed (the surface stress, which is an output of the model saved every 5-  
484 day and cannot be used here) and the other is based on the 5-day mean wind speed. Both indices  
485 are averaged over the study area. The two groups of large melt events associated with large  
486 values (i.e  $> 1$  std) of these two wind indices are very similar, suggesting the robustness of our  
487 analysis based on 5-day averages.

488 Only 55% of the large stress events present in the simulation are associated with large melt  
489 events, 45% of the large stress events are thus associated with moderate melt events. Part of the  
490 explanation may be found in the different surface stress distributions between the two groups  
491 of melt events which reveal a comparatively lower mean stress and absence of very large stress  
492 events in the moderate melt group. However, even when restricting the comparison between  
493 the two groups to events with stress lower than  $0.13 \text{ N.m}^{-2}$  (so that the mean stress of the two  
494 groups become similar) we find that the vertical shear of the current at the base of the mixed  
495 layer in the *melt area* is indeed drastically reduced during moderate melt events compared with  
496 the large melt events (the difference between the mean shear of the two groups is large and  
497 statistically significant). This increased shear is also accompanied by a significant strengthening  
498 of the stratification during the large melt events, but overall, the area where the Richardson

499 number remains in the range of the critical values is significantly larger in the large melt group,  
500 suggesting that shear instability could contribute to enhance the vertical heat fluxes and the melt  
501 rate during large melt events.

502 The mean melt rate in our simulation is clearly shaped by the depth of the upper boundary of  
503 the AW layer (taking, e.g., the depth of the 0°C isotherm as a proxy for the upper limit of this  
504 layer), with the largest melt located above regions where the AW layer is shallower. This is a  
505 strong suggestion that the topography of the AW layer maintained by the boundary current  
506 exerts some preconditioning with regards to further vertical mixing events. However, melt  
507 events triggered by large surface stress (which represent the large majority of the large melt  
508 events) are not sensitive to the depth of the AW layer. While this may be partly due to the weak  
509 variability of the AW layer, it also suggests that upwelling (in particular shelf break upwelling  
510 in the region of enhanced melt) cannot be a leading mechanism in producing large sea ice melt,  
511 a result which is probably linked to the rather deep shelf found north of Svalbard (as already  
512 mentioned by Randelhoff and Sundfjord, 2018). Some occurrences of AW upwelling can  
513 however be identified which lead to subsurface warming and enhanced upward heat fluxes and  
514 sea ice melt. In such cases, the AW layer rise would accompany a strengthening of the boundary  
515 current which may have been forced by an anticyclonic wind anomaly centered over the  
516 northern Barents Sea. This latter finding is consistent with conclusions by Lien et al. (2013)  
517 and Herbaut et al. (2017) who suggested that negative wind curl over the Northern Barents Sea  
518 could enhance the anticyclonic circulation on the shelf around the Svalbard Archipelago, and  
519 in particular strengthen the boundary current in the western Eurasian basin. However, the  
520 impact of the current on its own may be limited since, despite 50% of those large melt events  
521 not associated with ice advection over warm SST appear to be preconditioned by large current,  
522 most of these large current events need large stress to trigger a large response of the sea ice  
523 melt.

524 None of the large melt events show significant deepening of the surface mixed layer in the  
525 simulation. In addition, buoyancy driven convection does not seem to play a role in the  
526 occurrence of large melt events since only one episode of deep convection is detected in our  
527 study area. The event occurs in the boundary current during periods of ice-free conditions and  
528 can lead to mixed layers as deep as 300-400 meters. Buoyancy driven convection has been  
529 mentioned as the process responsible for winter ventilation of the Atlantic Water layer in the  
530 interior ice pack in the deep Nansen Basin (Polyakov et al., 2017, 2020). Occurrences of ocean  
531 convection under low sea ice concentrations, as found in our analysis, have also been reported  
532 in the literature in observations (Pérez-Hernández et al., 2019) and ocean model operational  
533 systems (Athanasé et al., 2020). While convective mixing in open water does not contribute  
534 directly to sea ice melt in our simulation, it does contribute in a delayed mode to the sea ice  
535 volume budget by transferring large amounts of heat to the surface layer and thus slowing down  
536 the return to heavier ice conditions. Advection of ice into surface waters which have been  
537 previously warmed by convective mixing can indeed generate large melt events subsequently  
538 to the convective events.

539 The January 2006 case study illustrates how large melt rates can contribute to the eastward  
540 retreat of the ice edge north of Svalbard in mid-winter. Composites of the SIC change during  
541 large melt events over the entire simulation show that the largest reduction in the sea ice  
542 concentration is usually found over the shelf north of Svalbard, decreasing to values less than  
543 10% over a 5-day period further offshore, while northeast of Svalbard, as far as 36°E, anomalies  
544 do not exceed 5% (Fig. 13a). However, similar composites based on the events of large ice  
545 transport divergence do not show the same decrease of ice concentrations northeast of Svalbard  
546 (Fig. 13b) which suggests that ocean-driven sea ice melt is the main mechanism responsible for  
547 the episodes of sea ice concentration decrease observed northeast of Svalbard. Conversely, on  
548 the shelf north of Svalbard, the SIC decrease is mainly explained by large divergence of the ice



549 volume transport, which is driven by southeasterly wind (not shown) enhancing northward  
550 advection of thinner ice from the south (Fig. 13c).

551 This mechanism was also mentioned by Onarheim et al., (2014) to explain the interannual sea  
552 ice variability in the same region. However, while these authors only stressed the impact of the  
553 wind stress on the ice transport, we could show that, in case of particularly high wind speed, a  
554 thermodynamic effect can add to the dynamic response of the ice. Because of this additional  
555 effect, the impact of the wind on the sea ice may be stronger than previously postulated. In  
556 particular, during periods of large southerly winds, larger ice divergence and enhanced  
557 entrainment could combine to drive particularly large ice cover decrease, with each process  
558 driving ice reduction over a specific distinct area. Actually, the four largest ice cover decrease  
559 events in 2000, 2006, 2013 and 2017, as identified by the EOF analysis of the winter SIC (Fig.  
560 2), occurred during periods of large southerly winds (not shown).

561 Graham et al., (2019), using ice drift and area from satellite observations and constant ice  
562 thickness to compute the ice volume budget, estimate that over a domain equivalent to our study  
563 area, ice-ocean heat flux up to  $600 \text{ W/m}^2$  (to account for the melt rate estimated) over about 2  
564 days could occur during a storm. In the simulation such large ice-ocean heat fluxes can happen  
565 in localized areas but on average over the study area, the 5-day heat fluxes exceed only twice  
566  $100 \text{ W/m}^2$  (Fig 2e). This discrepancy between the heat fluxes can be due to the model limitations  
567 and to the large uncertainties inherent of the method used by Graham et al. (2019).

568 Over the time of the simulation, we do not find significant correlation between a preferred wind  
569 direction and sea ice melt index. This finding is in contrast to Duarte et al.'s (2020) conclusion  
570 which highlights significant correlations between southeasterly winds and sea ice melt rates.  
571 The fact that Duarte et al. calculation is done over a region that is located closer to the ice edge  
572 and further west compared to our study area may explain the discrepancy. This limitation in the

573 comparison is even reinforced by the fact that, as aforementioned, the processes highlighted in  
574 our simulation take place in an area which is mainly covered by compact ice. In contrast, the  
575 domain analyzed by Duarte et al. includes a larger fraction of the MIZ where, under northerly  
576 winds, ice advection can lead to large melt when ice encounters warm waters.

577 The goal of the present analysis was to explore the sensitivity of the Arctic sea ice cover to  
578 some ocean processes. As such, it emphasizes the role of the mechanical surface forcing in  
579 controlling important processes like vertical mixing, advection or sea ice motion, but also the  
580 upper ocean temperature distribution and stratification. The sensitivity of these processes to the  
581 forcing is however likely to depend on the mean sea ice state. Underestimated sea ice thickness  
582 in the simulation, for instance, should imply a more efficient energy transfer between the  
583 atmosphere and the ocean, an effect however modulated by the higher compactness of the ice  
584 cover so that the net impact is difficult to estimate. On the other hand, due to too compact sea  
585 ice in the model climatology, the SIC variability is underestimated along the AW boundary  
586 current (Fig. 2) and the ice conditions simulated in this region may therefore be more  
587 representative of more compact ice conditions observed farther downstream. One of the  
588 consequences of this could be an underestimation of the number of large melt events associated  
589 with ice advection over warm SST: with a MIZ, which would be located closer to the boundary  
590 current, close-up of the ice following opening should be accompanied by larger ice-ocean heat  
591 flux. While the simulated winter AW temperature distribution in the West Spitsbergen Current  
592 matches reasonably well the observations in Fram Strait, the AW core appears too warm in  
593 winter northeast of Svalbard. While this discrepancy might be a consequence of a shift in the  
594 seasonal cycle as discussed earlier, the overestimated ice extent and compactness during most  
595 of the year may also contribute to limiting direct heat loss through air sea interactions. In  
596 principle, the warm bias in the upper part of the AW layer (that part which is eroded during  
597 large melt events) should lead to overestimated upward heat flux and ice bottom melt under

598 vertical mixing. However, larger sea ice extent in the model is also associated with a fresher  
599 ocean surface layer and a stronger upper halocline which limits the mixing efficiency and  
600 mitigates the impact of warmer subsurface water, therefore suggesting that our results are  
601 conservative regarding the impact of vertical mixing on the sea ice. Still, further investigation  
602 would be necessary to disentangle the different contributions to the model sensitivity with  
603 respect to mixing. More generally, additional observations are needed to validate the  
604 relationship between winter sea ice melt and the underlying ocean dynamics and its dependency  
605 on the different sea ice regimes, a prerequisite to assessing their impact on the sea ice cover at  
606 the basin scale.

