

1 **Arctic sea level variability from high-resolution model simula-**  
2 **tions and implications for the Arctic observing system**

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8 **Abstract.** Two high-resolution model simulations are used to investigate the spatio-temporal variability of the  
9 Arctic Ocean sea level. The model simulations reveal barotropic sea level variability at periods <30 days, which  
10 is strongly captured by bottom pressure observations. The seasonal sea level variability is driven by volume  
11 exchanges with the Pacific and Atlantic Oceans and the redistribution of the water by the wind. Halosteric ef-  
12 fects due to river runoff and evaporation minus precipitation (EmPmR), ice melting/formation also contribute in  
13 the marginal seas and seasonal sea ice extent regions. In the central Arctic Ocean, especially the Canadian Basin,  
14 the decadal halosteric effect dominates sea level variability. **The study confirms that satellite altimetric observa-**  
15 **tions and Gravity Recovery and Climate Experiment (GRACE) could infer the total freshwater content changes**  
16 **in the Canadian Basin at periods longer than one year, but they are unable to depict the seasonal and subseasonal**  
17 **freshwater content changes.** The increasing number of profiles seems to capture freshwater content changes  
18 since 2007, encouraging further data synthesis work with a more complicated interpolation method. Further, in-  
19 situ hydrographic observations should be enhanced to reveal the freshwater budget and close the gaps between  
20 satellite altimetry and GRACE, especially in the marginal seas.

## 21 **1 Introduction**

22 The Arctic Ocean is experiencing pronounced changes (e.g., Perovich et al., 2020; AMAP, 2019). Observa-  
23 tions have revealed increased warm inflows through the Bering Strait (Woodgate et al., 2012) and the Fram  
24 Strait (Polyakov et al., 2017), and an unprecedented freshening of the Canadian Basin, especially the Beaufort  
25 Gyre (Proshutinsky et al., 2019). The rapid changes potentially impact the weather and climate of the northern  
26 hemisphere (Overland et al., 2021).

27 As an integrated indicator, sea level change reflects changing ocean conditions caused by ocean dynamics,  
28 atmospheric forcing, and terrestrial processes (Stammer et al., 2013). Satellite altimetry, together with bottom  
29 pressure observations from Gravity Recovery and Climate Experiment (GRACE), has been applied to infer  
30 ocean temperature and salinity changes that are not measured directly in the Arctic Ocean (e.g., Armitage et al.,  
31 2016) and in the deep ocean (e.g., Llovel et al., 2014), enhancing our ability to monitor ocean changes.

32 Over the past decades, coupled ocean-sea ice models and observations have advanced our understanding of  
33 the Arctic Ocean variability. Proshutinsky and Johnson (1997) demonstrated wind-forced cyclonic/anticyclonic  
34 ocean circulation patterns accompanied by dome-shaped sea levels variation using a barotropic model simula-  
35 tion. Further, ocean circulation changes in the Canadian Basin result in freshwater accumulation and release,  
36 which is very well correlated to sea level changes (Koldunov et al., 2014; Proshutinsky et al., 2002). Given that  
37 sea level changes reflect freshwater content changes in the Canadian Basin, Giles et al. (2012) and Morison et al.  
38 (2012) proposed to use satellite altimetry observations and GRACE observations to infer freshwater content  
39 changes. The method was then applied to explore the freshwater content changes in the Beaufort Gyre  
40 (Armitage et al., 2016; Proshutinsky et al., 2019) at seasonal to decadal timescales. In the Barents Sea, Volkov et  
41 al. (2013) used altimetric sea level observations and the ECCO reanalysis (Forget et al., 2015) to explore sea-  
42 sonal to interannual sea level anomalies, revealing different roles of mass-related changes, thermosteric and  
43 halosteric effects on different regions of the Barents Sea.

44 However, the sparseness of in-situ profiles, coarse resolution and significant uncertainties of satellite altim-  
45 etry and GRACE observations result in large gaps in understanding the spatio-temporal variability of the Arctic  
46 sea level and its relations to the thermo/halosteric effects and mass changes (Ludwigsen and Andersen, 2021).  
47 Previous studies mainly focus on the decadal sea level variability (e.g., Koldunov et al., 2014; Proshutinsky et al.,  
48 2007; Proshutinsky and Johnson, 1997), and no study has yet fully explored the Arctic sea level variability at  
49 different spectral bands, and its dependence on the mass component and the vertical oceanic variability. Such a  
50 study could help identify critical regions and environmental parameters that need to be observed coordinately  
51 and point out observational gaps that need to be filled in the future.

