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1 **Greenland freshwater pathways in the sub-Arctic Seas from model experiments with**  
2 **passive tracers**

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31           **Abstract**

32           Accelerating since the early 1990s, the Greenland Ice Sheet mass loss exerts a  
33 significant impact on thermohaline processes in the sub-Arctic seas. Surplus freshwater  
34 discharge from Greenland since the 1990s, comparable in volume to the amount of freshwater  
35 present during the Great Salinity Anomaly events, could spread and accumulate in the sub-  
36 Arctic seas and influence convective processes there. However, hydrographic observations in  
37 the Labrador Sea and the Nordic Seas, where the Greenland freshening signal might be  
38 expected to propagate, do not show a persistent freshening in the upper ocean during last two  
39 decades. This raises the question of where the surplus Greenland freshwater has propagated.  
40 In order to investigate the fate, pathways, and propagation rate of Greenland freshwater in the  
41 sub-Arctic seas, several numerical experiments using a passive tracer to track the spreading  
42 of Greenland freshwater have been conducted as a part of the Forum for Arctic Ocean  
43 Modeling and Observational Synthesis effort. The models show that Greenland freshwater  
44 propagates and accumulates in the sub-Arctic seas, although there is disagreement among the  
45 models on the amount and rate of tracer propagation into the convective regions. Results  
46 highlight the differences in simulated physical mechanisms at play in different models and  
47 underscore the continued importance of intercomparison studies. It is estimated that surplus  
48 Greenland freshwater flux should have caused a salinity decrease by 0.06—0.08 in the sub-  
49 Arctic seas, which is smaller than the recently observed salinization (0.15–0.2) of Atlantic  
50 Water that has possibly obscured the freshening signal.

51           **Key words:** Greenland Ice Sheet melting, Greenland freshwater, thermohaline  
52 circulation, Nordic Seas, sub-Arctic seas, Baffin Bay, Labrador Sea

53           **3 key points:**

54           Results from passive tracer model experiments are presented  
55           Greenland freshwater pathways and accumulation in the sub-Arctic seas are analyzed  
56           Meltwater influence on salinity change in the sub-Arctic is discussed

57

57

## 58 **1. Introduction**

59           Observational and modeling studies indicate that the Greenland Ice Sheet and other  
60 Arctic land ice are melting at a rate that has dramatically increased since the early 1990s  
61 [Pritchard *et al.*, 2009; Velicogna, 2009; Gardner *et al.*, 2011; Bamber *et al.*, 2012; Box and  
62 Colgan, 2013]. The volume of surplus freshwater discharge from Greenland into the sub-  
63 Arctic seas (Greenland, Iceland, Norwegian, Labrador seas, Baffin Bay, and the Subpolar  
64 region in the North Atlantic, Figure 1) since 1990 (in 2010  $\sim 3,200 \text{ km}^3$ ) is approaching the  
65 magnitude of the freshwater anomaly ( $10,000 \text{ km}^3$ ) advected into the North Atlantic during  
66 the 1970 Great Salinity Anomaly event (GSA) [Dickson *et al.*, 1998; Bamber *et al.*, 2012],  
67 suggesting that this freshwater source may have significant impact on ocean conditions in the  
68 region.

69           The sensitivity of the thermohaline circulation in the North Atlantic to freshwater  
70 balance has been widely discussed in the literature [Stommel, 1961; Rooth, 1982; Manabe  
71 and Stouffer, 1988; Rahmstorf, 1995; Malmberg and Jonsson, 1997; Stouffer *et al.*, 2006;  
72 Jahn and Holland, 2013]. Considerable progress has been made in understanding the role of  
73 the thermohaline circulation in abrupt climate change from the paleoclimate perspective [e.g.,  
74 Manabe and Stouffer, 1993; 1995; Clark *et al.*, 2002], thus providing a link between the  
75 freshwater budget of the sub-Arctic and climate variability. The concept of freshwater  
76 influence on the thermohaline circulation has been generally accepted and used to explain  
77 observed climate variability in the Arctic and North Atlantic [Stouffer *et al.*, 2015 and  
78 references therein]. The Arctic Ocean and the Greenland Ice Sheet are two sources of  
79 freshwater that potentially may impact the thermohaline circulation in the North Atlantic and  
80 influence the Atlantic Meridional Overturning Circulation (AMOC).

81 Past GSA events resulting from anomalously high freshwater flux from the Arctic  
82 Ocean have been shown to have impacted thermohaline conditions and climate in the North  
83 Atlantic in the 20<sup>th</sup> century [*Dickson et al.*, 1988; *Belkin*, 2004; *Belkin et al.*, 2008]. Further,  
84 conceptual and idealized models have put forward mechanisms by which a freshwater flux  
85 from the Arctic Ocean may be a key factor in controlling the thermohaline circulation in the  
86 North Atlantic [e.g., *Ikeda*, 1990; *Mysak and Venegas*, 1998; *Ikeda et al.*, 2001; *Goosse et al.*,  
87 2002; *Proshutinsky et al.*, 2002; *Dukhovskoy et al.*, 2004].

88 Meltwater from the Greenland Ice Sheet is the other major source of freshwater  
89 considered to influence thermohaline circulation and climate [*Rahmstorf*, 2002; *Fichefet et*  
90 *al.*, 2003; *Rahmstorf*, 2005; *Ridley et al.*, 2005; *Rudels*, 2011; *Castro de la Guardia et al.*,  
91 2015]. With growing evidence of accelerated Greenland ice melt [*Velicogna*, 2009; *Bamber*  
92 *et al.*, 2012; *Box and Colgan*, 2013], increasing attention has been given to the role of  
93 meltwater in the high-latitude climate [*Hu et al.*, 2011; *Weijer et al.*, 2012; *Rahmstorf et al.*,  
94 2015]. *Proshutinsky et al.* [2015] conjecture that the recent observed disruption of quasi-  
95 decadal oscillations of the Arctic wind-driven circulation regimes may be attributed to a  
96 surplus freshwater flux from Greenland. The underlying assumption is that a substantial  
97 volume of Greenland freshwater spreads laterally from the Greenland coast into convective  
98 sites of the interior sub-Arctic seas. On the other hand, observational studies suggest that  
99 freshwater predominantly flows in boundary currents along the margins of the Nordic Seas  
100 (the East Greenland Current) and the Labrador Sea (the West Greenland Current and the  
101 Labrador Current) [e.g., *Bacon et al.*, 2002; *Dickson et al.*, 2007; *Sutherland and Pickart*,  
102 2008; *Myers et al.*, 2009]. Hence, most of the Greenland freshwater discharge should travel  
103 along the margins of the sub-Arctic seas on its way to the North Atlantic (Figure 1). It  
104 remains unclear how, where, and on what timescales this freshwater can impact convective  
105 regions [*Moore et al.*, 2015]. The rate of Greenland freshwater flux into the interior seas,

106 which is unknown, is also important. Previous model experiments have shown that the  
107 thermohaline circulation in the North Atlantic responds to freshwater perturbations on order  
108 of 0.1 Sv [*Rahmstorf, 1995; Fanning and Weaver, 1997; Clark et al., 2002*].

109 In order to understand the mechanisms of climate change driven by changes in  
110 AMOC in response to increased ice sheet melting, a number of “hosing” experiments have  
111 been performed with Ocean General Circulation Models and climate models [*Huybrechts et*  
112 *al., 2002; Fichefet et al., 2003; Ridley et al., 2005; Jungclaus et al., 2006; Gerdes et al.,*  
113 *2006; Stouffer et al., 2006; Swingedouw et al., 2007; Vizcaino et al., 2008; Hu et al., 2011*].  
114 Overall, these experiments show different AMOC states as a function of freshwater flux into  
115 the North Atlantic. These numerical experiments are not designed to address any of the  
116 questions raised above. Typically in these hosing experiments surplus freshwater flux is  
117 imposed over some region of the North Atlantic. Although such simplification of freshwater  
118 input into the North Atlantic seems to have low impact on climate simulations [*Kleinen et al.,*  
119 *2009*], misrepresentation of actual pathways of Greenland freshwater into the ocean may be  
120 important for the realistic responses of thermohaline processes in the sub-Arctic seas, as  
121 discussed in *Stammer [2008], Marsh et al. [2010], and Weijer et al. [2012]*.

122 In order to investigate the influence of surplus Greenland Ice Sheet meltwater on the  
123 sub-Arctic seas and North Atlantic several high-resolution ocean experiments have been  
124 conducted. A model study of *Marsh et al. [2010]* includes 0.25° and 1° NEMO simulations  
125 integrated for 8 years. They found that the freshwater signal tends to stay along the narrow  
126 boundary current of the Labrador Sea and only a very small volume of freshwater has  
127 reached the interior Subpolar Gyre in the simulation after 8 years. The largest volume of  
128 surplus freshwater accumulated in Baffin Bay. The authors mention that the East Greenland  
129 Coastal Current, which acts as a freshwater conduit, is not resolved in the NEMO

130 experiments. *Bacon et al.* [2014] conclude that at least 1/12 degree model resolution is  
131 needed to resolve freshwater fluxes associated with East Greenland Coastal Current.

132 Another recent model study of *Weijer et al.* [2012] compares the adjustment of the  
133 AMOC to the surplus of Greenland freshwater in multidecadal simulations with 0.1°  
134 (“strongly eddying”) and 1° (“non-eddying”) global configurations from the Los Alamos  
135 Parallel Ocean Program (POP). In these experiments, a passive tracer is released along with  
136 the freshwater anomaly around Greenland in order to diagnose the freshwater propagation  
137 rate. The numerical experiments predict rapid spreading of the tracer at shallow depths over  
138 the sub-Arctic seas (within a year) with a distinct delay (about a year) in the arrival time for  
139 the tracer in the interior Nordic Seas. Both POP configurations predict decline in convective  
140 activity in both the Labrador Sea and Nordic Seas. The strongest freshening is simulated in  
141 the Labrador Sea and Baffin Bay, whereas the decline in surface salinity in the Nordic Seas is  
142 much weaker.

143 The goal of this paper is to investigate the pathways, mechanisms and time scales of  
144 Greenland meltwater propagation within the sub-Arctic seas with particular focus on the  
145 Nordic Seas, the Labrador Sea, and Baffin Bay (Figure 1). We start our analysis with a  
146 description of the Greenland Ice Sheet runoff data used in the model experiments (section 2)  
147 and a brief overview of observed salinity changes in the sub-Arctic seas that could provide  
148 evidence of the influence of surplus Greenland freshwater influence (section 3). Design and  
149 analysis of numerical experiments employing three regional coupled ocean–sea ice models  
150 with different resolution forced by realistic Greenland freshwater fluxes are discussed in  
151 section 4. The experiments use a passive tracer released at exact locations of the Greenland  
152 freshwater sources. The model results are analyzed from the perspective of observed salinity  
153 changes in the sub-Arctic seas (section 5).

154

## 155 **2. Greenland Freshwater Flux**

### 156 **2.1. Data**

157 Greenland freshwater fluxes, between 1958 and 2010, from *Bamber et al.* [2012] are  
158 employed to investigate the fate of this freshwater and assess its role in thermohaline  
159 variability over the past decade. The data are a monthly gridded product (5x5km grid) with  
160 realistic geographic distribution and temporal variability (Figure 2a). Greenland runoff was  
161 derived from a reconstruction of the surface mass balance of the Greenland Ice Sheet and  
162 surrounding tundra using a high-resolution regional climate model, RACMO2 [*Ettema et al.*,  
163 2009] forced with ERA-40 re-analysis data. The runoff from the ice sheet and surrounding  
164 tundra was combined with observations of solid ice discharge derived from satellite  
165 observations of ice velocity to produce the total freshwater flux.