## 607 **5. Summary and conclusions**

608 Using an ice-ocean simulation forced by the ERA-I reanalysis over the period 1997-2017, large  
609 ice melt events have been identified in winter in the area north of Svalbard. These short duration  
610 (5-10 days) melt events, which show enhanced signature along the AW pathway and can occur  
611 in the pack ice, can contribute to 40% of the total winter ice melt. The net growth pattern  
612 associated with the large melt events clearly reflects the mean pattern of the 0°C isotherm.  
613 Different types of events have been established depending on the scenario responsible for  
614 enhanced sea ice melt. Enhanced melt can happen concomitantly to large ice transport over  
615 warm surface waters in the ice edge region, a scenario which predominates during the close-up  
616 following a large sea ice cover opening. This process is also a factor controlling the SIC budget  
617 on interannual time scales (Lundesgaard et al., 2021) and may become more frequent in the  
618 area in the future with the expansion of the ice-free areas and the ice becoming more mobile.

619 Apart from these events, large melt rates are mainly driven by increased mixing and entrainment  
620 of warm water into the mixed layer. Strong winds are the main driver of the large mixing events  
621 but the shoaling of the pycnocline during episodes of strengthened boundary current can also

622 modulate the vertical heat flux into the mixed layer. Shear instability is likely to enhance the  
623 heat transfer to the surface layer. While wind driven mixing appears to be the controlling factor  
624 of the winter sea ice melt in the western Eurasian Basin, thermohaline convection has been  
625 pointed out as the main driver of the winter ice melt in the eastern Eurasian Basin (Polyakov et  
626 al., 2017). With the expansion of the ocean conditions prevailing north of Svalbard to the  
627 eastern Eurasian Basin as a result of the Atlantification of the Arctic Ocean (Polyakov, 2017;  
628 Polyakov et al., 2020a), the nature of the drivers of the ice melt in this part of the basin may  
629 also change and the impact of wind driven mixing and boundary current dynamics on the future  
630 of the Arctic sea ice cover become visible on a wider scale.

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638 The model initialization is based on the PHC3 climatology  
639 ([http://psc.apl.washington.edu/nonwp\\_projects/PHC/Climatology.html](http://psc.apl.washington.edu/nonwp_projects/PHC/Climatology.html)). Continental runoff is  
640 extracted from the monthly climatology of Dai and Trenberth (2002)  
641 (<https://rda.ucar.edu/datasets/ds551.0/>). The atmospheric forcing is based on the ERA-Interim  
642 reanalysis (<https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim>). The  
643 ISCCP climatology of the radiation fluxes over the North Atlantic  
644 (<https://isccp.giss.nasa.gov/projects/flux.html>) and the surface air temperature from the IABP  
645 climatology ([http://research.jisao.washington.edu/data\\_sets/iabppoles/](http://research.jisao.washington.edu/data_sets/iabppoles/)) have been used.  
646 Precipitation is extracted from the DFS5 dataset (<https://ige-meom-opensap.univ-grenoble->

647 [alpes.fr/thredds/catalog/meomopendap/extract/FORCING\\_ATMOSPHERIQUE/DFS5.2/ALL](https://alpes.fr/thredds/catalog/meomopendap/extract/FORCING_ATMOSPHERIQUE/DFS5.2/ALL)  
648 [/catalog.html](#)). The sea ice concentration data were retrieved from the Integrated Climate  
649 Data Center of Hamburg University ([https://www.cen.uni-](https://www.cen.uni-hamburg.de/en/icdc/data/cryosphere/seaiceconcentration-asi-ssmi.html)  
650 [hamburg.de/en/icdc/data/cryosphere/seaiceconcentration-asi-ssmi.html](https://www.cen.uni-hamburg.de/en/icdc/data/cryosphere/seaiceconcentration-asi-ssmi.html)).

651

652 **References:**  
653

- 654 Årthun, M., Eldevik, T., Smedsrud, L.H., 2019. The Role of Atlantic Heat Transport in Future Arctic  
655 Winter Sea Ice Loss. *J. Clim.* 32, 3327–3341. <https://doi.org/10.1175/JCLI-D-18-0750.1>
- 656 Årthun, M., Eldevik, T., Smedsrud, L.H., Skagseth, Ø., Ingvaldsen, R.B., 2012. Quantifying the  
657 Influence of Atlantic Heat on Barents Sea Ice Variability and Retreat. *J. Clim.* 25, 4736–4743.  
658 <https://doi.org/10.1175/JCLI-D-11-00466.1>
- 659 Årthun, M., Eldevik, T., Viste, E., Drange, H., Furevik, T., Johnson, H.L., Keenlyside, N.S., 2017. Skillful  
660 prediction of northern climate provided by the ocean. *Nat. Commun.* 8, 15875.  
661 <https://doi.org/10.1038/ncomms15875>
- 662 Athanase, M., Provost, C., Pérez-Hernández, M.D., Sennéchaël, N., Bertosio, C., Artana, C., Garric, G.,  
663 Lellouche, J.-M., 2020. Atlantic Water Modification North of Svalbard in the Mercator  
664 Physical System From 2007 to 2020. *J. Geophys. Res. Oceans* 125, e2020JC016463.  
665 <https://doi.org/10.1029/2020JC016463>
- 666 Beszczynska-Möller, A., Fahrbach, E., Schauer, U., Hansen, E., 2012. Variability in Atlantic water  
667 temperature and transport at the entrance to the Arctic Ocean, 1997–2010. *ICES J. Mar. Sci.*  
668 69, 852–863. <https://doi.org/10.1093/icesjms/fss056>
- 669 Blanke, B., Delecluse, P., 1993. Variability of the Tropical Atlantic Ocean Simulated by a General  
670 Circulation Model with Two Different Mixed-Layer Physics. *J. Phys. Oceanogr.* 23, 1363–1388.  
671 [https://doi.org/10.1175/1520-0485\(1993\)023<1363:VOTTAO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1993)023<1363:VOTTAO>2.0.CO;2)
- 672 Carmack, E., Polyakov, I., Padman, L., Fer, I., Hunke, E., Hutchings, J., Jackson, J., Kelley, D., Kwok, R.,  
673 Layton, C., Melling, H., Perovich, D., Persson, O., Ruddick, B., Timmermans, M.-L., Toole, J.,  
674 Ross, T., Vavrus, S., Winsor, P., 2015. Toward Quantifying the Increasing Role of Oceanic Heat  
675 in Sea Ice Loss in the New Arctic. *Bull. Am. Meteorol. Soc.* 96, 2079–2105.  
676 <https://doi.org/10.1175/BAMS-D-13-00177.1>
- 677 Close, S., Houssais, M.-N., Herbaut, C., 2015. Regional dependence in the timing of onset of rapid  
678 decline in Arctic sea ice concentration. *J. Geophys. Res. Oceans* 120, 8077–8098.  
679 <https://doi.org/10.1002/2015JC011187>
- 680 Cokelet, E.D., Tervalon, N., Bellingham, J.G., 2008. Hydrography of the West Spitsbergen Current,  
681 Svalbard Branch: Autumn 2001. *J. Geophys. Res. Oceans* 113.  
682 <https://doi.org/10.1029/2007JC004150>
- 683 Crews, L., Sundfjord, A., Hattermann, T., 2019. How the Yermak Pass Branch Regulates Atlantic Water  
684 Inflow to the Arctic Ocean. *J. Geophys. Res. Oceans* 124, 267–280.  
685 <https://doi.org/10.1029/2018JC014476>
- 686 Dai, A., Trenberth, K.E., 2002. Estimates of Freshwater Discharge from Continents: Latitudinal and  
687 Seasonal Variations. *J. Hydrometeorol.* 3, 660–687. [https://doi.org/10.1175/1525-7541\(2002\)003<0660:EOFDFC>2.0.CO;2](https://doi.org/10.1175/1525-7541(2002)003<0660:EOFDFC>2.0.CO;2)
- 689 Dee, D.P., Uppala, S.M., Simmons, A.J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda,  
690 M.A., Balsamo, G., Bauer, P., Bechtold, P., Beljaars, A.C.M., Berg, L. van de, Bidlot, J.,  
691 Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A.J., Haimberger, L., Healy, S.B.,  
692 Hersbach, H., Hólm, E.V., Isaksen, I., Kållberg, P., Köhler, M., Matricardi, M., McNally, A.P.,  
693 Monge-Sanz, B.M., Morcrette, J.-J., Park, B.-K., Peubey, C., Rosnay, P. de, Tavolato, C.,  
694 Thépaut, J.-N., Vitart, F., 2011. The ERA-Interim reanalysis: configuration and performance of  
695 the data assimilation system. *Q. J. R. Meteorol. Soc.* 137, 553–597.  
696 <https://doi.org/10.1002/qj.828>
- 697 Duarte, P., Sundfjord, A., Meyer, A., Hudson, S.R., Spreen, G., Smedsrud, L.H., 2020. Warm Atlantic  
698 Water Explains Observed Sea Ice Melt Rates North of Svalbard. *J. Geophys. Res. Oceans* 125,  
699 e2019JC015662. <https://doi.org/10.1029/2019JC015662>
- 700 Dussin, R., Barnier, B., Brodeau, L., Molines, J.-M., 2018. The making of the Drakkar Forcing set DFS5.  
701 <https://doi.org/10.5281/zenodo.1209243>

702 Falk-Petersen, S., Pavlov, V., Berge, J., Cottier, F., Kovacs, K.M., Lydersen, C., 2015. At the rainbow's  
703 end: high productivity fueled by winter upwelling along an Arctic shelf. *Polar Biol.* 38, 5–11.  
704 <https://doi.org/10.1007/s00300-014-1482-1>

705 Graham, R.M., Itkin, P., Meyer, A., Sundfjord, A., Spreen, G., Smedsrud, L.H., Liston, G.E., Cheng, B.,  
706 Cohen, L., Divine, D., Fer, I., Fransson, A., Gerland, S., Haapala, J., Hudson, S.R., Johansson,  
707 A.M., King, J., Merkouriadi, I., Peterson, A.K., Provost, C., Randelhoff, A., Rinke, A., Rösel, A.,  
708 Sennéchaël, N., Walden, V.P., Duarte, P., Assmy, P., Steen, H., Granskog, M.A., 2019. Winter  
709 storms accelerate the demise of sea ice in the Atlantic sector of the Arctic Ocean. *Sci. Rep.* 9,  
710 1–16. <https://doi.org/10.1038/s41598-019-45574-5>