52 Our study systematically explores the Arctic sea level variability as a function of timescale and geographic  
53 location using daily and monthly outputs of two high-resolution model simulations. Contributions from  
54 barotropic changes expressed in bottom pressure variations and baroclinic processes represented by  
55 thermo/halosteric changes are quantified at different timescales. **Altimetric and GRACE measurements, in-situ**  
56 **hydrographic observations mapped with different interpolation schemes (e.g., Haine et al., 2015; Polyakov et al.,**  
57 **2008; Rabe et al., 2014; Rabe et al., 2011), and ocean reanalyses have been used to infer the basin-scale**  
58 **freshwater changes during the unprecedented freshwater changes since the 2000s. However, Solomon et al.**  
59 **(2021) pointed out that significant uncertainties and discrepancies remain in revealing the regional patterns. This**  
60 **study further discusses the existing Arctic observing system's capability to monitor the Arctic freshwater content**

61 **variability and identify observational gaps.**

62 The structure of the remaining paper is as follows: the numerical models and the observations from the bot-  
63 tom pressure sensor, GRACE, and satellite altimetry are described in Section 2, together with different compo-  
64 nents of sea level changes. We compare the model simulations against observations in Section 3. Section 4 ana-  
65 lyzes sea level variability and associated mechanisms at high frequency (<30 days), seasonal cycles, and decadal  
66 timescales. The relations with bottom pressure and thermos/halosteric components are demonstrated, pointing  
67 out key regions and parameters we need to observe. Further, we analyze the ability of satellite altimetry,  
68 GRACE, and the in-situ profiler system to monitor the Arctic freshwater content variability in Section 5. Sec-  
69 tion 6 provides a summary and conclusions.

## 70 **2 Model Simulations and observations**

### 71 **2.1 Atlantic-Arctic simulations**

72 This study relies on two ocean high-resolution numerical simulations using the MIT general circulation  
73 model (Marshall et al., 1997). A dynamic thermodynamic sea ice model (Hibler, 1979, 1980;Zhang and  
74 Rothrock, 2000), implemented by Losch et al. (2010), is employed to simulate sea ice processes. The model  
75 domain covers the entire Arctic Ocean north of the Bering Strait and the Atlantic Ocean north of 33°S. In the  
76 horizontal, the model uses a curvilinear grid with resolutions of ~8 km (ATLARC08km) and ~4 km (AT-  
77 LARC04km). **In the vertical, ATLARC08km has 50 levels with resolution ranging from 10 m over the top 130**  
78 **m to 456.5 m in the deep basin. And ATLARC04km has 100 z-levels ranging from 5 m over the top 200 m to**  
79 **185 m in the deep basin.**

80 At the ocean surface, the model simulations are forced by momentum, heat, and freshwater fluxes comput-  
81 ed using bulk formulae and either the 6-hourly NCEP RA1 reanalysis (Kalnay et al., 1996) (ATLARC08km) or  
82 the 6-hourly ECMWF ERA-Interim reanalysis (Dee et al., 2011) (ATLARC04km). A virtual salt flux parame-  
83 terization is used to mimic the dilution and salinification effects of rainfall, evaporation, and river discharge.  
84 The models are forced by the monthly output from the GECCO2 (Köhl, 2015) global model configuration at the  
85 open boundaries. The river runoff is applied at river mouths by seasonal climatology (Fekete et al., 2002). Bot-  
86 tom topography is derived from the ETOPO 2-min (Smith and Sandwell, 1997) database. ATLARC08km is  
87 initialized with annual mean temperature and salinity from the World Ocean Atlas 2005 (Boyer et al., 2005) and  
88 covers 1948 to 2016, and ATLARC04km starts from the initial condition, **including velocity, temperature, and**  
89 **salinity**, of ATLARC08km at the start of the year 2002. Table 1 summarizes both the simulations and their main  
90 characteristics.

91 Table 1. Summary of model simulations used in this study.

	<i>Horizontal resolution</i>	<i>Vertical grid</i>	<i>Surface forcing</i>	<i>periods</i>	<i>Output Frequency</i>	<i>Variables used</i>
<i>ATLARC08km</i>	~8 km	50 z-levels	NCEP-RA1	1948-2016 05.01.2003- 01.12.2010	monthly daily	<b>Potential temperature</b> <b>Salinity</b> <b>Sea surface height</b>