166

### 167 **2.2. Variability and Trends, 1990-2010**

168 According to *Bamber et al.* [2012], total Greenland freshwater flux has been  
169 predominantly  $>1000 \text{ km}^3 \text{ yr}^{-1}$  since 1998 having increased from  $870\text{--}900 \text{ km}^3 \text{ yr}^{-1}$  in the  
170 early 1990s to  $1100\text{--}1200 \text{ km}^3 \text{ yr}^{-1}$  in the late 2000s (Figure 2b). A somewhat smaller  
171 estimate of the total Greenland freshwater flux is reported in *Box and Colgan* [2013], but it  
172 did not include tundra runoff that accounts for an additional  $100\text{--}200 \text{ km}^3 \text{ yr}^{-1}$ . However, their  
173 estimate however indicates a similar increase of the Greenland freshwater flux of around  $200$   
174  $\text{km}^3 \text{ yr}^{-1}$  between 1990 and 2010.

175 The volume of freshwater supplied by the Greenland Ice Sheet [*Bamber et al.*, 2012]  
176 is roughly 29% - 42% of the total annual river runoff into the Arctic Ocean ( $3500 \text{ km}^3 \text{ yr}^{-1}$   
177 according to *Curry and Mauritzen* [2005] or  $2500 \text{ km}^3 \text{ yr}^{-1}$  as reported by *Aagaard and*  
178 *Carmack* [1989] based on climatological data of the 1980s). The change of the freshwater  
179 flux between 1990 and 2010 is equivalent to a freshwater input increase of 6.3 mSv to 10.5



180 mSv. This increase is smaller than those of an earlier study of *Gregory and Lowe* [2000]  
181 noting a 17.4 mSv increase. In addition, between 2004 and 2009 mass loss from the Canadian  
182 Arctic Archipelago (CAA), primarily into Baffin Bay, changed from 31 to 92 km<sup>3</sup> yr<sup>-1</sup>  
183 [*Gardner et al*, 2011], an increase from about 1 to 3 mSv.

184         Estimated from the Greenland freshwater data of *Bamber et al.* [2012], the cumulative  
185 Greenland freshwater flux anomaly between 1990-2010 is 2830±337 km<sup>3</sup> relative to the pre-  
186 1990 mean flux of 876 km<sup>3</sup> yr<sup>-1</sup>. This amount is 28% of the freshwater volume advected to  
187 the sub-Arctic from the Arctic Ocean during the 1970s GSA event, totaling 10,000 km<sup>3</sup>  
188 [*Dickson et al.*, 1988]. If the Greenland freshwater flux increases at the same rate for the next  
189 50 years, the surplus freshwater will reach the GSA freshwater volume by the mid 2060s. It is  
190 noteworthy that rates of mass loss from Greenland have been steadily increasing since 2010  
191 [*Helm et al*, 2014].

192

### 193 **3. Observed Salinity Changes in the Sub-Arctic Seas**

194         Recent changes in salinity fields in the sub-Arctic seas are discussed below in an  
195 attempt to identify observational evidence of freshwater release from Greenland that has been  
196 increasing since the early 1990s.

#### 197 **3.1. Baffin Bay**

198         Hydrographic characteristics of Baffin Bay are largely determined by outflow from  
199 the Arctic Ocean and inflow from the Labrador Sea that enters the bay as the West Greenland  
200 Current (WGC, Figure 1). Greenland freshwater flux to Baffin Bay has substantially  
201 increased since the early 1990s (200—250 km<sup>3</sup> yr<sup>-1</sup>) contributing about 316 km<sup>3</sup> yr<sup>-1</sup> in the  
202 late 2000s (Figure 2b). This is ~13% of the freshwater transport through the northern  
203 channels of the Canadian Arctic Archipelago (CAA) estimated as 2450 km<sup>3</sup> yr<sup>-1</sup> combining  
204 estimates in *Peterson et al.* [2012], *Münchow and Melling* [2008], *Rabe et al.* [2012], *Agnew*  
205 *et al.* [2008], and *Kwok* [2005].

206 A detailed analysis of warming and freshening in Baffin Bay over 1916-2003 based  
207 on historical hydrographic data was conducted by *Zweng and Munchow* [2006]. The majority  
208 of the observations used were collected between 1950 and the early 2000s. The study  
209 revealed significant freshening over the shelf and slope regions of Baffin Bay in the layer  
210 between 50–200 m depth range. The largest freshening ( $-0.086 \pm 0.039$  decade<sup>-1</sup>) was found on  
211 the Baffin Island shelf. A smaller freshening (from  $-0.048$  to  $-0.066$  decade<sup>-1</sup>) was reported  
212 for the continental slopes off Baffin Island. The surface waters of the Greenland shelf and  
213 slope regions to the north of Davis Strait had similar freshening trends ( $-0.04$  decade<sup>-1</sup>). The  
214 authors suggest Greenland meltwater runoff might have contributed to this freshening.  
215 Recently *Myers and Ribergaard* [2013] also found freshening in the upper layer of Disko  
216 Bay, as well as offshore in the West Greenland Current in the 2000s as compared to the  
217 1980s and 1990s.

218 Moored observations in Davis Strait described in *Curry et al.* [2014] indicate slight  
219 freshening of several water mass classes (Figures 3a-c). Although no statistical significance  
220 can be drawn from these short time series, all but two water masses (West Greenland  
221 Irminger Water, WGIW and Transitional Water, TrW) tend to have negative salinity  
222 anomalies in recent years relative to the 2003-2013 mean (Figures 3b and c). UK Met Office  
223 Hadley Centre objectively mapped monthly salinity data (version EN.4.1.1) [*Good et al.*,  
224 2013] have been examined in the region (Figure 3d). Salinity anomalies relative to the 1960-  
225 1990 monthly climatology indicate apparent freshening signal in the upper 50 m over 1990 to  
226 2014 in BB. Over the past decade, fresh surface water has expanded southwards (by about  
227 10° latitude) into the northern Labrador Sea.

228

### 229 **3.2. The Labrador Sea**

230           The Labrador Sea receives freshwater exported from the Arctic Ocean through the  
231 CAA via Davis Strait and Hudson Strait and through Fram Strait with the EGC and WGC  
232 (Figure 1). Additionally, the Labrador Sea also receives salty modified Atlantic Water  
233 originating from the North Atlantic Current (called the Subpolar Mode Water in [McCartney  
234 and Talley, 1982]). The warm and salty Atlantic water flows along the continental slope  
235 around the basin as the Irminger Current. It is generally accepted that eddies shed by the  
236 Irminger Current control the heat and salt budget of the interior Labrador Sea [e.g., Straneo,  
237 2006], governing winter convective activity when the intermediate and deep waters of the  
238 North Atlantic are formed [Lazier et al., 2002]. Deep convection controls formation and  
239 properties of Labrador Sea Water that occupies the central Labrador Sea [Talley and  
240 McCartney, 1982]. Deep winter convection in 1987—1994 produced an anomalously high  
241 volume of the deepest and densest Labrador Sea Water [Yashayaev, 2007]. After 1994, the  
242 freshening of the water column reversed, especially in the layers below 1000 m attributed to  
243 weaker convection in the central Labrador Sea with exception to the deep convection events  
244 during winter of 2007-2008 [Yashayaev and Loder, 2009].

245           Hydrographic observations in the central Labrador Sea indicate considerable  
246 variability of the temperature and salinity characteristics of the Labrador Sea Water over the  
247 last four decades [Yashayaev, 2007]. Oceanographic monitoring of the Labrador Sea was  
248 mostly done on the basis of annual hydrographic sections across the central Labrador Sea (for  
249 more details, see Yashayaev [2007]) until 2002 when the international Argo program  
250 (<http://www.argo.ucsd.edu>) was launched. In the early 1970s, the Labrador Sea had a  
251 noticeable freshening trend that continued until 1995 [Dickson et al., 2002; Zweng and  
252 Munchow, 2006; Yashayaev and Seidov, 2015]. During 1988 through 1994, strong freshening  
253 was observed in the upper 1500 m in the Labrador Sea caused by the formation of a large  
254 volume of cold and fresh Labrador Sea Water [Sy et al., 1997; Lazier et al., 2002]. After

255 1995, the salinity trend changed to positive and remained positive at least through the late  
256 2000s. *Yashayaev and Loder* [2009] analyzed time evolution of hydrography fields in the  
257 central Labrador Sea on the basis of the Argo float data and in situ observations. Their results  
258 demonstrated a strong positive salinity anomaly developed in the 200-500 m layer during  
259 2000-2009 with the highest salinity >34.9 in 2008 (compared to 34.78-34.82 in the late  
260 1980s). Recent Argo observations in the Labrador Sea show a slight decrease in salinity in  
261 the 200-800 m layer after 2012 caused by several deep convection events, with the strongest  
262 convection occurred in 2014 [*Kieke and Yashayaev*, 2015]. Freshening in the upper 200m has  
263 been observed in the western Subpolar Gyre region and the Labrador Sea since 2010  
264 [*Beszczyńska-Møller and Dye*, 2013; *Yashayaev et al.*, 2015].

265

### 266 **3.3. The Nordic Seas**

267 Hydrographic characteristics of the upper ocean water masses in the Nordic Seas are  
268 influenced by the northward flowing Atlantic Water and southward flowing Polar Water from  
269 the Arctic Ocean carried by the EGC [*Aagaard et al.*, 1985]. Whereas the eastern and western  
270 parts of the Nordic seas are largely determined by the characteristics of the Polar Water and  
271 Atlantic Water, respectively, the interior part, that includes convective regions in the Iceland  
272 and Greenland seas, has a more complex relationship between hydrography and contributions  
273 from Polar Water and Atlantic Water. The Greenland and Iceland seas have many features in  
274 common [*Malmberg and Jonsson*, 1997]. Hydrography of the upper layer in both seas is  
275 determined by water mass transformations and modifications through lateral advection of  
276 Polar Water and Atlantic Water from the rims of the seas and convection driven by surface  
277 cooling and salt flux during ice formation. The North Atlantic water undergoes substantial  
278 cooling and freshening as it mixes with the ambient water before entering the Greenland

279 Gyre. Upon mixing, the two water masses form the upper Greenland and Iceland water mass,  
280 sometimes called Arctic Surface Water [*Carmack, 1990*].