711 Guthrie, J.D., Morison, J.H., Fer, I., 2013. Revisiting internal waves and mixing in the Arctic Ocean. *J.*  
712 *Geophys. Res. Oceans* 118, 3966–3977. <https://doi.org/10.1002/jgrc.20294>

713 Ivanov, V., Alexeev, V., Koldunov, N.V., Repina, I., Sandø, A.B., Smedsrud, L.H., Smirnov, A., 2015.  
714 Arctic Ocean Heat Impact on Regional Ice Decay: A Suggested Positive Feedback. *J. Phys.*  
715 *Oceanogr.* 46, 1437–1456. <https://doi.org/10.1175/JPO-D-15-0144.1>

716 Ivanov, V.V., Alexeev, V.A., Repina, I., Koldunov, N.V., Smirnov, A., 2012. Tracing Atlantic Water  
717 Signature in the Arctic Sea Ice Cover East of Svalbard [WWW Document]. *Adv. Meteorol.*  
718 <https://doi.org/10.1155/2012/201818>

719 Kaleschke, L., Lüpkes, C., Vihma, T., Haarpaintner, J., Bochert, A., Hartmann, J., Heygster, G., 2001.  
720 SSM/I Sea Ice Remote Sensing for Mesoscale Ocean-Atmosphere Interaction Analysis. *Can. J.*  
721 *Remote Sens.* 27, 526–537. <https://doi.org/10.1080/07038992.2001.10854892>

722 Koenig, Z., Provost, C., Sennéchaël, N., Garric, G., Gascard, J.-C., 2017. The Yermak Pass Branch: A  
723 Major Pathway for the Atlantic Water North of Svalbard? *J. Geophys. Res. Oceans* 122, 9332–  
724 9349. <https://doi.org/10.1002/2017JC013271>

725 Koenig, Z., Provost, C., Villacieros-Robineau, N., Sennéchaël, N., Meyer, A., 2016. Winter ocean-ice  
726 interactions under thin sea ice observed by IAOOS platforms during N-ICE2015: Salty surface  
727 mixed layer and active basal melt. *J. Geophys. Res. Oceans* 121, 7898–7916.  
728 <https://doi.org/10.1002/2016JC012195>

729 Large, W.G., Yeager, S.G., n.d. Diurnal to Decadal Global Forcing For Ocean and Sea-Ice Models: The  
730 Data Sets and Flux Climatologies 113.

731 Lind, S., Ingvaldsen, R.B., 2012. Variability and impacts of Atlantic Water entering the Barents Sea  
732 from the north. *Deep Sea Res. Part Oceanogr. Res. Pap.* 62, 70–88.  
733 <https://doi.org/10.1016/j.dsr.2011.12.007>

734 Lundesgaard, Ø., Sundfjord, A., Renner, A.H.H., 2021. Drivers of Interannual Sea Ice Concentration  
735 Variability in the Atlantic Water Inflow Region North of Svalbard. *J. Geophys. Res. Oceans*  
736 126, e2020JC016522. <https://doi.org/10.1029/2020JC016522>

737 Madec, Gurvan, 2008. NEMO ocean engine. Note du Pôle de modélisation, Institut Pierre-Simon  
738 Laplace (IPSL), France, No 27, ISSN No 1288-1619.

739 Menze, S., Ingvaldsen, R.B., Haugan, P., Fer, I., Sundfjord, A., Beszczynska-Moeller, A., Falk-Petersen,  
740 S., 2019. Atlantic Water Pathways Along the North-Western Svalbard Shelf Mapped Using  
741 Vessel-Mounted Current Profilers. *J. Geophys. Res. Oceans* 124, 1699–1716.  
742 <https://doi.org/10.1029/2018JC014299>

743 Meyer, A., Fer, I., Sundfjord, A., Peterson, A.K., 2017a. Mixing rates and vertical heat fluxes north of  
744 Svalbard from Arctic winter to spring. *J. Geophys. Res. Oceans* 4569–4586.  
745 [https://doi.org/10.1002/2016JC012441@10.1002/\(ISSN\)2169-9291.NICE1](https://doi.org/10.1002/2016JC012441@10.1002/(ISSN)2169-9291.NICE1)

746 Meyer, A., Sundfjord, A., Fer, I., Provost, C., Robineau, N.V., Koenig, Z., Onarheim, I.H., Smedsrud,  
747 L.H., Duarte, P., Dodd, P.A., Graham, R.M., Schmidtko, S., Kauko, H.M., 2017b. Winter to  
748 summer oceanographic observations in the Arctic Ocean north of Svalbard. *J. Geophys. Res.*  
749 *Oceans* 122, 6218–6237. <https://doi.org/10.1002/2016JC012391>

750 Onarheim, I.H., Eldevik, T., Smedsrud, L.H., Stroeve, J.C., 2018. Seasonal and Regional Manifestation  
751 of Arctic Sea Ice Loss. *J. Clim.* 31, 4917–4932. <https://doi.org/10.1175/JCLI-D-17-0427.1>

752 Onarheim, I.H., Smedsrud, L.H., Ingvaldsen, R.B., Nilsen, F., 2014. Loss of sea ice during winter north  
753 of Svalbard. *Tellus Dyn. Meteorol. Oceanogr.* 66, 23933.  
754 <https://doi.org/10.3402/tellusa.v66.23933>

755 Pérez-Hernández, M.D., Pickart, R.S., Pavlov, V., Våge, K., Ingvaldsen, R., Sundfjord, A., Renner,  
756 A.H.H., Torres, D.J., Erofeeva, S.Y., 2017. The Atlantic Water boundary current north of  
757 Svalbard in late summer. *J. Geophys. Res. Oceans* 122, 2269–2290.  
758 <https://doi.org/10.1002/2016JC012486>

759 Pérez-Hernández, M.D., Pickart, R.S., Torres, D.J., Bahr, F., Sundfjord, A., Ingvaldsen, R., Renner,  
760 A.H.H., Beszczynska-Möller, A., Appen, W.-J. von, Pavlov, V., 2019. Structure, Transport, and  
761 Seasonality of the Atlantic Water Boundary Current North of Svalbard: Results From a  
762 Yearlong Mooring Array. *J. Geophys. Res. Oceans* 124, 1679–1698.  
763 <https://doi.org/10.1029/2018JC014759>

764 Pickart, R.S., Moore, G.W.K., Torres, D.J., Fratantoni, P.S., Goldsmith, R.A., Yang, J., 2009. Upwelling  
765 on the continental slope of the Alaskan Beaufort Sea: Storms, ice, and oceanographic  
766 response. *J. Geophys. Res. Oceans* 114. <https://doi.org/10.1029/2008JC005009>

767 Polyakov, I.V., Alkire, M.B., Bluhm, B.A., Brown, K.A., Carmack, E.C., Chierici, M., Danielson, S.L.,  
768 Ellingsen, I., Ershova, E.A., Gårdfeldt, K., Ingvaldsen, R.B., Pnyushkov, A.V., Slagstad, D.,  
769 Wassmann, P., 2020a. Borealization of the Arctic Ocean in Response to Anomalous Advection  
770 From Sub-Arctic Seas. *Front. Mar. Sci.* 7. <https://doi.org/10.3389/fmars.2020.00491>

771 Polyakov, I.V., Pnyushkov, A.V., Alkire, M.B., Ashik, I.M., Baumann, T.M., Carmack, E.C., Goszczko, I.,  
772 Guthrie, J., Ivanov, V.V., Kanzow, T., Krishfield, R., Kwok, R., Sundfjord, A., Morison, J.,  
773 Rember, R., Yulin, A., 2017. Greater role for Atlantic inflows on sea-ice loss in the Eurasian  
774 Basin of the Arctic Ocean. *Science* 356, 285–291. <https://doi.org/10.1126/science.aai8204>

775 Polyakov, I.V., Rippeth, T.P., Fer, I., Alkire, M.B., Baumann, T.M., Carmack, E.C., Ingvaldsen, R., Ivanov,  
776 V.V., Janout, M., Lind, S., Padman, L., Pnyushkov, A.V., Rember, R., 2020b. Weakening of Cold  
777 Halocline Layer Exposes Sea Ice to Oceanic Heat in the Eastern Arctic Ocean. *J. Clim.* 33,  
778 8107–8123. <https://doi.org/10.1175/JCLI-D-19-0976.1>

779 Polyakov, I.V., Rippeth, T.P., Fer, I., Baumann, T.M., Carmack, E.C., Ivanov, V.V., Janout, M., Padman,  
780 L., Pnyushkov, A.V., Rember, R., 2020c. Intensification of Near-Surface Currents and Shear in  
781 the Eastern Arctic Ocean. *Geophys. Res. Lett.* 47, e2020GL089469.  
782 <https://doi.org/10.1029/2020GL089469>

783 Polyakov, I.V., Timokhov, L.A., Alexeev, V.A., Bacon, S., Dmitrenko, I.A., Fortier, L., Frolov, I.E.,  
784 Gascard, J.-C., Hansen, E., Ivanov, V.V., Laxon, S., Mauritzen, C., Perovich, D., Shimada, K.,  
785 Simmons, H.L., Sokolov, V.T., Steele, M., Toole, J., 2010. Arctic Ocean Warming Contributes  
786 to Reduced Polar Ice Cap. *J. Phys. Oceanogr.* 40, 2743–2756.  
787 <https://doi.org/10.1175/2010JPO4339.1>

788 Randelhoff, A., Sundfjord, A., 2018. Short commentary on marine productivity at Arctic shelf breaks:  
789 upwelling, advection and vertical mixing. *Ocean Sci.* 14, 293–300.  
790 <https://doi.org/10.5194/os-14-293-2018>