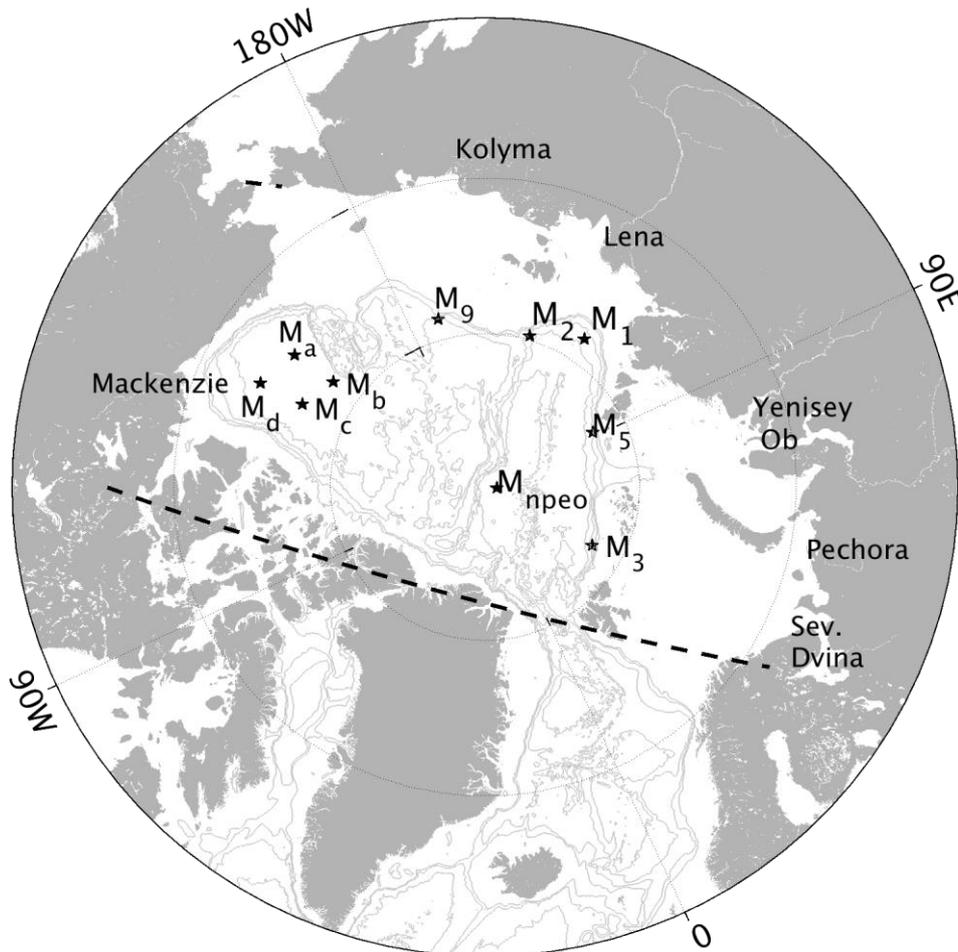
ATLARC04km	~4 km	100 z-levels	ERA-Interim	01.01.2003-23.08.2012	daily	Wind stress
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93 **2.2 Satellite and in-situ observations**

94 Koldunov et al. (2014) have validated ATLARC08km against tide gauge observations. We further compare  
 95 the two model simulations against in-situ bottom pressure observations, GRACE observations, and satellite  
 96 altimetric observations.

97 The monthly altimetric sea level observations from Armitage et al. (2016,  
 98 [http://www.cpom.ucl.ac.uk/dynamic\\_topography/](http://www.cpom.ucl.ac.uk/dynamic_topography/)) and GRACE measurements (Chambers and Bonin, 2012,  
 99 [doi:10.5067/TEOCN-3AJ64](https://doi.org/10.5067/TEOCN-3AJ64)) are used in comparison with the model simulations. For the very high-frequency  
 100 variability, bottom pressure records supplied by the Beaufort Gyre Exploration Project (BGEP, M<sub>a</sub>, M<sub>b</sub>, M<sub>c</sub>, and  
 101 M<sub>d</sub> in Fig. 1, <https://www2.whoi.edu/site/beaufortgyre/data/mooring-data/>) and the North Pole Environmental  
 102 Observatory (NPEO, M<sub>npeo</sub> in Fig.1, [ftp://northpoleftp.apl.washington.edu/NPEO\\_Data\\_Archive/](ftp://northpoleftp.apl.washington.edu/NPEO_Data_Archive/)) are used.  
 103 Tidal signals are removed using the T\_TIDE Matlab program (Pawlowicz et al., 2002) since the model did not  
 104 include tidal forcing.