281 Large salinity changes have been observed in the Nordic Seas since the 1950s. A  
282 long-term freshening of the North Atlantic in surface, intermediate and deep water masses  
283 has been reported in *Curry et al. [2003]*, *Dickson et al. [2003]*, and *Curry and Mauritzen*  
284 *[2005]*. Hydrographic observations conducted in the Greenland and Iceland seas (regions  
285 “GS” and “IS” in Figure 1) reveal strong negative salinity anomalies ( $<34.3$  to  $34.5$ ) in the  
286 upper 25-m layer during the late 1960s-1970s, late 1980s, and the late 1990s (Figure 4). The  
287 anomalies are related to the 1970s, 1980s and 1990s GSAs, respectively [*Dickson et al.*,  
288 *1988*; *Belkin et al., 1998*; *Belkin, 2004*]. In the mid 2000s, high salinity ( $>34.7$ ) was observed  
289 in both seas in the layer from 0 to 200m. The positive salinity event was replaced by a strong  
290 freshening signal ( $<34.5$ ) in 2008 in the upper Greenland Sea (above 50 m) but not in the  
291 Iceland Sea (Figure 4). A persistent salinity increase in the layers below 50–75 m was  
292 observed in the regions, especially apparent in the Greenland Sea. The observations in Figure  
293 4 agree with the previous studies that reported a positive salinity trend in the Nordic Seas  
294 since the late 1990s [*Avsic et al., 2006*; *Hátún et al., 2005*; *Holliday et al., 2008*; *Somavilla et*  
295 *al., 2013*]. Recent analysis of Argo float observations in the Nordic Seas, between 2001-  
296 2007, shows a slightly positive salinity trend for the upper 500 m [*Latarius and Quadfasel,*  
297 *2010*]. Increased salinity in the upper Nordic Seas is related to increased salinity in the  
298 Atlantic water flowing into the sub-Arctic seas [*Walczowski and Piechura, 2006*; *Holliday et*  
299 *al., 2008*; *Yashayaev and Seidov, 2015*], likely related to increased salinity in subtropical and  
300 mid-latitudes Atlantic Water [*Hátún et al., 2005*].

301 To summarize, historic observations in Baffin Bay show persistent freshening over  
302 the last two decades, whereas salinity in the other sub-Arctic seas has both positive and  
303 negative anomalies. Thus, no definite conclusion can be made on the linkage between

304 observed salinity anomalies in the sub-Arctic seas and increased Greenland freshwater  
305 discharge observed during last decades. We believe that more information and additional  
306 studies are needed to reveal mechanisms and processes regulating freshwater fluxes and fresh  
307 water transformations under different complex processes. A better knowledge about  
308 freshwater pathways, accumulation rates and time scales of freshwater propagation is  
309 necessary in order to understand the disagreement between proposed freshening of the sub-  
310 Arctic seas caused by surplus Greenland freshwater flux and the observed salinity changes.  
311 Note that observational studies referenced in this section were primarily focused on  
312 hydrographic changes in the water column below the near-surface layer (upper 50–200 m).  
313 However, results from model experiments with passive tracers described in the following  
314 sections suggest that Greenland freshening should primarily manifest in the in the upper 200  
315 m and mostly in the 0–50 m layer.

316

#### 317 **4. Results from the Passive Tracer Experiments**

318 Three groups participating in the Forum for Arctic Ocean Modeling and  
319 Observational Synthesis (FAMOS) project have run ocean-sea ice models with passive  
320 tracers released at the Greenland freshwater sources. The models used in this study are from  
321 the Florida State University (AO-HYCOM), the University of Alberta (NEMO-LIM2), and  
322 the Institute of Computational Mathematics and Mathematical Geophysics (ICMMG). The  
323 simulations have different model forcing and configurations. The length of integration is 14  
324 years. The models have different resolutions; AO-HYCOM has a high-resolution ( $0.08^\circ$ )  
325 configuration and NEMO-LIM2 and ICMMG have  $0.25^\circ$  and  $0.5^\circ$  grid spacing, respectively.  
326 The only coordinated forcing is the Greenland freshwater flux data (described in section 2.1).  
327 In the model experiments, Greenland freshwater is tracked by a passive tracer released at the  
328 locations of Greenland freshwater sources (Figure 2a). The tracer is treated as a scalar

329 variable and its time evolution is described by scalar advection and diffusion equations  
330 similar to temperature and salinity. Specifics of the models and experiments are summarized  
331 in Appendix 1.

332

#### 333 ***4.1. Evolution of Tracer Concentration in the Upper Ocean***

334 Time evolution of the passive tracer concentration in the simulations (Figures 5—7)  
335 reveals general agreement among the models. In particular, during the first several years, the  
336 models simulate rapid spreading of the freshwater tracer into the Labrador Sea (mainly with  
337 the Labrador Current) and Baffin Bay where the highest concentration is maintained  
338 throughout the simulation. On the eastern side of Greenland, the tracer is exported from the  
339 Nordic Seas by the EGC into the Labrador Sea and the Subpolar Gyre region, and then the  
340 tracer is advected back into the Nordic Seas with the North Atlantic current. Yet there are  
341 obvious discrepancies in the simulated fields. The differences among the models are found in  
342 tracer concentration (and mass) in the sub-Arctic seas, tracer pathways and spreading into the  
343 interior basins, and timing of tracer advection into the Nordic Seas.

344 Specifically, during the first 5 years of the AO-HYCOM simulation the tracer stays  
345 within the EGC. There is an indication of some advection of the tracer with the East Icelandic  
346 Current to the east (Years 1, 2, 5, and 6). However, the interior Nordic Seas show no  
347 signature of the tracer until year 6. After that, the interior Nordic Seas slowly fill up with the  
348 tracer being brought by the North Atlantic Current after the tracer has travelled around the  
349 Subpolar Gyre (Figure 5).

350 The NEMO-LIM2 simulation (Figure 6) agrees with AO-HYCOM showing very  
351 limited lateral advection of the tracer off the EGC during the first 5 years of the simulation.  
352 Yet in contrast to AO-HYCOM, tracer spreading into the interior Nordic Seas by means of  
353 the west-to-east Jan Mayen current is evident in NEMO-LIM2, . The timing of tracer

354 propagation into the convective regions of the Nordic Seas is remarkably similar between the  
355 two models. Nevertheless, there is a notable distinction in how this propagation is simulated.  
356 In NEMO-LIM2 the tracer spreads from the south with the North Atlantic Current and from  
357 west off the EGC, whereas in AO-HYCOM the tracer is mainly advected by the North  
358 Atlantic Current.

359         The ICMMG simulation (Figure 7) qualitatively agrees well with both AO-HYCOM  
360 and NEMO-LIM2. Yet the ICMMG simulation does not predict any substantial tracer  
361 spreading off the EGC into the interior Nordic Seas, unlike the NEMO-LIM2 simulation. In  
362 ICMMG, it takes much longer (10 years) for the tracer to travel around the Subpolar Gyre  
363 before returning to the Nordic Seas with the North Atlantic Current, compared to AO-  
364 HYCOM and NEMO-LIM2. Another discrepancy between ICMMG and the two other  
365 models is sluggish propagation of the tracer into the interior of the basins in the ICMMG  
366 simulation. For example, propagation of the tracer in Baffin Bay is predominantly along the  
367 coast in the ICMMG simulation. By contrast, there is substantial lateral advection of the  
368 tracer into the interior Baffin Bay in both AO-HYCOM and NEMO-LIM2. Similarly in the  
369 Labrador Sea and the Nordic Seas, ICMMG simulates a slower rate of tracer spreading into  
370 the interior basins.

371         Another disagreement between ICMMG and the two other models is the evolution of  
372 the tracer concentration to the north and west off Greenland. ICMMG simulates an extensive  
373 spread of the tracer into the Arctic Ocean from the northern Greenland coast, whereas in both  
374 AO-HYCOM and NEMO-LIM2 simulations tracer propagation into the Arctic Ocean is  
375 limited and confined to the narrow shelf north of Greenland and the CAA.

376

#### 377 *4.2. Tracer budget analysis*



378 In order to characterize propagation and accumulation of the tracer in the sub-Arctic  
 379 seas, a tracer budget is calculated for the six regions shown in Figure 1. The study area is  
 380 divided into three basins: the Nordic Seas (NS), the Labrador Sea (LS), and Baffin Bay (BB).  
 381 Within the NS and LS, convective sites in the Greenland Sea (GS), the Iceland Sea (IS), and  
 382 interior Labrador Sea (IL) are designated. For each region, passive tracer storage (volume-  
 383 integrated tracer content or mass,  $M_{tr}$ ) is calculated as

$$384 \quad M_{tr}(t) = \int \int \int_V C_{tr}(x, y, z, t) dV, \quad (1)$$

386 where  $C_{tr}$  is tracer concentration ( $\text{kg m}^{-3}$ ). Within a region, tracer fluxes ( $F_{tr}$ ) are estimated  
 387 for each segment bounding the region by integrating over the total water depth ( $H$ ) along the  
 388 segment  $l$  of length  $L$

$$390 \quad F_{tr} = \int_L \int_H C_{tr}(l, z, t) \mathbf{u}(l, z, t) \cdot \mathbf{n}(l) dz dl \quad (2)$$

392 where  $\mathbf{n}$  is an inward unit normal vector (positive flux into the region).  
 393

#### 394 4.2.1. The Nordic Seas

395 All the models show a gradual increase of the volume-integrated tracer content  
 396 signifying tracer accumulation in the Nordic Seas region over the simulation time (Figure 8a).  
 397 However, the models disagree on the evolution rate of the tracer content. In the AO-HYCOM  
 398 simulation, concentration of the tracer starts growing after the first 3 years. In the other two  
 399 simulations, the content demonstrates a nearly steady increase over the whole integration  
 400 time. All models show a negative flux across Denmark Strait (section 1) owing to the tracer  
 401 export with the EGC (Figures 8 b—d); all models simulate a reasonably robust seasonal  
 402 signal in this transport. In AO-HYCOM (Figure 8b), the tracer net flux is negative during the  
 403

404 first 5 years when tracers are exported out of the region. After 5 years of the simulation,  
405 tracer inflow to the Nordic Seas starts increasing through Iceland-Faroe-Shetland segments  
406 (sections 2 and 3 corresponding to the green and blue lines, respectively) and the net oceanic  
407 flux becomes positive exceeding the outflow through section 1. In NEMO-LIM2 (Figure 8c)  
408 the net flux is negative dominated by the EGC export of the tracer during the simulation.  
409 Contrary to the AO-HYCOM results, tracer fluxes across Iceland-Faroe-Shetland segments  
410 exhibit only minor increase after year 6. The increase in tracer content in the NS is due to the  
411 tracer advection with the East Icelandic and Jan Mayen currents in NEMO-LIM2 (Figure 6).  
412 Differing from AO-HYCOM, ICMMG (Figure 8d) simulates considerably smaller tracer  
413 inflow into NS with the North Atlantic Current through sections 2 and 3. The inflow cannot  
414 compensate tracer outflow through Denmark Strait. The imbalance in the net advective tracer  
415 fluxes is compensated by internal tracer flux along the Greenland coast, whereas in the AO-  
416 HYCOM the net advective flux is positive after 5 years indicating that tracer accumulation in  
417 the NS region is mainly attributed to the tracer advection by the North Atlantic Current.

418

#### 419 *4.2.2. The Greenland Sea*

420 Tracer accumulation in the GS deduced from the time series of the volume-integrated  
421 tracer content is largest for the AO-HYCOM model (Figure 9a). During the first 2-3 years,  
422 there is no tracer signature in the region for all models. After several years, the tracers start  
423 spreading into the interior Greenland Sea. However, the timing and the propagation rate  
424 differ among the models. In AO-HYCOM, the tracer content starts growing after 4 years.  
425 During the last 4 years of the simulation, the accumulation rate slows down. Both NEMO-  
426 LIM2 and ICMMG simulate accumulation of the tracers in the GS at a slower rate than in  
427 AO-HYCOM.