791 Renner, A.H.H., Sundfjord, A., Janout, M.A., Ingvaldsen, R.B., Beszczynska-Möller, A., Pickart, R.S.,  
792 Pérez-Hernández, M.D., 2018. Variability and Redistribution of Heat in the Atlantic Water  
793 Boundary Current North of Svalbard. *J. Geophys. Res. Oceans* 123, 6373–6391.  
794 <https://doi.org/10.1029/2018JC013814>

795 Ricker, R., Kauker, F., Schweiger, A., Hendricks, S., Zhang, J., Paul, S., 2021. Evidence for an Increasing  
796 Role of Ocean Heat in Arctic Winter Sea Ice Growth. *J. Clim.* 34, 5215–5227.  
797 <https://doi.org/10.1175/JCLI-D-20-0848.1>

798 Rigor, I.G., Colony, R.L., Martin, S., 2000. Variations in Surface Air Temperature Observations in the  
799 Arctic, 1979–97. *J. Clim.* 13, 896–914. [https://doi.org/10.1175/1520-0442\(2000\)013<0896:VISATO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<0896:VISATO>2.0.CO;2)

800  
801 Rossow, W.B., Schiffer, R.A., 1999. Advances in Understanding Clouds from ISCCP. *Bull. Am.*  
802 *Meteorol. Soc.* 80, 2261–2288. [https://doi.org/10.1175/1520-0477\(1999\)080<2261:AIUCFI>2.0.CO;2](https://doi.org/10.1175/1520-0477(1999)080<2261:AIUCFI>2.0.CO;2)

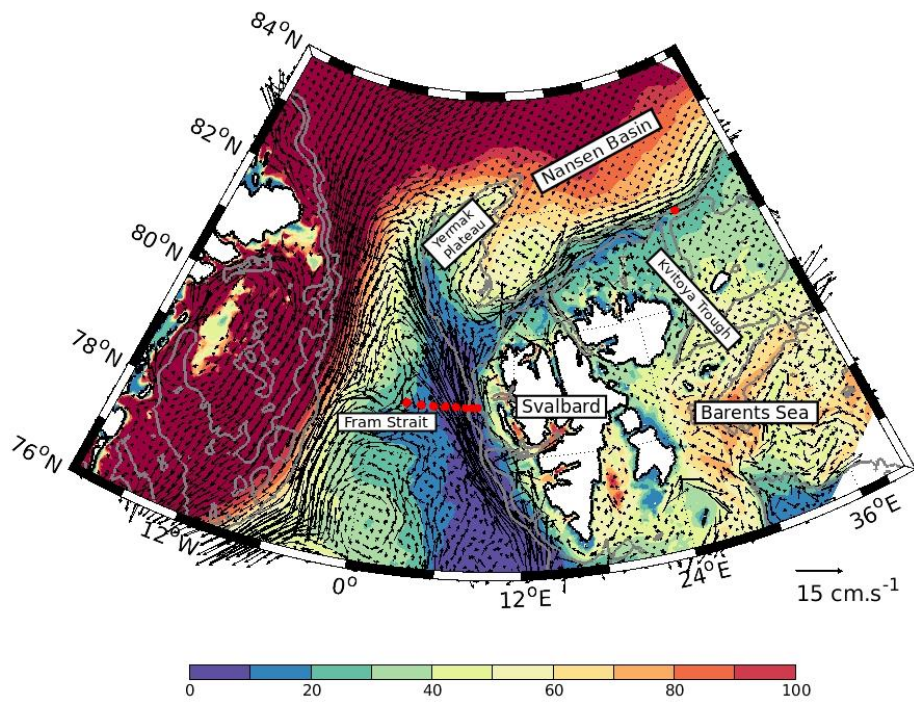
803



804 Rousset, C., Vancoppenolle, M., Madec, G., Fichefet, T., Flavoni, S., Barthélemy, A., Benshila, R.,  
805 Chanut, J., Lévy, C., Masson, S., Vivier, F., 2015. The Louvain-La-Neuve sea ice model LIM3.6:  
806 global and regional capabilities. *Geosci. Model Dev.* 8, 2991–3005.  
807 <https://doi.org/10.5194/gmd-8-2991-2015>  
808 Rudels, B., Jones, E.P., Schauer, U., Eriksson, P., 2004. Atlantic sources of the Arctic Ocean surface  
809 and halocline waters. *Polar Res.* 23, 181–208. [https://doi.org/10.1111/j.1751-](https://doi.org/10.1111/j.1751-8369.2004.tb00007.x)  
810 [8369.2004.tb00007.x](https://doi.org/10.1111/j.1751-8369.2004.tb00007.x)  
811 Sirevaag, A., Fer, I., 2009. Early Spring Oceanic Heat Fluxes and Mixing Observed from Drift Stations  
812 North of Svalbard. *J. Phys. Oceanogr.* 39, 3049–3069.  
813 <https://doi.org/10.1175/2009JPO4172.1>  
814 Skagseth, Ø., Furevik, T., Ingvaldsen, R., Loeng, H., Mork, K.A., Orvik, K.A., Ozhigin, V., 2008. Volume  
815 and Heat Transports to the Arctic Ocean Via the Norwegian and Barents Seas, in: Dickson,  
816 R.R., Meincke, J., Rhines, P. (Eds.), *Arctic–Subarctic Ocean Fluxes: Defining the Role of the*  
817 *Northern Seas in Climate*. Springer Netherlands, Dordrecht, pp. 45–64.  
818 [https://doi.org/10.1007/978-1-4020-6774-7\\_3](https://doi.org/10.1007/978-1-4020-6774-7_3)  
819 Spall, M.A., 2013. On the Circulation of Atlantic Water in the Arctic Ocean. *J. Phys. Oceanogr.* 43,  
820 2352–2371. <https://doi.org/10.1175/JPO-D-13-079.1>  
821 Steele, M., Morley, R., Ermold, W., 2001. PHC: A Global Ocean Hydrography with a High-Quality  
822 Arctic Ocean. *J. Clim.* 14, 2079–2087. [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0442(2001)014<2079:PAGOHW>2.0.CO;2)  
823 [0442\(2001\)014<2079:PAGOHW>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<2079:PAGOHW>2.0.CO;2)  
824 Våge, K., Pickart, R.S., Pavlov, V., Lin, P., Torres, D.J., Ingvaldsen, R., Sundfjord, A., Proshutinsky, A.,  
825 2016. The Atlantic Water boundary current in the Nansen Basin: Transport and mechanisms  
826 of lateral exchange. *J. Geophys. Res. Oceans* 121, 6946–6960.  
827 <https://doi.org/10.1002/2016JC011715>  
828 Yeager, S.G., Karspeck, A.R., Danabasoglu, G., 2015. Predicted slowdown in the rate of Atlantic sea  
829 ice loss. *Geophys. Res. Lett.* 42, 10,704–10,713. <https://doi.org/10.1002/2015GL065364>  
830

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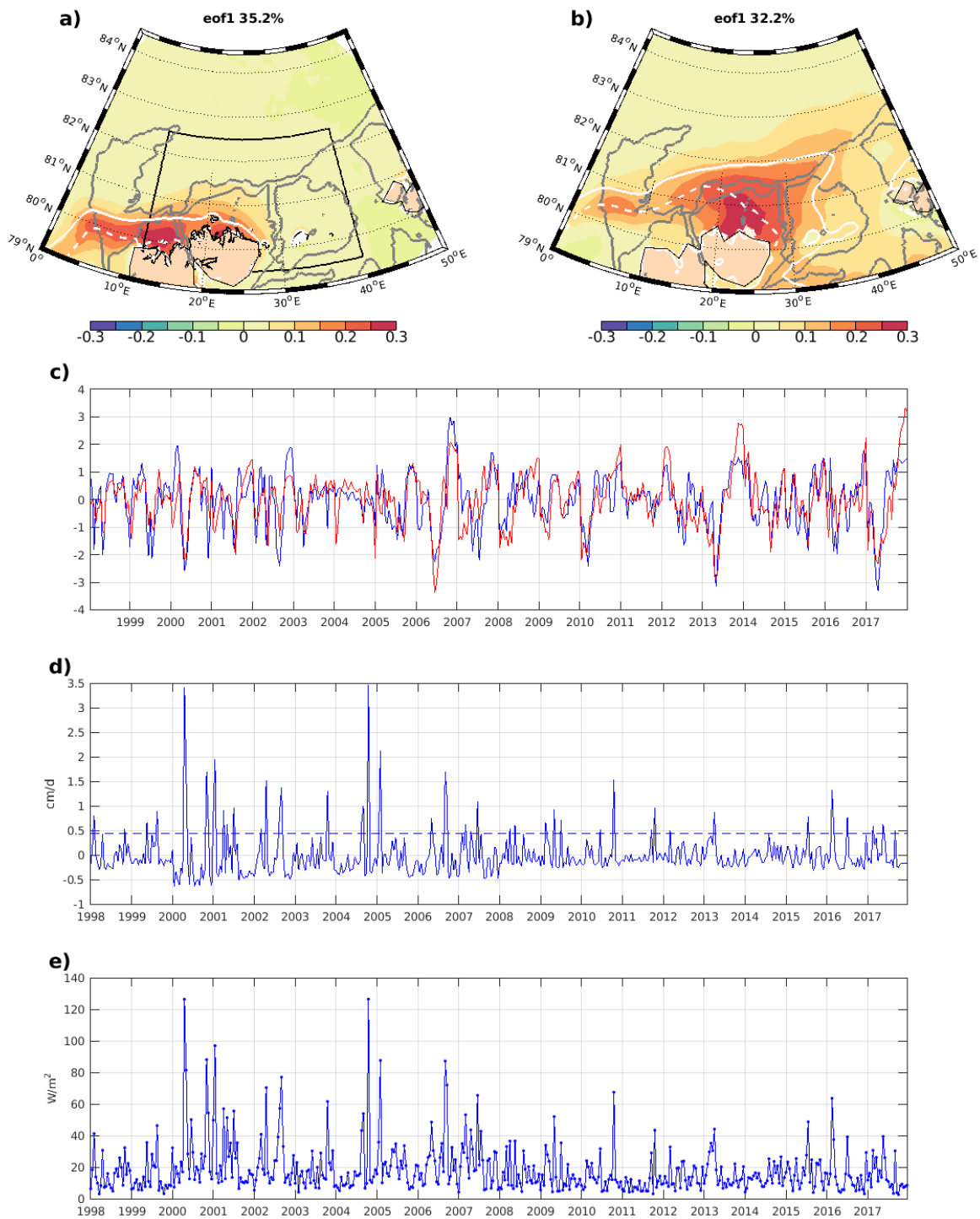


833

834 Figure 1: Simulated winter (December-March) velocity at 60 m depth (arrows) and depth (m)  
835 of the 0°C isotherm (color) averaged over the 1997-2017 period. The red dots indicate the  
836 positions of the moorings which were used for the model-observation comparison (see  
837 supporting information).