105

106 **Figure 1.** A map of the pan-Arctic Ocean presenting the locations of moorings deployed by the Nansen and  
 107 Amundsen Basin Observational System (NABOS, black pentagrams labeled with M<sub>1</sub>, M<sub>2</sub>, M<sub>3</sub>, M<sub>5</sub>, and M<sub>9</sub>), by  
 108 the Beaufort Gyre Exploration Project (BGEP, black pentagrams marked with M<sub>a</sub>, M<sub>b</sub>, M<sub>c</sub>, M<sub>d</sub>), and by the

109 North Pole Environmental Observatory (NPEO, black pentagram labeled with  $M_{\text{npeo}}$ ). **The black dashed lines**  
 110 **enclose the Arctic regions used in the following sections.** Bathymetry contours of 500, 1000, 2000, 3000, and  
 111 4000 m are drawn with grey lines. Main rivers are labeled with their names near the river mouths.

### 112 **2.3 Relation between sea level, bottom pressure, and thermo/halosteric components**

113 Following Ponte (1999) and Calafat et al. (2013), sea level anomaly  $\eta'$ , can be separated into a steric com-  
 114 ponent  $\eta'_s$  due to density change, an inverse barometer effect  $\eta'_{IB}$ , and a mass (measured by bottom pressure  
 115 observations) component  $\eta'_m$ :

$$116 \eta' = -\frac{1}{\rho_0} \int_{-H}^0 \rho' dz + \frac{1}{\rho_0 g} (\bar{P}'_a - P'_a) + \frac{1}{\rho_0 g} (P'_b - \bar{P}'_a) \quad (1)$$

117 where  $g=9.8 \text{ m s}^{-2}$  is the gravitational acceleration. The first term on the right-hand side represents the steric  
 118 effect  $\eta'_s$ , with  $\rho_0$  being a reference density ( $1025.0 \text{ kg/m}^3$  in this study) and  $\rho'$  being the density change. The  
 119 second term is the inverse barometer effect  $\eta'_{IB}$ :  $\bar{P}'_a$  and  $P'_a$  represent air pressure anomalies average over the  
 120 global ocean and at the observing location, respectively. The last term defines the mass component  $\eta'_m$ .  **$P'_b$  is the**  
 121 **bottom pressure anomalies in equivalent meters of water.**

122 Since the model simulations do not include the impacts of surface air pressure anomalies, the model-  
 123 simulated sea level changes due to steric and mass components are simplified as:

$$124 \eta' = -\frac{1}{\rho_0} \int_{-H}^0 \rho' dz + \frac{1}{\rho_0 g} (P'_b) \quad (2).$$

125 Separating density changes into temperature and salinity changes, we decompose the steric height  $\eta'_s$  into  
 126 thermosteric height  $\eta'_{st}$  (due to temperature anomalies) and halosteric height  $\eta'_{ss}$  (due to salinity anomalies):

$$127 \eta'_{st} = -\frac{1}{\rho_0} \int_{-H}^0 (\rho(T, \bar{S}, p) - \rho(\bar{T}, \bar{S}, p)) dz \quad (3),$$

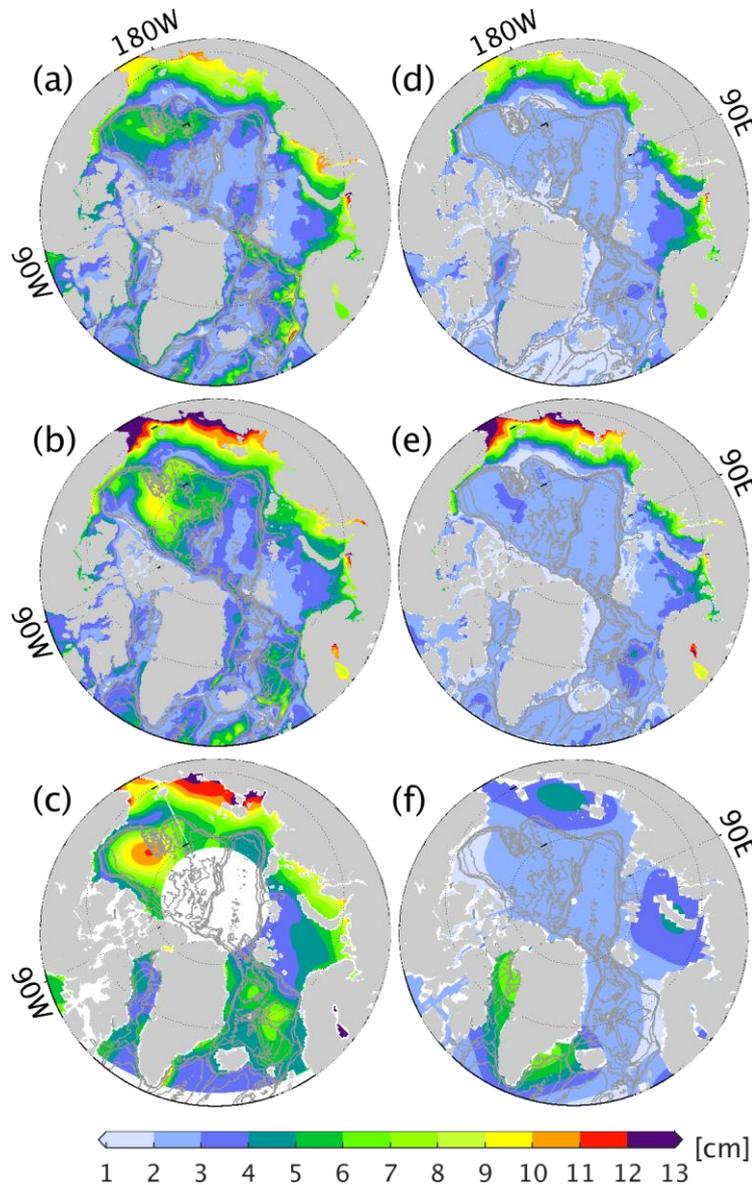
$$128 \eta'_{ss} = -\frac{1}{\rho_0} \int_{-H}^0 (\rho(\bar{T}, S, p) - \rho(\bar{T}, \bar{S}, p)) dz \quad (4),$$