428           There is good agreement between the general trend of tracer fluxes in AO-HYCOM  
429 and NEMO-LIM2, with much smaller fluxes in general for the ICMMG simulations (Figures  
430 9b and d). Tracer fluxes across sections 9 and 12 dominate but nearly cancel each other due  
431 to the throughflow of the recirculating Atlantic waters in the northern side of the GS. It is the  
432 balance between the fluxes through the eastern and southern sections that determines the  
433 tracer budget inside the box. In the AO-HYCOM experiment, the inflow across the eastern  
434 section (10) is mostly positive and the southern flux oscillates between positive and negative.  
435 The fluxes in ICMMG remain near zero through year 13.

436           Tracer fluxes across sections 9 and 12 in NEMO-LIM2 have a strong seasonal signal  
437 due to the propagation path of the tracer from the EGC (Figure 6), which has strong seasonal  
438 change in the tracer concentration related to seasonal variability of Greenland freshwater flux  
439 (Figure 2a). After year 8, the seasonal signals of the fluxes across sections 9 and 12 are  
440 superimposed on positive and negative trends, respectively, attributed to tracer that has  
441 travelled around the Subpolar Gyre and has been advected into the GS with the Norwegian  
442 Atlantic Current.

443

#### 444 *4.2.3. The Iceland Sea*

445           Accumulation of the passive tracer in the IS is delayed by about 1 year in NEMO-  
446 LIM2, 5 years in AO-HYCOM and 10 years in ICMMG (Figure 10a). The earlier  
447 accumulation of the tracer in the IS region in NEMO-LIM2 is related to eastward tracer  
448 transport with the East Icelandic Current (Figure 6) that is not evident in the other two  
449 simulations.

450           Tracer fluxes across the sections have noticeable disagreement among the models  
451 (Figures 10b—d). AO-HYCOM simulates major tracer outflow through the southern section  
452 (16) and the inflow is mainly through the northern (14) and eastern (15) boundaries. Tracer

453 flux through the western boundary (13) is intermittent but becomes predominantly positive by  
454 the end of the simulation. Note that after year 6, tracer flux across the southern section (16) is  
455 always negative, whereas the northern flux (14) remains positive. Similarly, ICMMG  
456 simulates persistent positive tracer flux through the northern section and negative tracer flux  
457 through the southern section (Figure 10d), however the northern inflow dominates the  
458 outflow (note the different scales). In the NEMO-LIM2 simulation, the fluxes through  
459 sections 14 and 16 are more sporadic and do not demonstrate any persistence, in contrast to  
460 AO-HYCOM and ICMMG. The oscillations across sections 13 and 16 are related to  
461 variability of the East Icelandic Current and the Jan Mayen Current that are substantial tracer  
462 sources for the interior Nordic Seas in the model.

463

#### 464 4.2.4. *The Labrador Sea*

465 All model experiments predict an overall increase of the passive tracer content in the  
466 LS during the simulation (Figure 11a). The rate of tracer increase is similar in all simulations.  
467 Tracer fluxes across the northern (7) and southern (8) sections exhibit strong seasonality  
468 associated with seasonal changes of meltwater flux from Greenland coast (Figures 11b—d).  
469 In AO-HYCOM and NEMO-LIM2, the southern flux through section 8 is positive during the  
470 first year of the simulation, and then it becomes predominantly negative. The shift is due to  
471 the fact that during the first year, the tracer is imported into the region with the WGC. After  
472 the tracer propagates around the Labrador Sea, it is exported out to the North Atlantic with  
473 the Labrador Current. Tracer flux between the Labrador Sea and Baffin Bay (section 7)  
474 oscillates on the seasonal time scale remaining mainly positive during the first half of the  
475 year and is reversed during the second half. Such a strong seasonal cycle is consistent with  
476 hydrographic observations in the West Greenland Current [Myers *et al.*, 2009; Curry *et al.*,  
477 2014; Rykova *et al.*, 2015]. In the ICMMG simulation, the tracer flux across section 8

478 remains predominantly positive through year 10 caused by tracer influx from the North  
479 Atlantic. In contrast to the AO-HYCOM and NEMO-LIM2, the tracer is exported to Baffin  
480 Bay (through section 7) with little tracer export from the bay until year 11.

481

#### 482 *4.2.5. The Interior Labrador Sea*

483 The rate and timing of tracer accumulation in the IL region agree between AO-  
484 HYCOM and ICMMG, whereas NEMO-LIM2 predicts a near-linear increase of the tracer  
485 until year 12 when tracer mass reaches maximum that is more than 2 times higher than in the  
486 AO-HYCOM simulation and nearly 5 times higher than in ICMMG (Figure 12a). Tracer  
487 fluxes across the sections bounding the IL box indicate different tracer advection pathways in  
488 the model experiments (Figures 12b—d). In AO-HYCOM, the fluxes do not have any  
489 persistent pattern. On average, tracer content in the IL is determined by the influx through the  
490 eastern (section 18) and northern (17) sides and the outflows through the western (20) and  
491 southern (19) sides. The NEMO-LIM2 experiment has the strongest tracer inflow through the  
492 southern section (19), while tracer flux across the northern section (17) is primarily negative.  
493 Similar to AO-HYCOM, in ICMMG the tracer budget in the IL box is largely influenced by  
494 the flux across the eastern side (18). Tracer outflow from the IL mostly occurs through the  
495 northern section (17).

496

#### 497 *4.2.6. Baffin Bay*

498 The BB region shows accumulation of passive tracer in all experiments with the  
499 accumulation rate differing among the models (Figure 13a). In AO-HYCOM, the tracer  
500 content rapidly increases during the first 2 years. After that, the tracer content slowly  
501 increases with superimposed seasonal variability. The NEMO-LIM2 simulation predicts  
502 faster and more persistent tracer accumulation resulting in two time more tracer mass by the

503 end of the integration compared to AO-HYCOM. Results from the ICMMG experiment  
504 demonstrate a near-linear increase in the tracer content in the BB region through the  
505 simulation also exceeding twice the tracer content in the AO-HYCOM simulation by the end  
506 of the experiment. Fluxes across the northern (6) and the southern (7) sections are dominated  
507 by seasonal change (Figures 13b—d). During the first half of the year the tracer is advected  
508 from the Nares Strait but during the second half of the year the tracer is advected from the  
509 Labrador Sea through Davis Strait. The seasonality is attributed to wind climatology in the  
510 area with strong southerly winds in winter and weaker northerly winds in the eastern bay  
511 during summer [*Tang et al.*, 2004].

512

#### 513 ***4.3. Tracer Distribution in the Labrador Sea, the Nordic Seas and Baffin Bay***

514 The numerical experiments illustrate that the tracer has been redistributed by ocean  
515 circulation within the entire sub-Arctic region over a relatively short time. At the beginning  
516 of the experiment, the tracer is predominantly located in the Labrador Sea and Baffin Bay  
517 (Figures 5—7). As the tracer spreads over the domain, the tracer content increases in the  
518 Nordic Seas. Analysis of the volume-integrated tracer content (section 4.2) reveals that tracer  
519 mass increases in all regions. Part of this increase is due to the overall increase of tracer mass  
520 in the domain and comparatively minor leakage of the tracer to the southern North Atlantic  
521 and the Arctic Ocean. However redistribution of the tracer within the sub-Arctic seas driven  
522 by the ocean circulation influences the rate of tracer accumulation rate in the basins.

523 In order to compare the accumulation rates, the ratio of volume-integrated tracer  
524 content (mass) in the Labrador Sea, the Nordic Seas and Baffin Bay relative to the total tracer  
525 mass released is analyzed (Figure 14). This ratio in the Labrador Sea and Baffin Bay  
526 decreases with time in all experiments. By the end of the simulations, the ratio of tracer mass  
527 in the Labrador Sea and Baffin Bay relative to the total tracer mass released is highest in the

528 ICMMG simulation compared to the other two models. This ratio indicates that the interior  
529 Labrador Sea (Figure 14, green numbers) loses the tracer over the course of the AO-HYCOM  
530 and ICMMG simulations, while in the NEMO-LIM2 simulation there is an accumulation of  
531 tracer in the region. In the Nordic Seas, the evolution of the tracer mass ratio is in the  
532 opposite sense between the AO-HYCOM the other two models indicating substantial  
533 differences in the simulated tracer spreading in the region. All the simulations indicate  
534 accumulation of tracer inside the GS and IS boxes (Figure 14, blue and red numbers). By the  
535 end of the simulations, the three basins contain 29.5% (AO-HYCOM), 28% (NEMO-LIM2),  
536 and 38.2% (ICMMG) of the total tracer mass. The remainder of the tracer is accumulated in  
537 the Subpolar Gyre region and in the Arctic Ocean. It is of note that the largest fraction of the  
538 tracer mass remains in the sub-Arctic seas (including the Subpolar Gyre region) after 14 years  
539 of the simulations, demonstrating that the tracer has been accumulating in the region while  
540 being transported around the sub-Arctic seas and propagating into the interior regions. The  
541 result corroborates previous studies that discuss recirculation of negative and positive salinity  
542 anomalies within the Labrador – Subpolar Gyre – Nordic Seas [*Belkin, 2004; Yashayaev and*  
543 *Seidov, 2015*].

544

#### 545 ***4.4. Tracer spreading into the deep layers***

546 Observations suggest that water masses propagate predominantly along isopycnal  
547 surfaces [e.g., *Ledwell et al., 1993*]. Greenland freshwater spreads into the sub-Arctic seas  
548 remaining near the surface and mixing with ambient water. In the sub-Arctic seas, freshwater  
549 is mixed downward by haline and thermal convection [e.g., *Watson et al., 1999; Rudels et al.,*  
550 *2012*]. Thus, Greenland meltwater ultimately penetrates into the deep layers in the convective  
551 areas. Several studies suggest that freshwater can be mixed downward into the deep layers  
552 and then laterally advected to distant basins. In the numerical experiments, the tracers being

553 released in the near-surface layer tagging buoyant meltwater quickly penetrate the subsurface  
554 layers through mechanical mixing and deep convection. Spurious diapycnal mixing can also  
555 enhance vertical propagation of the tracer in the non-isopycnal NEMO-LIM2 and ICMMG  
556 models [*Bleck, 2002; Hill et al., 2012*].

557         The models predict tracer maximum concentration in the near-surface layer (see  
558 sections in Figure 15). However, the maximum is strongly diffused in NEMO-LIM2 and  
559 ICMMG occupying depths between 200 and 500 m, by contrast, it remains within a narrow  
560 layer (upper 100 m) in AO-HYCOM. In the AO-HYCOM experiment, the deepest  
561 propagation of the tracer occurs in the North Atlantic (Figures 15a and b). In the Nordic Seas,  
562 the tracer is at notably shallower depths (~800m) with distinct minima in the Greenland Sea  
563 (shallower than ~700 m). Similarly, in Baffin Bay the tracer has not spread deeper than  
564 ~800m. Development of deep local tracer maxima is evident in the southern Labrador Sea by  
565 the end of year 7 (also observed in Figure 14a). After 14 years of the model experiment, the  
566 passive tracer has penetrated depths exceeding 1600-1800 m in the North Atlantic including  
567 the Labrador Sea and the southern Irminger Sea. The concentration decreases with depth and  
568 is maximal in the surface layer (<200 m).