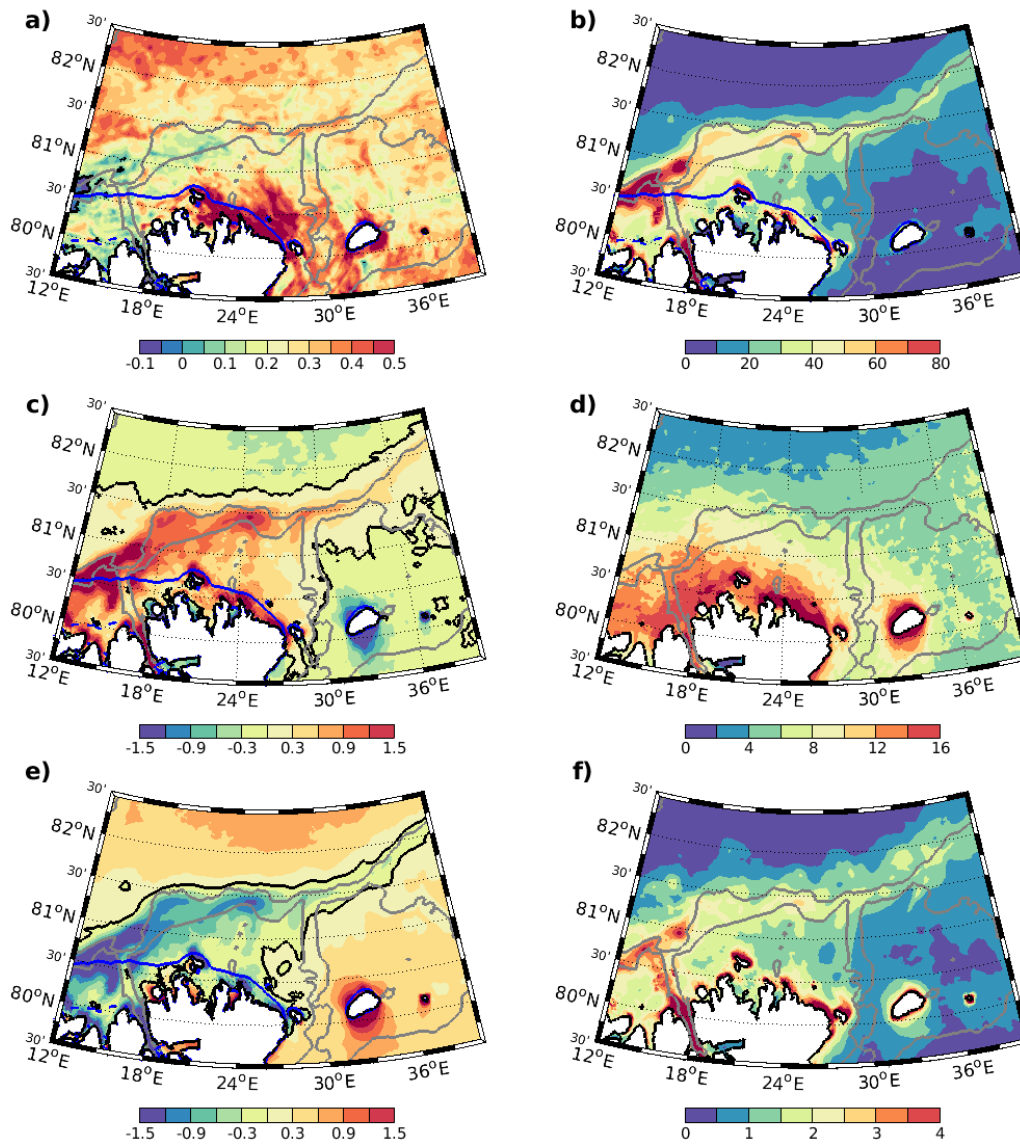
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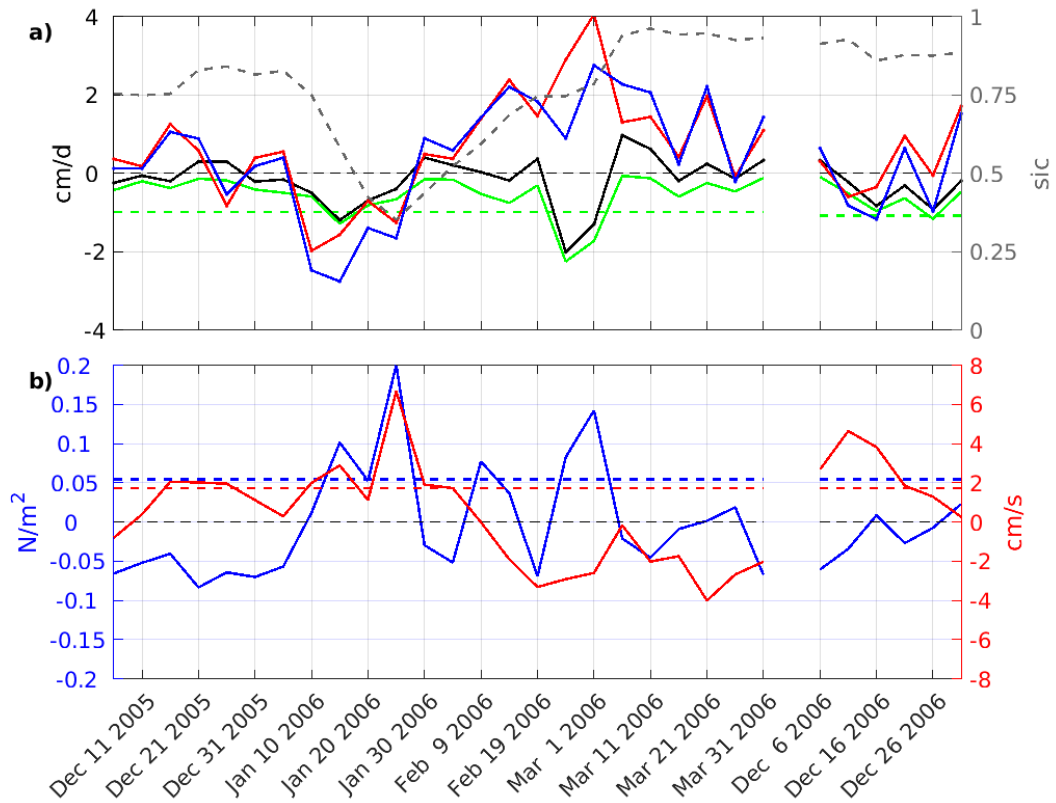
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841 Figure 2: (a)-(b) Loading patterns of the first EOF mode of the winter (DJFM) SIC in (a) the  
 842 model simulation and (b) the observations. (c) Normalized Principal Component associated  
 843 with the first EOF mode of the (blue) simulated and (red) observed winter SIC. SIC anomalies  
 844 are constructed by removing the mean of each winter. (d) Melt index (see definition in text)  
 845 with its standard deviation (dashed line). (e) Ice-ocean heat flux ( $W m^{-2}$ , positive values indicate  
 846 upward fluxes) averaged over the study area (shown as a black box in (a)). Also shown in (a)  
 847 and (b) are the 250 m and 1000 m isobaths (grey), the 50% (dashed white contour) and 80%  
 848 (solid white contour) isopleths of mean winter SIC.



849

850 Figure 3: Left panels: Simulated mean winter (DJFM) ice volume change (cm day<sup>-1</sup>): (a) net  
 851 volume change and the contributions from (c) the ice volume transport convergence and (e) the  
 852 net thermodynamic growth. Right panels: (b) ice-ocean heat flux (W m<sup>-2</sup>, positive values  
 853 indicate upward fluxes), (d) standard deviation of the ice volume transport convergence (cm  
 854 day<sup>-1</sup>) and (f) standard deviation of the net thermodynamic growth (cm day<sup>-1</sup>). Standard  
 855 deviations calculated over all winters of the simulation. Also shown in (a), (b) (c), and (e)  
 856 are the 50% (dashed blue) and 80% (solid blue) isopleths of the mean winter SIC. Grey contours  
 857 represent the 250 m and 1000 m isobaths. The black contour in (c) and (e) represents the 0  
 858 isopleth.