129 where  $T$ ,  $S$ , and  $p$  represent seawater temperature, salinity, and pressure. The overbars denote the average over  
 130 the simulation time.

131 Before comparing the model simulation with the GRACE measurements and mooring-based bottom pres-  
 132 sure observations, we remove air pressure anomalies averaged over the global ocean  $\bar{P}'_a$ , and then global-mean  
 133 mass changes from GRACE-based bottom pressure observations since the virtual salt flux parameterization does  
 134 not include mass transfer from land to ocean. In total, this process removes a seasonal cycle with an amplitude  
 135 of  $\sim 1$ - $1.5 \text{ cm}$  from the measurements.

### 136 **3 Testing simulations against observations**

137 Koldunov et al. (2014) have demonstrated that the interannual sea level variability in ATLARC08km and  
 138 tide gauges match very well. In the present study, we further evaluate the skill of the model-simulated sea level  
 139 and bottom pressure variability by comparing the root mean square (RMS) variability of sea level and bottom  
 140 pressure against altimetric data (Armitage et al., 2016) and GRACE data (Chambers and Willis, 2010). In addi-  
 141 tion, high-frequency bottom pressure observations from BGEP and NPEO are compared with the two model  
 142 simulations.



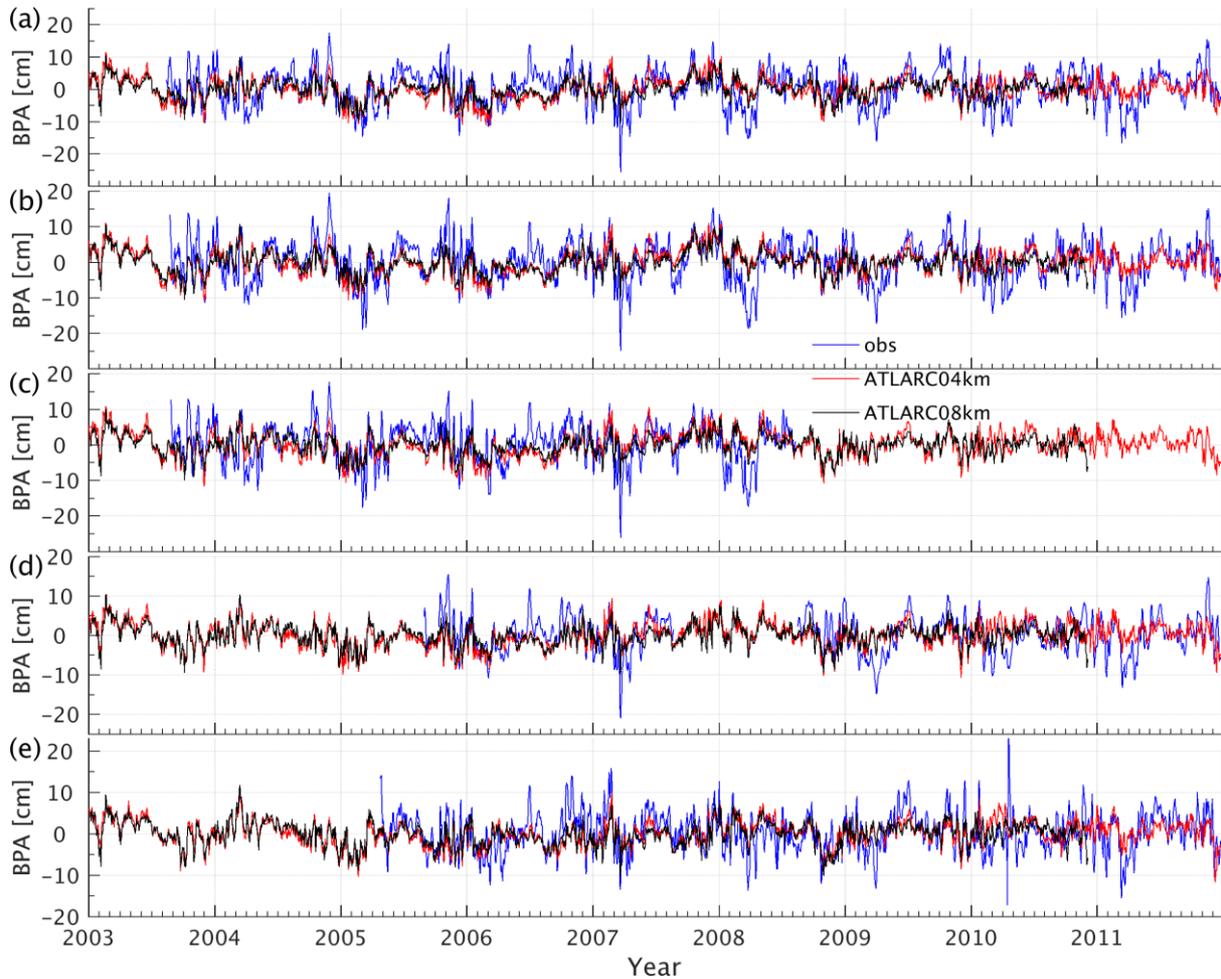
143

144 **Figure 2.** RMS variability of (a-c) sea level and (d-f) bottom pressure in (a, d) ATLARC08km, (b, e) ATLARC04km, (c)  
 145 satellite altimetry, and (f) GRACE. We computed the RMS variability using monthly data from January 2003 to December  
 146 2011. Bathymetry contours of 500, 1000, 2000, and 3000 m are drawn with grey lines.