569         In contrast to AO-HYCOM, NEMO-LIM2 simulates tracer spreading through the  
570 whole water column in Baffin Bay (Figures 15c and d). In the Labrador Sea, the tracer is  
571 distributed through most of the water column down to ~2800 m after 7 years and all the way  
572 to the bottom after 14 years. This is related to excessive Labrador Sea convection in this  
573 simulation. After 7 years, the tracer propagation shallows downstream of the North Atlantic  
574 flow with very shallow tracer signatures in the Nordic Seas owing to insufficient propagation  
575 of the tracer into this region. By the end of the simulation, the tracer spreads to the bottom in  
576 the North Atlantic and below 600 m in the Nordic Seas.



577 In the ICMMG experiment (Figures 15e and f), vertical propagation of the tracer  
578 generally agrees with AO-HYCOM. After 14 years, the model simulates deeper propagation  
579 of the tracer in the Labrador Sea and the North Atlantic basin and substantially shallower  
580 propagation in the Nordic Seas and Baffin Bay. ICMMG tracer distribution after 7 years  
581 differs from both the AO-HYCOM and NEMO-LIM2 simulations. Specifically, in ICMMG  
582 the tracer spreads deeper in the Labrador Sea, compared to the AO-HYCOM but shallower  
583 than in NEMO-LIM2. There is no tracer signature in the Nordic Seas in the ICMMG. There  
584 is good agreement in tracer depth in Baffin Bay between AO-HYCOM and ICMMG. After  
585 14 years, tracer distribution in the North Atlantic basin in the ICMMG looks more different  
586 from AO-HYCOM. Interestingly there is no tracer in the near-surface layers in the North  
587 Atlantic basin in the ICMMG simulation, suggesting the advection origin of the subsurface  
588 tracer presence in this part of the section. The absence of the tracer in the central Labrador  
589 Sea and Nordic Seas is due to the tracer propagation with the boundary currents in this  
590 simulation.

591

## 592 **5. Discussion**

### 593 ***5.1. Pathways of Greenland Freshwater***

#### 594 *5.1.1. Horizontal propagation*

595 The pathways of Greenland freshwater propagation follow the general circulation in  
596 the sub-Arctic seas (Figure 1). The presented tracer experiments demonstrate relatively rapid  
597 spreading of the passive tracer in the sub-Arctic seas reaching the most remote regions within  
598 5 – 7 years. Estimated travel time of the tracer in the model experiments agrees well with the  
599 estimates of propagation rates of the GSAs in the sub-Arctic inferred from observations  
600 [Dickson *et al.*, 1988; Belkin *et al.*, 1998; Belkin, 2004; Yashayaev and Seidov, 2015]. Based  
601 on the observed cooling and freshening signals in the region, Dickson *et al.* [1988] traced the

602 spreading of the 1970s GSA and suggested that it took the GSA 7-8 years to propagate from  
603 Fylla Bank (off the southwestern Greenland coast) to the Norwegian Sea and 9-10 years to  
604 Spitsbergen. Upstream, propagation timescales for the GSA were 1 year longer from the  
605 eastern Greenland coast (north of Iceland). The GSAs of the 1980s and the 1990s spread  
606 faster propagating from Fylla Bank to the Norwegian Sea in about 5-6 years [*Belkin et al.*,  
607 1998; *Belkin*, 2004]. The major inflow of the tracer into the Nordic Seas in AO-HYCOM and  
608 ICMMG occurs with the North Atlantic Current. This result is in agreement with other  
609 studies where the dominant role of negative salinity anomalies carried by the Atlantic inflow  
610 on the freshening signal in the Nordic Seas has been proposed [e.g., *Glessmer et al.*, 2014;  
611 *Reverdin*, 2014].

612         The discrepancies in the numerical solutions are apparent in tracer propagation into  
613 the interior regions of the sub-Arctic seas. There are markedly stronger horizontal gradients  
614 in tracer concentration fields simulated in ICMMG, compared to the other two models. In the  
615 coarse-resolution ICMMG simulation, the interior slowly fills with the tracer that tends to  
616 remain with the current following the boundary regions (Figure 7). For example, in ICMMG  
617 the tracer is advected into the Nordic Seas after 5 years of the simulation in agreement with  
618 the AO-HYCOM experiment, yet it takes another 7 – 8 years for the tracer to spread into the  
619 interior GS and IS in the ICMMG simulation compared to 2 – 3 years in the AO-HYCOM  
620 experiment (Figure 5). One possible reason for the observed discrepancies in the tracer  
621 distribution between ICMMG and the other two models could be the coarse grid spacing of  
622 the former. The coarse resolution of the ICMMG model does not permit the simulation of  
623 mesoscale eddies that may be a key mechanism for advection of tracer into the interior  
624 regions.

625         This assumption is supported by previous studies suggesting a leading role of small-  
626 scale eddies in lateral advection and spreading of both freshwater and Atlantic Water in the

627 Labrador Sea [Saenko *et al.*, 2014] and the Nordic Seas [Budeus and Ronski, 2009;  
628 Yashayaev and Seidov, 2015]. For example, Budeus and Ronski [2009] reported a patchy  
629 distribution of Atlantic Water observed inside the Greenland Gyre. Diameters of these  
630 “Atlantic Water patches” were only 20 km with vertical extent varying from 200 to 1000 m.  
631 The findings of Budeus and Ronski [2009] support the idea of eddy transport of Atlantic  
632 Water into the interior Greenland Gyre (see also Yashayaev and Seidov [2015] for an  
633 interesting analogy of Atlantic Water circulation and a pinball machine). This view is distinct  
634 from the idea of a diffusion type of penetration that assumes gradual lateral spreading  
635 resulting in a smooth and steady transition from salinity and temperature values at the  
636 boundaries of the gyres to the interior values [e.g., Karstensen *et al.*, 2005]. This type of  
637 advection is present in the ICMMG (non-eddy) simulation.

638         The above discussion raises the question about the meaning of “eddy-resolving” with  
639 respect to Arctic Ocean models. An “eddy-resolving” model has to be capable to adequately  
640 representing the eddy field, i.e. model horizontal grid spacing should be at least two grid  
641 points per Rossby radius of deformation and the grid spacing is measured as the grid-diagonal  
642 distance (so called “effective spacing” in Hallberg [2013]). In the Nordic Seas, the first  
643 baroclinic Rossby radius of deformation is ~4–8 km [Nurser and Bacon, 2013]. Hence, an  
644 “eddy-resolving” model should have a grid spacing about 2–4 km. Only AO-HYCOM with  
645 0.08° horizontal grid (effective spacing ~4–5 km in the Greenland Sea) is close to this  
646 estimate still being marginal between “eddy-resolving” and “eddy-permitting” (1 grid point  
647 per radius [Nurser and Bacon, 2013]). No other model from the previous studies mentioned  
648 earlier (section 1) satisfies this resolution criterion. With >20km horizontal grid spacing, in  
649 the ICMMG model tracer propagation into the gyres is dominated by horizontal diffusion,  
650 which is much slower than transport and mixing by eddies in the other two models.

651 The apparent disagreement in the NEMO-LIM2 solution with the other models is that  
652 NEMO-LIM2 shows stronger tracer propagation from the EGC with the Jan Mayen Current.  
653 The causes of this intensification are not clear. We speculate that this may be related to  
654 misrepresentation of the narrow East Greenland Coastal Current, which serves as a  
655 freshwater conduit in the Nordic Seas. This problem has been discussed with respect to  
656 another 0.25° NEMO simulation in *Marsh et al.* [2010]. Alternatively, it could be due to the  
657 stronger East Icelandic Current in NEMO-LIM2 compared to the other two models.

658

### 659 5.1.2. Non-Eddying and Eddy Fluxes

660 The general circulation in the sub-Arctic seas follows the boundaries of the basins  
661 suggesting no tracer propagation into the interior Labrador and Nordic Seas. The AO-  
662 HYCOM and NEMO-LIM2 simulations, however, indicated tracer spreading into the interior  
663 regions (section 4) in striking contrast to the coarse-resolution ICMMG simulation that  
664 showed very minor propagation of the tracer into the interior boxes. Visual inspection of  
665 figures 5–7 support the rationale suggested above that different eddy activity explains the  
666 different spreading of the tracer into the interior regions. To validate this assumption, eddy  
667 tracer flux and mean (non-eddy) tracer flux are calculated by separating velocities and  
668 concentration into time-mean velocity ( $\bar{\mathbf{u}}$ ) and tracer concentration ( $\bar{C}_{tr}$ ) and corresponding  
669 time-fluctuating components or eddy terms ( $\mathbf{u}'$  and  $C'_{tr}$ ). After time averaging, the tracer  
670 fluxes may be separated into mean and eddy fluxes

$$671 \overline{\mathbf{u} \cdot C_{tr}} = \bar{\mathbf{u}} \cdot \bar{C}_{tr} + \overline{\mathbf{u}' \cdot C'_{tr}}, \quad (3)$$

672 where the overbar denotes 30-day time averaging.

673 Figure 16 presents one-year-averaged mean and eddy tracer fluxes for year 9 from the  
674 simulations. The mean tracer flux (Figures 16 a, c, and e) highlights the pathways of the  
675 tracer in the sub-Arctic seas that follow the general large-scale circulation. The general

676 features of the mean tracer flux are similar in the models. Yet there are notable discrepancies  
677 in the smaller-scale pathways. For example, in contrast to the higher-resolution simulations,  
678 in ICMMG boundary currents are broad occupying nearly whole Baffin Bay. At the same  
679 time, the interior Nordic Seas is nearly quiescent. Note that the mean tracer flux in the Nares  
680 Strait is northward, in contradiction to AO-HYCOM and NEMO-LIM2 (as well as the  
681 observations). The eddy tracer fluxes (Figures 16 b, d, and f) are predominantly directed  
682 normal to the mean flux, indicating lateral advection of the tracer into the interior regions. All  
683 the models simulate increased eddy tracer flux along southwestern and southeastern  
684 Greenland coast. The magnitude of the eddy tracer flux in the Labrador Sea is markedly  
685 higher in AO-HYCOM compared to the other two experiments (note the natural-log scale).  
686 Also AO-HYCOM fields predict a local maximum of the eddy tracer fluxes in the Norwegian  
687 Sea that is smaller in NEMO-LIM2 and is absent in ICMMG.

688 Next, time-integrated eddy tracer flux ( $F_{eddy}$ ) and time-mean tracer flux ( $F_{mean}$ ) in the  
689 upper 150 m are calculated for the interior boxes in the Greenland, Iceland, and Labrador  
690 seas

$$691 \quad F_{mean}(t) = \int_t \int_H \int_L \bar{u}_\perp \cdot \bar{C}'_{tr} dl dz dt, \quad (4)$$

$$692 \quad F_{eddy}(t) = \int_t \int_H \int_L \overline{u'_\perp \cdot C'_{tr}} dl dz dt, \quad (5)$$

693 where  $u_\perp$  is the velocity component normal to the section.

694 Shown in Figure 17 are time-integrated eddy tracer flux (Eq. 4) and mean tracer flux  
695 (Eq. 5) calculated for years 8—12. All the models demonstrate modest contribution of eddy  
696 tracer flux compared to the mean flux in the Nordic Seas. This is an expected result from the  
697 previous analysis of Figure 16. Strikingly different from the Nordic Seas is the substantial  
698 contribution of eddy flux in the Labrador Sea in the AO-HYCOM simulation. The amount of  
699 tracer advected by eddies into the IL box is higher than the amount advected by the mean

700 fluxes. This result is different from the NEMO-LIM2 and ICMMG simulations where the  
701 eddy tracer flux is negligibly small in the IL region.