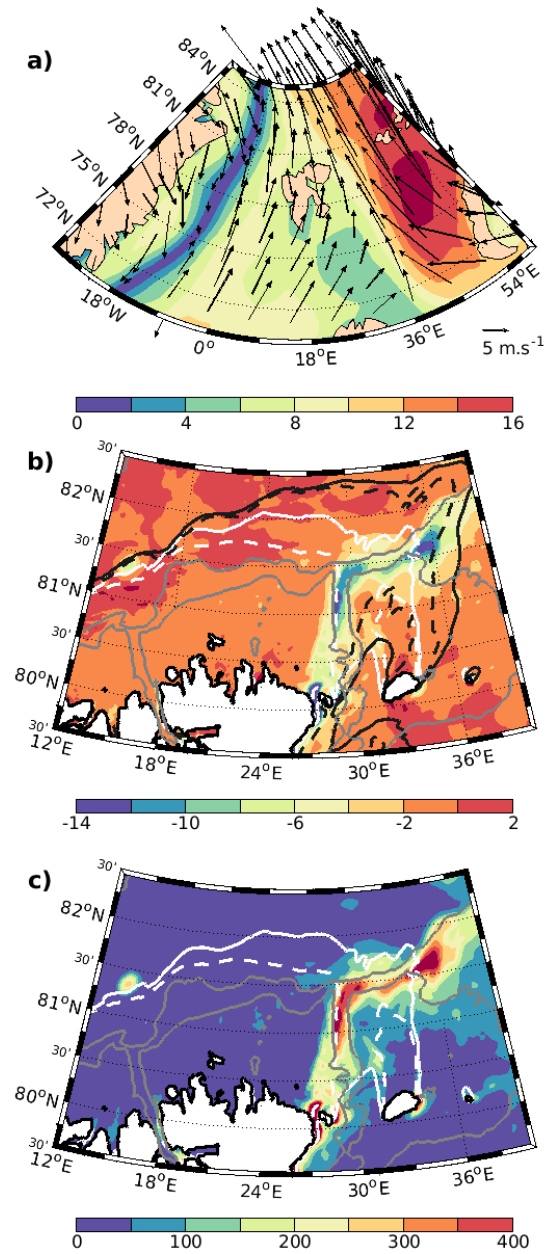


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860 Figure 4: (a) 5-day ice volume change ( $\text{cm day}^{-1}$ ) averaged over the study area over 2005-2006:  
 861 net volume change (blue) and the part contributed by subregions of net ice melt (solid green),  
 862 contributions from the ice volume transport convergence (red) and the thermodynamic ice  
 863 growth (black). Also shown in (a) are, the limit on the mean winter melt rate for identification  
 864 of large melt events (dashed green) and the SIC averaged over the study area (dashed grey). (b)  
 865 Surface stress index (blue) and boundary current index (red), with their limits for identification  
 866 of large stress events (dashed blue) and strong current events (dashed red), respectively (see  
 867 sections 3.4.2.2 and 3.4.2.3 for the definition of the indices).

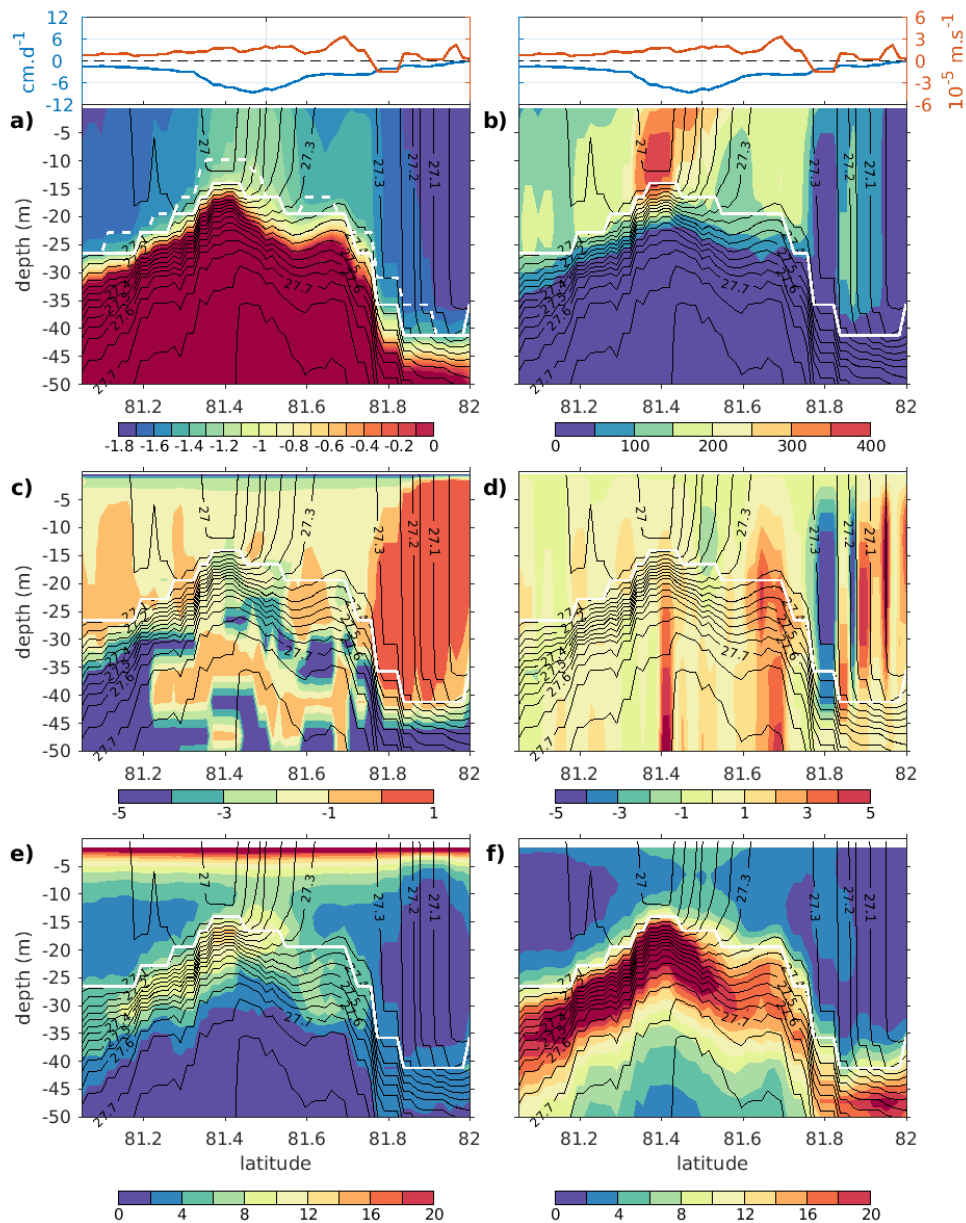
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871 Figure 5: (a) 10 m wind vector (arrows) and wind speed ( $\text{m s}^{-1}$ ), (b) thermodynamic ice  
 872 growth ( $\text{cm day}^{-1}$ ) and (c) ice-ocean heat flux ( $\text{W m}^{-2}$ , positive values indicate upward fluxes)  
 873 on January 15, 2006. The 50% (dashed line) and 80% (solid line) SIC isopleths are also shown  
 874 in (b) and (c) for January 15, 2006 (white) and in (b) for January 20, 2006 (black). Grey contours  
 875 represent the 250 m and 1000 m isobaths.



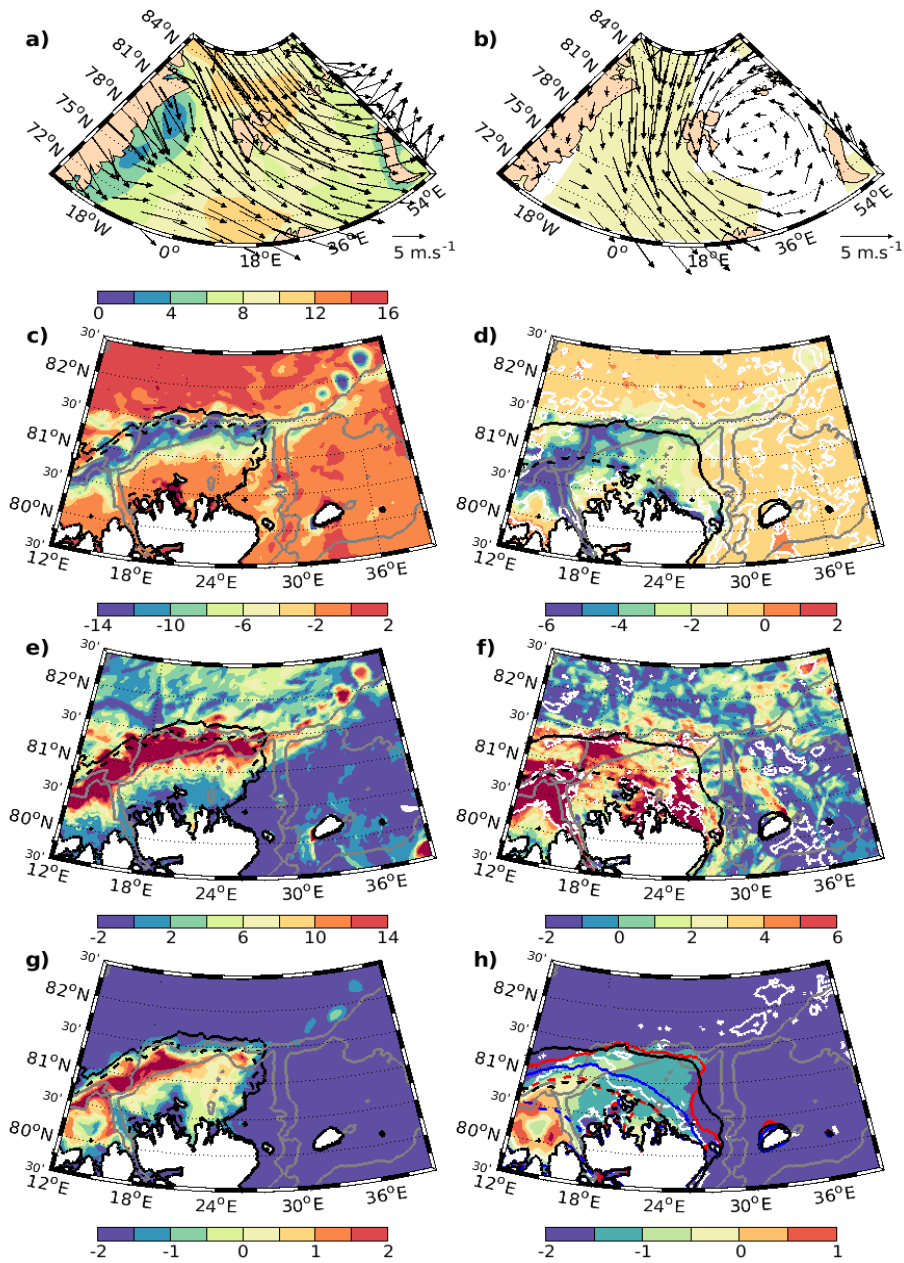
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878 Figure 6: (a) Temperature ( $^{\circ}\text{C}$ ), (b) vertical diffusive heat flux ( $\text{W m}^{-2}$ , positive values indicate  
 879 upward fluxes), (c) vertical diffusion coefficient ( $\text{m}^2 \text{s}^{-1}$ ) expressed as  $\log_{10}(K_v)$ , (d) vertical  
 880 velocity ( $10^{-5} \text{ m s}^{-1}$ ), (e) velocity shear ( $10^{-3} \text{ s}^{-1}$ ), and (f) Brunt Väisälä frequency ( $10^{-3} \text{ s}^{-1}$ ) on a  
 881 meridional section along  $31^{\circ}\text{E}$  on January 15, 2006. On top panels are shown the  
 882 thermodynamic ice growth ( $\text{cm day}^{-1}$ , blue) and the Ekman pumping velocity ( $10^{-5} \text{ m s}^{-1}$ , red).  
 883 The mixed layer depth is shown as a white line in all panels for January 15 (solid) and in (a) for  
 884 January 10 (dashed).