147 The model simulations (Figs. 2a, b) and satellite altimetry (Fig. 2c) reveal pronounced sea level variability  
 148 in the Canadian Basin and along the coast. In the Canadian Basin, where a characteristic scale of the Rossby  
 149 radius is  $\sim 10\text{--}15$  km (Nurser and Bacon, 2014), ATLARC04km starts to resolve transient eddies and thereby  
 150 simulates more significant sea level variability than ATLARC08km, and matches better with the observed sea  
 151 level variability. Still, ATLARC04km and ATLARC08km underestimate the observed sea level variability in  
 152 the Canadian Basin. Along the Arctic coast, the pronounced sea level variability is related to the seasonal river  
 153 runoff, the redistribution of water due to the shifting of basin-scale cyclonic/anticyclonic wind (Proshutinsky  
 154 and Johnson, 1997). Again, ATLARC04km simulates much stronger sea level variability than ATLARC08km  
 155 and is comparable to the altimetric observations. Bottom pressure shows significant variability in the Arctic  
 156 marginal seas (Figs. 2d-f), especially in the East Siberian Sea. However, due to the smoothing process applied  
 157 on GRACE measurements (a 500 km Gaussian filter), both the model simulations simulate much stronger RMS

158 variability of bottom pressure. The coarse GRACE resolution, uncertainties in the altimetric measurements, and  
159 a lack of in-situ hydrographic observations results in gaps in closing the budget of sea level trend and changes,  
160 especially in the Kara, Laptev, and the East Siberian seas (Ludwigsen and Andersen, 2021) where in-situ hydro-  
161 graphic data are rare and altimetric measurements are less correlated with tide gauge data (Armitage et al.,  
162 2016).