702 Thus, analysis of the tracer fluxes simulated in the models has demonstrated the  
703 predominant role of the mean tracer flux into the interior Nordic Seas. The eddy tracer flux  
704 dominates the non-eddy tracer flux in the interior Labrador Sea in the AO-HYCOM  
705 simulation. Calculated eddy fluxes do not support the idea of eddy advection as the dominant  
706 mechanism of tracer transport into the interior Nordic Seas, in contrast to the Labrador Sea.

707

## 708 ***5.2. Greenland Freshwater Influence on Salinity in the sub-Arctic Seas from AO-HYCOM***

709 The simulations with the passive tracer have demonstrated rapid tracer propagation  
710 over the sub-Arctic seas suggesting widespread influence of Greenland freshwater flux in the  
711 region in agreement with the hypothesis discussed in Section 1. The tracer concentration  
712 increases in all the sub-Arctic seas during 14 years of model simulations indicating  
713 accumulation of Greenland freshwater in the basins. The results imply the growth of  
714 freshwater content that should manifest in increasing negative salinity anomaly in the upper  
715 sub-Arctic seas. Nevertheless, there is no obvious observational evidence that could relate  
716 freshening signals in the region with Greenland freshwater flux. According to the model  
717 results, accumulation over the last two decades of surplus Greenland freshwater should have  
718 resulted in a negative salinity trend in the near-surface layer. Instead, observed salinity  
719 changes alternate between positive and negative anomalies with no persistent temporal  
720 pattern (as discussed in section 3). In the following section, the influence of the Greenland  
721 freshwater flux on salinity changes in the sub-Arctic region is analyzed using results from  
722 only the AO-HYCOM simulation, which has the highest resolution.

723

### 724 ***5.2.1. Greenland Freshwater Flux and Salinity Change in the Sub-Arctic***

725 The AO-HYCOM experiment was initialized from another simulation that had been  
726 integrated without Greenland runoff. In the tracer experiment, Greenland freshwater flux is  
727 “turned on” at the first time step of the model integration. Hence, there is a discrete jump in  
728 the freshwater forcing from the initial conditions with zero Greenland runoff to  $>1000 \text{ km}^3$   
729  $\cdot\text{yr}^{-1}$  (0.032 Sv) in the simulation. This provides easier detection and tracking of Greenland  
730 freshwater propagation and associated freshwater anomalies in the sub-Arctic seas. The  
731 drawback of the experiment design is that simulated negative salinity changes in the sub-  
732 Arctic seas can be too extreme for direct comparison with observational records. A better  
733 approach would be to initialize the experiment with the ocean fields from another experiment  
734 forced by realistic Greenland runoff, although in practice this still complicates the process of  
735 detection of freshwater anomalies in the basin. The AO-HYCOM results are employed here  
736 to establish relationships between Greenland freshwater flux and salinity changes in the sub-  
737 Arctic. This relationship will be used to estimate freshening in the sub-Arctic seas caused by  
738 the surplus Greenland freshwater flux in 1990-2010.

739 Both freshwater and salinity changes in the sub-Arctic basins have linear trends  
740 during the simulation (Figure 18). For every region (Figure 1), salinity is spatially averaged  
741 within the slab 0-400 m inside the 500-m isobath. A linear relationship is sought between the  
742 spatially averaged salinity and cumulative Greenland freshwater mass  $Q_r$ .

$$743 \quad Q_r(t) = \int_0^t F_r dt, \quad (6)$$

744 where  $F_r$  is Greenland freshwater flux. The tracer experiments have demonstrated a delay  
745 between the time when freshwater (or passive tracer) is released at the Greenland coast and  
746 the time of appearance of the freshwater signature in the sub-Arctic basins (Figure 4). The  
747 time lag ( $\tau$ ) varies for the basins from 1 year for Baffin Bay to more than 5 years for the  
748 Greenland and Iceland Seas. Thus, the following regression is fit to the data

749

$$750 \quad dS(t) = \alpha_0 + \alpha_1 \cdot Q_r(t - \tau), \quad (7)$$

751

752 where  $dS$  is salinity change averaged in the upper 400 m within a sub-Arctic basin at time  $t$   
753 (Figure 18b). The time lag is defined iteratively until the best fit to the data is provided, based  
754 on the highest value of the coefficient of determination ( $R^2$ ). The high values of  $R^2$  in the  
755 analyzed cases indicate a strong linear relationship between the Greenland freshwater flux  
756 and the magnitude of salinity anomaly in the regions.

757 The next step is to “remap”  $dS$  onto the observational Greenland freshwater changes  
758 reported in *Bamber et al.* [2012]. Greenland freshwater flux increased by  $\sim 200 \text{ km}^3 \text{ yr}^{-1}$   
759 ( $\sim 0.006 \text{ Sv}$ ) from 1990 to 2010. Assuming a linear increase in the Greenland freshwater flux  
760 over this period, Eq. (7) provides the estimated salinity change ( $dS$ ) in the six sub-Arctic  
761 regions for 1990–2015 (Figure 18c). The magnitude of salinity anomaly is modest in the  
762 basins, especially in the interior Nordic Seas (GS and IS), suggesting a relatively small  
763 contribution of the Greenland runoff on the freshening of the sub-Arctic seas over 20 years of  
764 increased Greenland freshwater flux. According to Figure 18c, by 2010 salinity anomaly  
765 should be  $-0.05$  in the GS,  $-0.07$  in the IS, and  $-0.08$  in the IL as a result of surplus  
766 Greenland freshwater flux.

767 An important caveat here is that salinity changes of the North Atlantic Current have  
768 not been taken into account. According to long-term observations in the Iceland-Faroe-  
769 Shetland section and in the southwestern Nordic Seas salinity of Atlantic water increased by  
770  $0.15 - 0.2$  in the 2000s compared to the long-term mean [*Holliday et al.*, 2008; *Yashayaev*  
771 *and Seidov*, 2015]. Such a strong positive salinity anomaly would dominate and undermine  
772 the impact of Greenland freshwater runoff on salinity in the sub-Arctic seas.

773



774

## 775 **6. Summary**

776           As Greenland Ice Sheet melting has been accelerating, increased freshwater discharge  
777 into the ocean can have dramatic consequences for thermohaline circulation of the sub-Arctic  
778 seas. Generally accepted ramifications of the surplus freshwater flux are increased water  
779 column stability and weakening of deep convection in the interior Labrador and Nordic Seas,  
780 not to mention sea level rise as the most prominent consequence. Several hypotheses of the  
781 current and future climate changes in the Arctic and North Atlantic have been suggested on  
782 the basis of this assumption. Yet pathways and time scales of freshwater propagation in the  
783 sub-Arctic seas are not known. The influence of Greenland freshwater on convective regions  
784 remains elusive. Moreover, recent observational records demonstrate increasing salinity in  
785 the upper Nordic Seas and the Labrador Sea during the 2000s contradicting the anticipated  
786 freshening caused by accelerated Greenland Ice Sheet melt. Although freshening in the 0–200  
787 m layer in the Labrador Sea was observed during the 2010s [*Beszczynska-Möller and Dye,*  
788 *2013; Yashayaev et al. 2015*]. In order to address these uncertainties, three numerical  
789 experiments were conceived during FAMOS discussions. In these experiments employing  
790 AO-HYCOM, NEMO-LIM2, and ICCMG ocean-sea ice models, a passive tracer was  
791 continuously released at Greenland freshwater source sites to track propagation of the  
792 freshwater in the sub-Arctic seas.

793           Results from the tracer experiments demonstrate general agreement among the models  
794 on timing and propagation of the tracer in the sub-Arctic seas. The tracer follows the large-  
795 scale ocean circulation pattern with the EGC, WGC and the Subpolar Gyre from Greenland  
796 to the North Atlantic and with the North Atlantic Current to the Nordic Seas. The tracer  
797 quickly propagates into Baffin Bay and the Labrador Sea. The interior Labrador Sea is  
798 impacted by the tracer within the first 2 years of the simulations. The major pathway of the

799 tracer to the Nordic Seas is via the North Atlantic Current after it has traveled around the  
800 Subpolar Gyre. It takes from 3 (AO-HYCOM) to 5 (NEMO-LIM2 and ICMMG) years for  
801 the tracer to reach the south Nordic Seas via the North Atlantic Current and another 2 years to  
802 enter convective sites in the interior Nordic Seas (except for ICMMG where tracer spreading  
803 into the interior region was not substantial over the simulation time). The Iceland Sea  
804 receives the tracer via the Iceland Current shortly after the simulation begins, yet the amount  
805 of this influx is not substantial, except in the NEMO-LIM2 simulation. Positive trends in the  
806 time series of the tracer mass inside the basins indicate that a steady state has not been  
807 reached in the simulation and the tracer was still accumulating in the sub-Arctic seas by the  
808 end of the model runs. Nevertheless, the accumulation has substantially slowed in Baffin Bay  
809 and the Labrador Sea to the end of the experiments.

810         The model experiments predict highest tracer concentration in Baffin Bay with the  
811 second highest concentration in the Labrador Sea. The concentration in these basins quickly  
812 reached a quasi-steady state (during the first 4-5 years). In the rest of the domain, the  
813 concentration increased as the tracer propagated with the Subpolar Gyre and the North  
814 Atlantic Current towards the Nordic Seas. During the 14-year time interval, the tracer  
815 accumulated in the sub-Arctic seas. However, the simulations were not sufficiently long to  
816 reach a steady state in tracer concentration within the domain precluding determination of the  
817 residence time of the tracer in the sub-Arctic basins.

818         The noticeable discrepancy in the numerical solutions is tracer propagation into the  
819 convective regions in the Labrador Sea and the Nordic Seas. In the ICMMG simulation the  
820 interior regions are filled slowly with the tracer resulting in strong horizontal gradients of the  
821 tracer concentration, whereas the other two models simulate faster propagation of the tracer  
822 into the interior seas. The disagreement in the simulations is probably due to inability of the  
823 coarse-resolution ICMMG to represent tracer advection by small-scale eddies that may play

824 an important role in distributing water from the currents following the margins of the sub-  
825 Arctic seas. Calculated eddy tracer fluxes demonstrated substantially higher eddy activity in  
826 the AO-HYCOM simulation, contrasting NEMO-LIM2 and ICMMG. Nevertheless,  
827 calculated eddy tracer flux into the Greenland and Iceland interior boxes did not demonstrate  
828 noticeable contribution of eddies to the total tracer flux. By contrast, eddy flux was a  
829 substantial contribution to the tracer flux into the interior Labrador Sea.

830 Modeled vertical distribution of tracer in the water column has a near-surface (top 200  
831 m) maximum in the sub-Arctic seas. Simulated vertical penetration of the tracer is deepest in  
832 the North Atlantic where the tracer spreads down to the near-bottom layers in AO-HYCOM  
833 and NEMO-LIM2 and down to ~2000m in ICMMG. The models disagree on the vertical  
834 tracer distribution in Baffin Bay where AO-HYCOM and ICMMG predicts tracer penetration  
835 down to ~800m and NEMO-LIM2 mixes the tracer all the way to the bottom. The causes of  
836 this disagreement are unclear and need further investigation.