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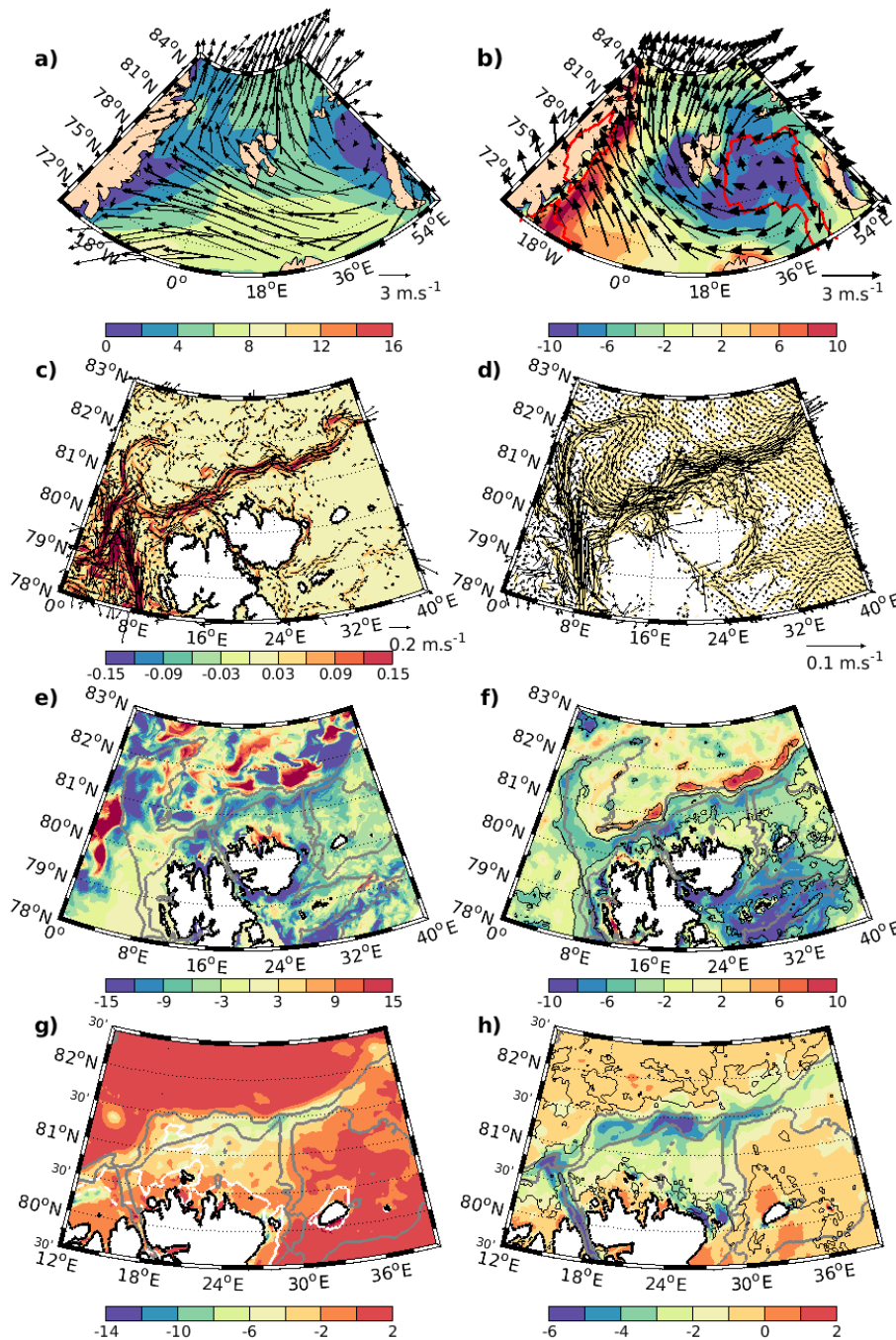
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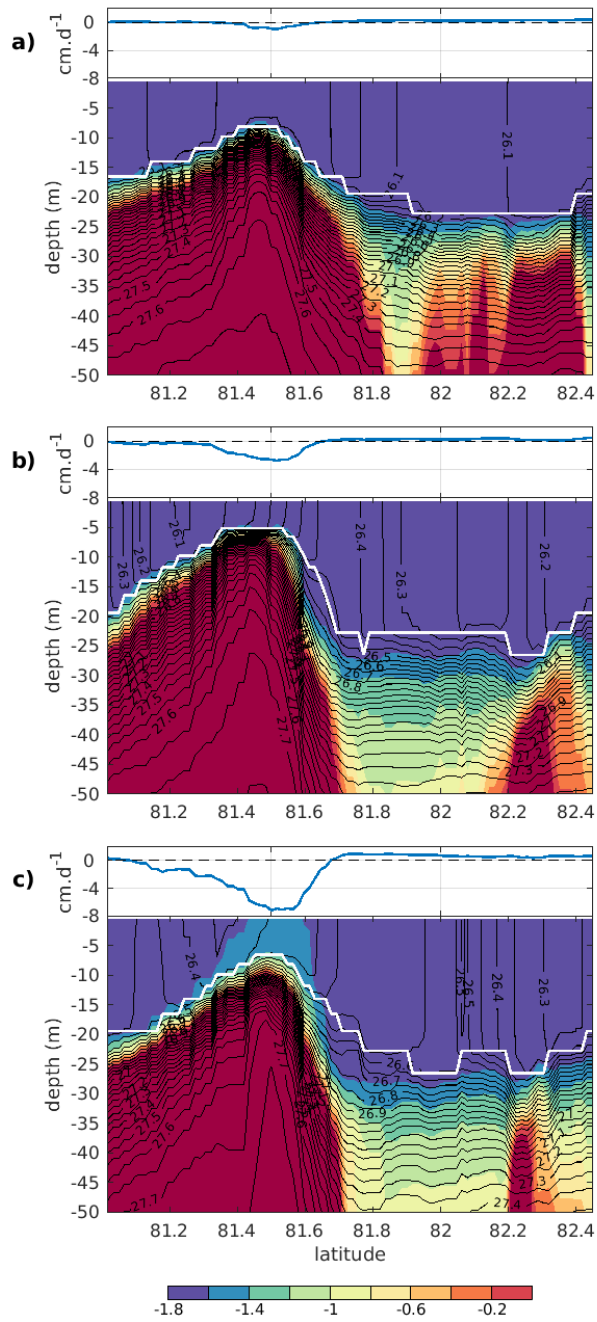
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 888 Figure 7: Left panels: (a) 10 m wind (arrow) and wind speed ( $\text{m s}^{-1}$ ) (color), (c) thermodynamic  
 889 ice growth ( $\text{cm day}^{-1}$ ), and (e) ice volume transport convergence ( $\text{cm day}^{-1}$ ) on February 24,  
 890 2006. (g) SST ( $^{\circ}\text{C}$ ) on February 19, 2006. Also shown in (c)-(g) as black contours are the 50%  
 891 (dashed) and 80% (solid) SIC isopleths on February 24. Right panels: (b, d, f) Composite  
 892 difference between large melt events associated with ice advection over warm SST and other  
 893 events for (b) the 10 m wind, (d) the thermodynamic ice growth ( $\text{cm day}^{-1}$ ), (f) the ice volume  
 894 transport convergence ( $\text{cm day}^{-1}$ ). Also shown in (d) and (f) as black lines are the 50% (dashed)  
 895 and 80% (solid) SIC isopleths averaged over these large melt events. (h) composites of the SST  
 896 ( $^{\circ}\text{C}$ ) five days prior to the large melt events. Also plotted in (h) are the 50% (dashed) and 80%  
 897 (solid) SIC isopleths 5 days before (red), in phase (black) and 5 days after (blue) the large melt  
 898 events. The area of 95% confidence level for the composites is delimited by yellow shading in  
 899 (b) and white lines in (d), (f) and (h). Grey contours represent the 250 m and 1000 m isobaths.