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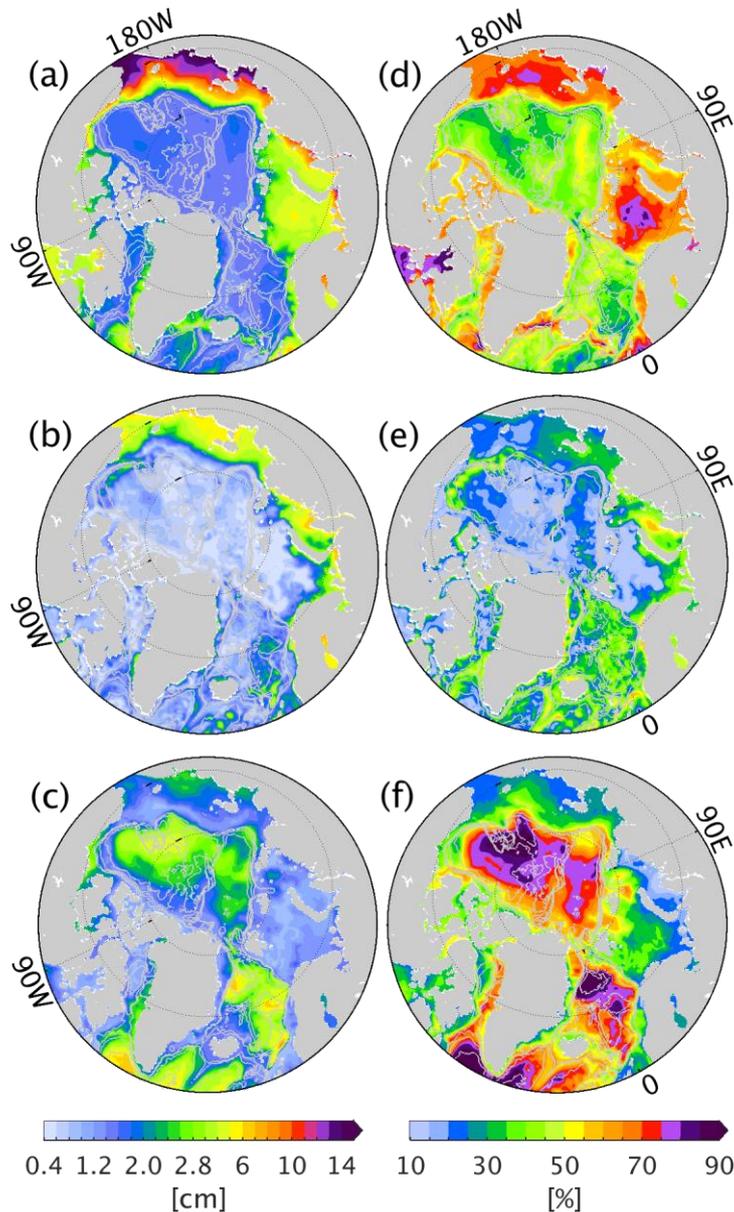
165 Figure 3. Time series of bottom pressure anomalies in ATLARC08km, ATLARC04km, and in-situ observations.  
166 Observations are derived from (a) mooring A, (b) mooring B, (c) mooring C, and (d) mooring D of BGEP. Panel  
167 (e) is from the NEPO moorings. Mooring locations are marked in Fig. 1.

168 Besides monthly to decadal variability of bottom pressure, both the model simulations and the in-situ ob-  
169 servations also demonstrate significant high-frequency bottom pressure anomalies (Fig. 3). Both model simula-  
170 tions correlate well with the observations ( $\sim 0.45-0.55$ ) in the five shown locations, but ATLARC04km and  
171 ATLARC08km underestimate the RMS variability by  $\sim 30-50\%$ , with ATLARC04km showing **relatively**  
172 **stronger** RMS variability.

173 The comparisons above indicate that the model simulations reasonably reproduce the observed sea level  
174 and bottom pressure variability at both high-frequency and low-frequency bands. In the following parts, we will  
175 use the daily output of ATLARC04km to reveal spatial variability of sea level at high frequency and seasonal  
176 periods and use the monthly output of ATLARC08km to explore the decadal sea level variability.

177 **4 Sea level variability and its relation with bottom pressure and steric height**

178 A model study (Proshutinsky et al., 2007) and satellite observations (Armitage et al., 2016) showed that the  
 179 Arctic sea level presents distinctive seasonal to decadal variability. In situ bottom pressure observations also  
 180 reveal energetic variability at sub-monthly frequencies. Here, we concentrate on sea level variability at very  
 181 high-frequency (<30 days), on the seasonal cycle, and at decadal timescales (>4 years).