837 Accumulation of the tracer in the interior regions of the sub-Arctic seas supports the  
838 idea of Greenland meltwater influence of the thermohaline processes and convection in the  
839 region. These results, however, are not directly supported by observations. No persistent  
840 negative salinity trends can be found in the reported hydrographic changes in the sub-Arctic  
841 region except for Baffin Bay. The estimates of the impact of the surplus Greenland  
842 freshwater flux on salinity changes in the upper 400 m layer suggest noticeable yet not  
843 dramatic freshening (from -0.04 to -0.12) of the sub-Arctic seas, especially in the interior  
844 regions. However, the magnitude of this freshening signal is smaller than the observed  
845 salinity increase in the Atlantic Water by 0.15–0.2 during the 2000s [Holliday *et al.*, 2008].  
846 The salinity increase in the Atlantic inflow can be linked to the propagating positive salinity  
847 anomaly from the subtropical Atlantic Ocean [Hátún *et al.*, 2005; Straneo and Heimbach,  
848 2013]. This positive salinity anomaly counteracts the freshening signal caused by Greenland

849 freshwater. The results of the model experiments also suggest that the accumulation of  
850 Greenland freshwater in the sub-Arctic seas caused by continuing Greenland Ice Sheet melt  
851 can amplify freshening in the Nordic Seas and the Labrador Sea during the period when  
852 salinity of the Atlantic inflow from southern North Atlantic decreases.

853 While the results here attest to efficient transport of freshwater derived from  
854 Greenland melt into the sub-Arctic seas, feedbacks and implications of this freshwater, as  
855 well as its relative importance with respect to influxes from the Arctic Ocean and Atlantic,  
856 remain an open question.

857

858

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870 study is that presented in *Bamber et al.* [2012] and is available on request as a gridded  
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881

## 881 **Appendix A. Characteristics of the numerical models**

### 882 *AI. AO-HYCOM*

883           The high resolution 0.08° regional Arctic Ocean HYbrid Coordinate Ocean Model  
884 (HYCOM) coupled to the Los Alamos National Laboratory Community Ice Code (CICE)  
885 [Hunke and Lipscomb, 2008] is used in this application (hereinafter referenced as AO-  
886 HYCOM). HYCOM is a primitive equation generalized coordinate (hybrid) ocean model  
887 [Bleck, 2002; Chassignet et al., 2006]. The model domain is a subset of the global HYCOM  
888 [Metzger et al., 2014] north of ~38°N. The computational grid of the AO-HYCOM is a  
889 Mercator projection from the southern boundary to 47°N. North of 47°N, it employs an  
890 orthogonal curvilinear Arctic dipole grid [Murray, 1996]. The model employs 32 hybrid  
891 vertical coordinate layers with potential density referenced to 2000m and includes the effect  
892 of thermobaricity [Chassignet et al., 2003]. The model is forced with atmospheric fields (2-m  
893 air temperature, 2-m atmospheric humidity, surface shortwave and longwave heat fluxes, and  
894 precipitation) that are derived from hourly fields of the Climate Forecast System Reanalysis  
895 (CFSR) [Saha et al., 2010]. Wind stress is estimated from the Cross-Calibrated Multi-  
896 Platform Ocean Surface Wind vector Analyses (CCMP) [Atlas et al., 2011]. The atmospheric  
897 fields employed for this experiment cover the period from 2004-2010 and have been recycled  
898 to provide forcing for 14 years. Surface latent and sensible heat fluxes, along with  
899 evaporation, are calculated using bulk formulas during model run time. The bulk transfer  
900 coefficients are parameterized following Kara et al.'s [2000] algorithm.

901           The model is initialized from an existing HYCOM-CICE simulation that was run with  
902 no Greenland runoff, and is integrated continuously for 14 years. Lateral open boundaries are  
903 derived from a climatology from the 0.08° Global HYCOM hindcast [Metzger et al., 2014].  
904 Thus, there is no interannual variability in the oceanic forcing superimposed at the open  
905 boundaries. It is noteworthy that surface salinity relaxation is minimal in the simulation with

906 the e-folding relaxation scale of  $1.59\text{e-}8 \text{ s}^{-1}$  corresponding to the restoring time scale of 4  
907  $\text{year}^{-1}$ .

908 Greenland freshwater sources are incorporated into AO-HYCOM using monthly  
909 inter-annual gridded data (Fig. 1b). The amount of tracer is proportional to the local  
910 freshwater flux rate and thus, tracer input replicates seasonal and interannual variability of the  
911 Greenland freshwater discharge. In HYCOM, runoff from a single source is distributed over  
912 several grid cells. From the beginning of the experiment, the passive tracer is continuously  
913 released in the upper 6m at every model grid cell that has non-zero Greenland freshwater  
914 influx along the Greenland coast. The tracer is prescribed in the model by relaxing tracer  
915 concentration in the specified locations, which are ocean grid cells nearest to the freshwater  
916 sources along the Greenland coast, to the maximum concentration value that is defined as  
917 follows. For the given Greenland freshwater flux at some location along the coast ( $F_r$ ,  $\text{m}^3/\text{s}$ )  
918 tracer concentration ( $\text{kg}/\text{m}^3$ ) in the grid cell is defined as

$$919 \quad C_{tr} = \frac{F_r \cdot t_{rlx} \cdot \rho_{tr}}{\Delta z \cdot \Delta x \cdot \Delta y}, \quad (\text{A1})$$

920 where  $F_r$  is Greenland runoff flux at a given location ( $\text{m}^3 \text{ s}^{-1}$ ),  $\rho_{tr}$  is tracer density ( $1000 \text{ kg m}^{-3}$ ),  
921  $t_{rlx}$  is tracer relaxation time scale (1 day, here) and  $\Delta z$ ,  $\Delta x$  and  $\Delta y$  are layer thickness and  
922 horizontal grid spacing (m). Tracers are released in the upper 2 layers in the grid cells where  
923 Greenland runoff is prescribed.

924

## 925 **A.2. NEMO-LIM2**

926 The Nucleus for European Modelling of the Ocean model (NEMO v3.4 – Madec,  
927 2008) is used in this experiment. The Arctic Northern Hemisphere Atlantic configuration  
928 (ANHA4) was based on the  $0.25^\circ$  tripolar grid extracted from the NEMO ORCA025  
929 configuration developed within the Mercator-Ocean and DRAKKAR collaboration [Barnier  
930 *et al.*, 2007]. The model consists of 50 vertical levels with a 1 m top layer decreasing in

931 resolution with increasing depth. The ANHA4 domain is contained within open boundaries at  
932 20°S latitude and Bering Strait. Lateral boundaries are free slip. Lateral mixing varies  
933 horizontally according to a bi-Laplacian operator with a horizontal eddy viscosity of  $1.5 \times$   
934  $10^{11} \text{ m}^4 \text{ s}^{-1}$ . For tracer lateral diffusion the model uses an isopycnal Laplacian operator with a  
935 horizontal eddy diffusivity of  $300 \text{ m}^2 \text{ s}^{-1}$ . Vertical mixing at sub-grid scales was  
936 parameterized using a turbulent kinetic energy (TKE) closure model [*Madec et al.*, 1998;  
937 *Axell*, 2002]. Background vertical eddy viscosity and diffusivity are  $10^{-4} \text{ m}^2 \text{ s}^{-1}$  and  $10^{-5} \text{ m}^2 \text{ s}^{-1}$   
938 respectively.

939 The sea ice module is from the Louvain-la-Neuve sea-ice model (LIM2) [*Fichefet and*  
940 *Maqueda*, 1997] with a modified elastic-viscous-plastic (EVP) ice rheology [*Hunke and*  
941 *Dukowicz*, 1997]. No-slip and free-slip boundary conditions are applied for sea ice and ocean,  
942 respectively.

943 The simulations presented here were forced with inter-annual atmospheric data  
944 derived from the Canadian Meteorological Centre's Global Deterministic Prediction System  
945 (CGRF, [*Smith et al.*, 2013]) with an hourly resolution in time and a spatial resolution of  
946  $0.45^\circ$  longitude and  $0.3^\circ$  latitude (minimal spacing is  $\sim 33$  km in the Labrador Sea).

947 The model is initialized with the output from the GLORYS2V3 reanalysis from  
948 MERCATOR and then run from 2002-2010 and then recycled over 2004-2010 to provide a  
949 14 year simulation. It is forced with monthly inter-annual runoff from *Dai et al.*, [2009].  
950 Greenland freshwater sources are incorporated using monthly inter-annual gridded data of  
951 [*Bamber et al.*, 2012]. Starting in January 2004, passive tracer is continuously released in the  
952 upper 10 m of the model at every freshwater source on the Greenland coast, with 5 tracers  
953 defined around the coasts of Greenland. The amount of tracer is proportional to the local  
954 freshwater discharge rate and thus, tracer flux replicates seasonal and inter-annual variability



955 of the Greenland freshwater discharge. The local change of tracer concentration in the model  
956 grid cells is defined as

957

$$958 \quad \Delta C_{tr} = \frac{F_r \Delta t}{\Delta z \Delta x \Delta y}. \quad (\text{A2})$$

959

### 960 **A.3. ICMMG**

961 A regional model of the Arctic and Atlantic Oceans of the Institute of Computational  
962 Mathematics and Mathematical Geophysics (ICCMG) is configured from 20°S in the Atlantic  
963 Ocean to 60°N in the Pacific Ocean [*Golubeva and Platov, 2007*]. The horizontal  
964 computational grid is bipolar curvilinear and has an equatorial resolution of 0.5° (minimal  
965 spacing is ~19km in the study region). The model uses a hydrostatic primitive formulation of  
966 Navier-Stokes equations with the rigid-lid approximation. The sea ice model is version 3 of  
967 CICE [*Hunke and Lipscomb, 2008*].