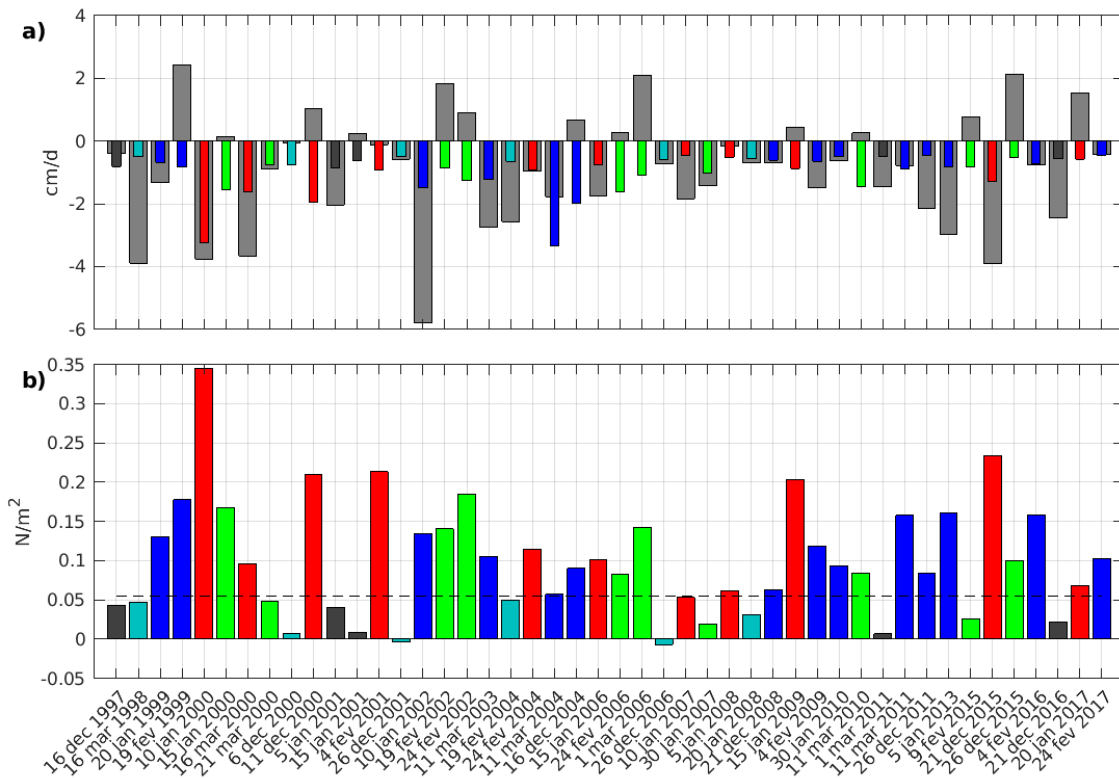
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901 Figure 8: Left panels: (a) 10 m wind (arrows) and wind speed ( $\text{m s}^{-1}$ , color shading) and (g)  
 902 thermodynamic ice growth ( $\text{cm day}^{-1}$ ) on December 26, 2006. (c) current velocity anomaly  
 903 (arrows) and velocity speed anomaly ( $\text{m s}^{-1}$ , color shading) averaged between 20 and 40 meters  
 904 and (e) anomaly of the depth (m) of the  $0^\circ\text{C}$  isotherm, on December 11, 2006. The 50% (dashed  
 905 white line) and 80% (solid white line) SIC isopleths are also shown in (g). Right panels:  
 906 Composites of anomalies of (b) the 10 m wind (arrows) and surface wind curl ( $10^{-6} \text{ s}^{-1}$ , color  
 907 shading), (d) the current velocity averaged between 20 and 40 m and (f) the depth (m) of the  
 908  $0^\circ\text{C}$  isotherm, for the strong current events, (h) composites of anomalies of the thermodynamic  
 909 ice growth ( $\text{cm day}^{-1}$ ) for large melt events associated with strong current. All anomalies are  
 910 calculated relative to the winter average.



911

912 Figure 9: Temperature ( $^{\circ}\text{C}$ , color shading) and density (sigma units, black contours) overlaid  
 913 with the mixed layer depth (white line) on a meridional section along  $31^{\circ}\text{E}$  on (a) December 1,  
 914 (b) December 16 and (c) December 26, 2006. On top of each panel, the thermodynamic ice  
 915 growth rate along the section is also plotted.

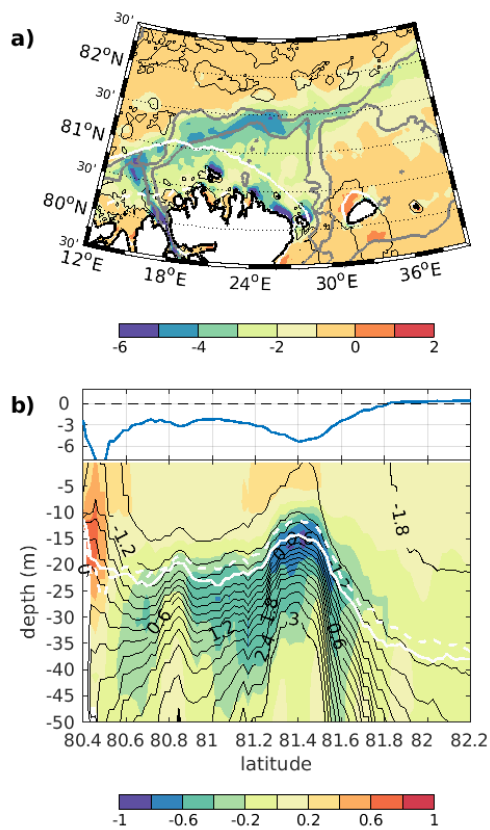


917

918 Figure 10: (a) ice volume change ( $\text{cm day}^{-1}$ ) averaged over the study area for all the large melt  
 919 events (grey bar), overlaid with the ice melt contribution (narrow bar) for those large melt  
 920 events only associated with large stress (blue), ice advection over warm SST (green), large  
 921 stress and strong current (red), and large current (cyan). Large melt events which are not  
 922 associated with any specific process are shown in black. (b) surface stress ( $\text{N m}^{-2}$ ) averaged  
 923 over the study area with same color code for the different types of large melt events defined in  
 924 (a), and its standard deviation (dashed line).

925

926



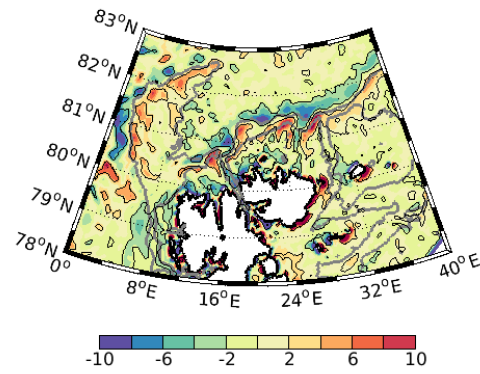
928

929 Figure 11: (a) Composite difference of the thermodynamic ice growth (cm day<sup>-1</sup>) between the  
 930 large melt events associated with large stress and other melt events. (b) Composites of the  
 931 temperature (°C, black contours) for the large melt events associated with large stress and the  
 932 difference between the in-phase temperature and the temperature 5 days prior to these events  
 933 (°C, color shading) on a meridional section along 24°E. Also shown as white lines in (a) are the  
 934 50% (dashed) and 80% (solid) isopleths of the SIC averaged over these large melt events and  
 935 in (b) is the mixed layer depth in phase with (solid) and five days prior to (dashed) the melt  
 936 event. On top of (b), the thermodynamic ice growth rate (cm day<sup>-1</sup>) is shown along the section.

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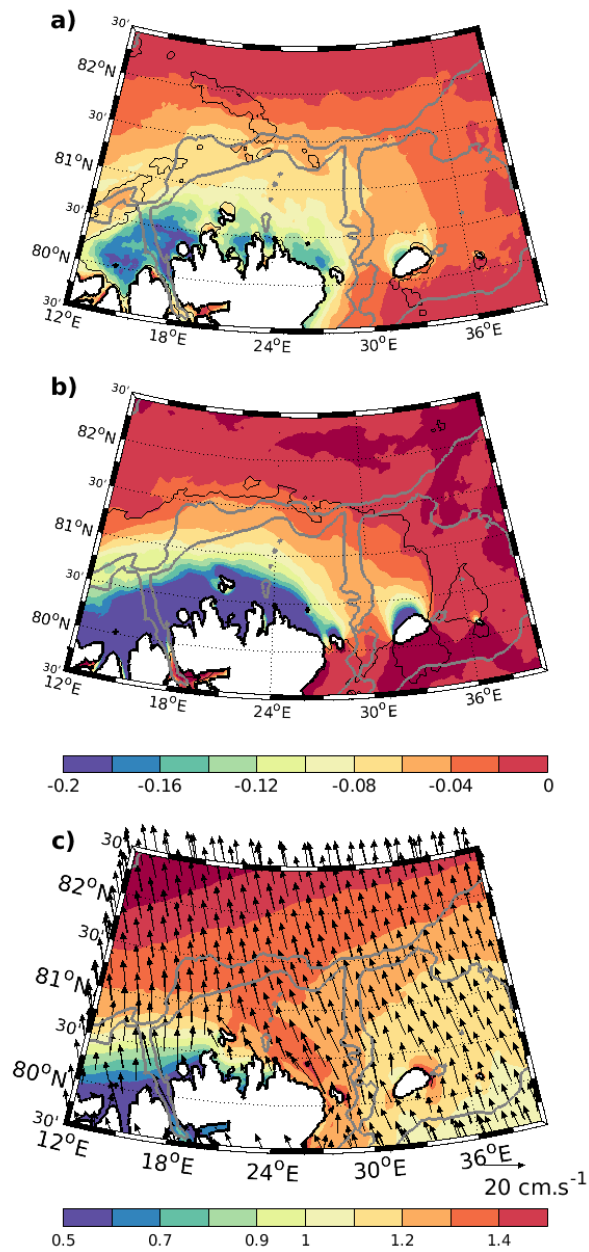


940

941 Figure 12: Composites of anomalies of the Ekman suction velocity ( $10^{-6} \text{ m s}^{-1}$ ) for the strong  
942 current events. Anomalies are calculated relative to the winter average.

943

944



945

946 Figure 13: Composites of the SIC change between two consecutive 5-day periods for (a) the  
 947 large melt events and (b) the events of large ice volume transport divergence. (c) Composites  
 948 of the ice drift anomalies (arrows) superimposed on the mean winter ice thickness (m, color  
 949 shading) for events of large ice volume transport divergence. In (a) and (b) the thin black line  
 950 represents the 95% confidence level for the composites. Grey contours represent the 250 m and  
 951 1000 m isobaths.