968 The model is forced by wind stress, sensible and latent heat fluxes, precipitation and  
969 evaporation, solar and longwave radiation derived from the NCEP/NCAR reanalysis  
970 [*Kanamitsu et al., 2002*]. Lateral open boundaries are provided by the PHC climatology  
971 ([http://psc.apl.washington.edu/nonwp\\_projects/PHC/Climatology.html](http://psc.apl.washington.edu/nonwp_projects/PHC/Climatology.html)). The experiment is  
972 initialized from an existing 1948-2003 model run with no Greenland runoff. Greenland  
973 freshwater input is implemented in the model as surface mass and salt fluxes at the ocean grid  
974 cells closest to the freshwater sources on the Greenland coast. Similar to the other model  
975 experiments, the amount of tracer is proportional to the local freshwater discharge rate.  
976 Tracers are prescribed as mass flux ( $\text{kg m}^{-2} \text{s}^{-1}$ ) into the near-surface layer calculated from the  
977 local runoff rate as

$$978 \quad F_{tr} = \frac{F_r \rho_{tr}}{\Delta x \Delta y}. \quad (\text{A3})$$

979

980

981 **Table A1. Characteristics of the numerical models**

	AO-HYCOM	NEMO-LIM2	ICMMG
Horizontal grid	Dipole curvilinear, 0.08°	Curvilinear, 0.25°	Dipole curvilinear, 0.5°
Vertical coordinates	32 hybrid layers	50 geopotential levels	38 geopotential levels
Free surface	Free surface, split time step	Linear filtered free surface <sup>1</sup>	Rigid-lid approximation
Baroclinic time step	240 s	1080 s	5400 s
Barotropic time step	7.5 s	none	Varying
Salinity relaxation	e-folding relaxation scale = $1.59 \times 10^{-8} \text{ s}^{-1}$ .	No relaxation	No relaxation
Bathymetry	DBDBV2 <sup>2</sup>	ETOPO1 <sup>3</sup> +GEBCOv1 <sup>4</sup>	2.5km IBCAO <sup>5</sup>
Scalar horizontal advection	Second-order flux-corrected transport	TVD <sup>6</sup>	Ultimate QUICKEST <sup>7</sup>
Horizontal diffusion	Laplacian diffusion = $u_d \times \Delta x$ , $u_d$ for scalars = 0.005 m/s, $u_d$ for momentum = $2.86 \times 10^{-3} \text{ m/s}$ . Biharmonic diffusion: $u_d \times \Delta x^3$ momentum dissipation = 0.03 m/s	Laplacian for tracer, maximum eddy diffusivity is $300 \text{ m}^2 \text{ s}^{-1}$ (proportional to grid size); Bilaplacian for momentum, maximum eddy viscosity is $-1.5 \times 10^1 \text{ m}^4/\text{s}$ (proportional to the cubic grid size)	Laplacian, $50 \text{ m}^2/\text{s}$
Vertical turbulence	KPP	TKE (turbulent kinetic energy) vertical mixing model	Richardson-based vertical mixing <sup>8</sup> and OPPS <sup>9</sup>
Diapycnal diffusivity	$1 \times 10^{-7} / \text{buoyancy freq.}, \text{ m}^2 \text{ s}^{-1}$	None	None
Background diffusivity	$0.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$	$0.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$	$0.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$
Background viscosity	$0.3 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$	$1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$	$1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$
Tracer input	Relaxed to tracer concentration based on local Greenland river runoff rate		Surface mass flux
Atmospheric forcing	CCMP winds and CFSR radiative fluxes	CGRF	NCEPR
Sea ice model	CICE v.4	LIM2	CICE v.3

982 <sup>1</sup> Roulet and Madec (2000)983 <sup>2</sup> Digital Bathymetric Data Base Variable Resolution (DBDBV2)984 <sup>3</sup> Global 1 min resolution relief dataset [*Amante and Eakins, 2009*]985 <sup>4</sup> General Bathymetric Chart of the Oceans986 <sup>5</sup> International Bathymetric Chart of the Arctic Ocean987 <sup>6</sup> Total Variation Dissipation scheme [*Lévy et al, 2001*]988 <sup>7</sup> [*Vested et al., 1992*]989 <sup>8</sup> [*Golubeva and Platov, 2007*]990 <sup>9</sup> Ocean penetrative plume scheme [*Paluszkiwicz and Romea, 1997*]

991

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1322 **Figure Captions**

1323 **Figure 1.** Map of the sub-Arctic seas with general circulation and geographic names  
1324 shown, including the regions selected for tracer budget analysis (section 4). Three main  
1325 basins include the Nordic Seas (NS), the Labrador Sea (LS), and Baffin Bay (BB). The boxes  
1326 are interior regions in the Greenland Sea (GS), Iceland Sea (IS), and interior Labrador Sea  
1327 (IL). The numbers indicate segments bounding the regions. Color coding is used for ease of  
1328 identification of the segments bounding the regions. The dashed black line across Davis Strait  
1329 shows the location of the 2004-present moored array. Abbreviated currents: CSC –  
1330 Continental Slope Current, FC – Faroe Current, SC – Shetland Current, NCC – Norwegian  
1331 Coastal Current, NwAC – Norwegian Atlantic Current, WSC – West Spitsbergen Current,  
1332 EGC – East Greenland Current, JMC – Jan Mayen Current, EIC – East Icelandic Current,  
1333 NIC – North Icelandic Current, WGC – West Greenland Current, BIC – Baffin Island  
1334 Current, LC – Labrador Current.

1335 **Figure 2.** Seasonal and interannual variability of Greenland freshwater fluxes from  
1336 *Bamber et al.*, [2012], 1990-2010. The fluxes include both solid and liquid discharge. (a)  
1337 Locations and volume fluxes ( $\text{km}^3 \text{ mo}^{-1}$ ) of Greenland freshwater sources along the coast.  
1338 The bar diagrams show monthly averaged Greenland freshwater fluxes between 1990-2010  
1339 ( $\text{km}^3 \text{ mo}^{-1}$ ) integrated over the regions delineated with the black lines. At the upper right,  
1340 monthly runoff integrated over the whole Greenland coast is shown. The vertical black lines  
1341 on top of the bars indicate the range of the monthly runoff during 1990-2010. Long-term  
1342 (1990-2010) mean runoff ( $\text{km}^3 \text{ yr}^{-1}$ ) is listed for every region. (b) Stacked bar diagram of the  
1343 annual Greenland freshwater flux in the regions. The black line depicts the total annual  
1344 freshwater flux.

1345 **Figure 3.** Mean salinity and salinity anomalies derived from the observational program in  
1346 Davis Strait data between 2004 and 2013 (the location of the moored array is shown in Figure  
1347 1). (a) Mean salinity. The major water masses are indicated: ArW – Arctic Water, TrW –  
1348 Transitional Water, WGIW – West Greenland Irminger Water, WGSW – West Greenland  
1349 Shelf Water. (b) Annual salinity anomalies ( $\Delta S$ ) relative to the 2004–2013 mean for the  
1350 individual water masses. The other water mass refers to the near surface water that is too  
1351 warm to be classified as ArW or is near surface water along the West Greenland slope right at  
1352 the shelf that is too salty to be WGSW and too fresh to be WGIW. (c) Annual salinity  
1353 anomalies relative to the 2003-2013 mean. (d) Time vs latitude diagram showing salinity  
1354 anomalies along the section shown with the red line on the map to the left for the time period  
1355 from 1990 to 2014 relative to the 1960-1990 monthly climatology. Salinity climatology and

1356 anomalies are derived from the Met Office Hadley Centre subsurface ocean salinity dataset  
1357 (EN.4.1.1).

1358 **Figure 4.** Time-depth diagrams of salinity in the central Greenland Sea (a) and Iceland  
1359 Sea (b) from hydrographic observations in boxes GS and IS shown in Figure 1.  
1360 Hydrographic measurements (salinity, temperature, and pressure from bottle and CTD data)  
1361 in the Greenland and Iceland seas from 1950 to 2010 are from the ICES Dataset on Ocean  
1362 Hydrography ([www.ocean.ices.dk](http://www.ocean.ices.dk)) and the Pangaea database ([www.pangaea.de](http://www.pangaea.de)). Note a very  
1363 shallow (upper 25 m) freshening signal at both locations.

1364 **Figure 5.** Daily mean tracer concentration in the first model layer (~3m) on December  
1365 17<sup>th</sup> for every model year from AO-HYCOM. The concentration is on the natural-log scale  
1366 ( $\text{kg m}^{-3}$ ).

1367 **Figure 6.** Five-day mean tracer concentration in the surface model layer (~1m) for output  
1368 including December 17<sup>th</sup> for every model year from NEMO-LIM2. The concentration is on  
1369 the natural-log scale ( $\text{kg m}^{-3}$ ).

1370 **Figure 7.** Daily mean tracer concentration in the surface model layer (~2m) on December  
1371 17<sup>th</sup> for every model year from ICMMG. The concentration is on the natural- log scale ( $\text{kg m}^{-3}$ ).  
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1373 **Figure 8.** Tracer budget for the Nordic Seas region. (a) Volume-integrated tracer content  
1374 (kg). Colors designate different models. (b—d) 60-day low-pass filtered time series of the  
1375 tracer fluxes ( $\text{kg s}^{-1}$ ) across the sections from the model experiments. Colors designate fluxes  
1376 across individual sections shown in the inset in (b). Section numbers correspond to Figure 1.  
1377 Note the different scale in 8b vs 8c and 8d.

1378 **Figure 9.** Same as Figure 8 but for the Greenland Sea region. Note the different scales in  
1379 9b–9d.

1380 **Figure 10.** . Same as Figure 8 but for the Iceland Sea region. Note the different scales in  
1381 10b–10d.

1382 **Figure 11.** Same as Figure 8 but for the Labrador Sea region.

1383 **Figure 12.** Same as Figure 8 but for the interior Labrador Sea region.

1384 **Figure 13.** Same as Figure 8 but for Baffin Bay. (d) In ICMMG, section 22 is over land,  
1385 thus the flux is not shown.

1386 **Figure 14.** Mass fraction (%) of the volume-integrated tracer in 3 regions (NS, LS, and  
1387 BB) relative to the total tracer mass integrated over the model domain (limited by ~38°N for  
1388 NEMO-LIM2 and ICMMG). (a) AO-HYCOM; (b) NEMO-LIM2; (c) ICMMG. Shown are  
1389 mass fraction values after 3, 6, 9, 12, and 14 years of the simulation. The colored bars inside

1390 the “Nordic Seas” and “Labrador Sea” illustrate the mass fraction of the tracer integrated over  
1391 the interior boxes (blue - GS, red - IS, and green - IL) relative to the total tracer mass (the  
1392 values are listed in colored numbers). Also listed is the total mass fraction of the tracer within  
1393 the three regions.

1394 **Figure 15.** Distribution of the tracer concentration ( $\text{kg m}^{-3}$ ) along the vertical section (g)  
1395 from the numerical experiments on December 17<sup>th</sup> after 7 and 14 years. In a–f, the vertical  
1396 axis is depth in meters. (a, b) AO-HYCOM: black contours are interfaces of the vertical  
1397 layers from a single model output time. (c) NEMO-LIM2. (e, f) ICMMG. (g) Section lines.  
1398 The red numbers on the blue line correspond to the distances (km) along the line also shown  
1399 on the horizontal axis on the diagrams.

1400 **Figure 16.** Mean (top row) and eddy (bottom row) tracer fluxes ( $\text{kg s}^{-1} \text{m}^{-1}$ ) integrated  
1401 over the upper 150 m calculated from daily-mean fields (a, b) AO-HYCOM, (c, d) NEMO-  
1402 LIM2, and (e, f) ICMMG for year 9. The fluxes are one-year-averaged.

1403 **Figure 17.** Mass gain (time integrated net tracer flux, kg) from the mean tracer flux (blue)  
1404 and eddy tracer flux (red) into the interior convective regions (Figure 1) during years 8–12 of  
1405 the simulations from (a–c) AO-HYCOM, (d–f) NEMO-LIM2, and (g–i) ICMMG. The  
1406 horizontal axis is time (model years).

1407 **Figure 18.** Hydrographic changes in the sub-Arctic regions (in rows) from AO-HYCOM.  
1408 (a) Time series of freshwater content (black) and salinity (blue) in the upper 400 m during the  
1409 simulation. (b) Salinity change of the upper 400m vs cumulative time-integrated Greenland  
1410 freshwater flux. Linear regression parameters (Eq. 5) and coefficient of determination are  
1411 listed. (c) Estimated salinity change in the upper 400m for time-integrated anomaly of  
1412 Greenland freshwater flux for 1990-2010.

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1413 **Figures**  
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