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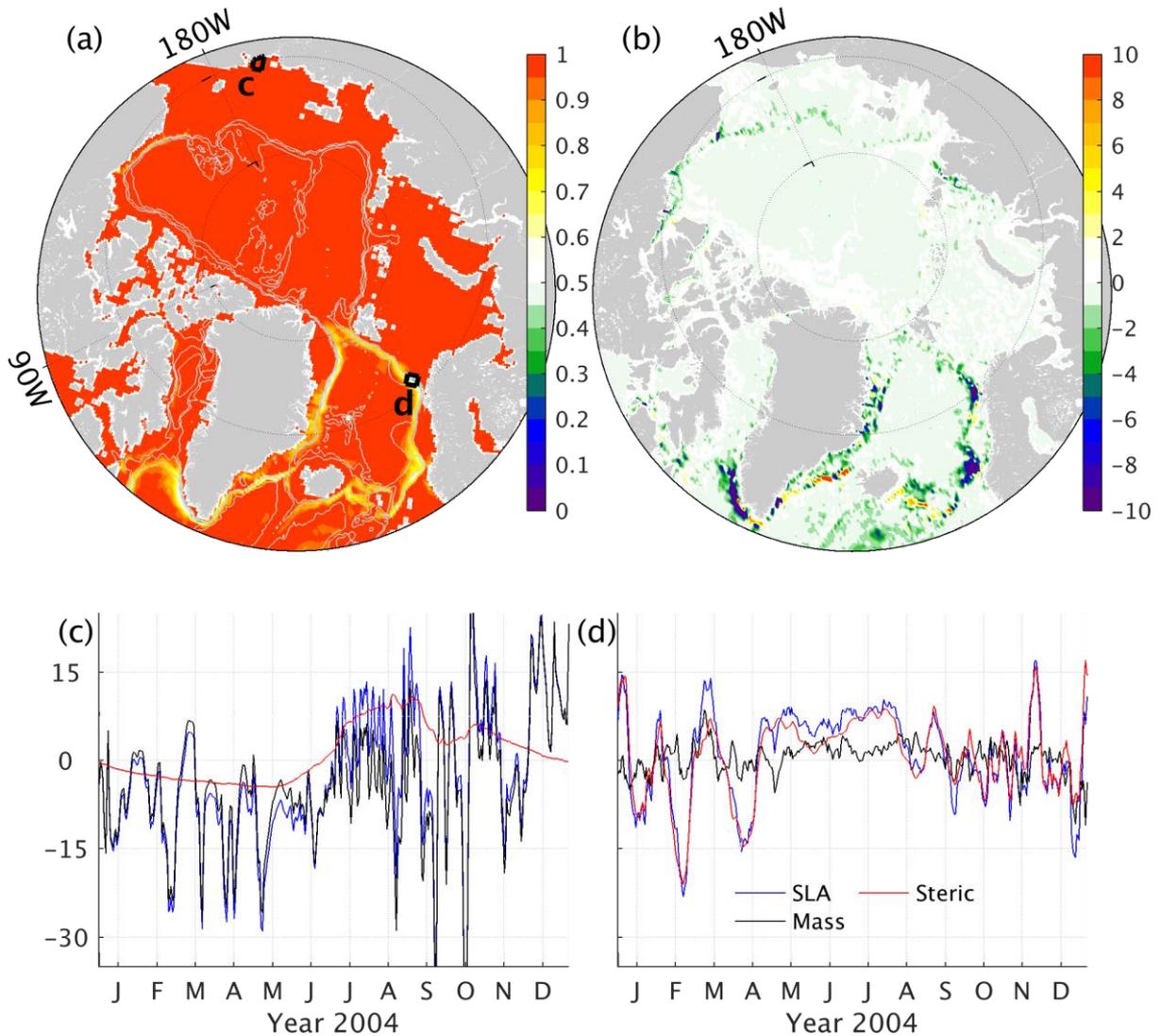
183 Figure 4. RMS variability (cm) of sea level (a) in the high-frequency band (<30 days), (b) at the seasonal cycle,  
 184 and (c) at decadal periods (>4 years). Panels (d)-(f) are the corresponding ratios (%) to the total sea level vari-  
 185 ance that panels (a)-(c) explained. **The high-frequency and seasonal variability (a, b, d, and e) uses the daily**  
 186 **output of ATLARC04km, and decadal variability (c and f) uses the monthly output from ATLARC08km.** The  
 187 grey lines denote bathymetry contours of 500, 1000, 2000, and 3000 m.

188 At period <30 days, RMS variability of sea level up to 14 cm appears in the marginal seas and along the  
 189 coasts (Fig. 4a), accounting for 60%~80% of the local sea level variance (Fig. 4d). The seasonal sea level varia-

190 bility is pronounced in the marginal seas and southern edge of the Beaufort Sea, and it explains 20%-40% of the  
 191 total sea level variance. In the deep regions of the pan-Arctic Ocean, the decadal variability dominates the sea  
 192 level variability, and it explains more than 70%~90% of the sea level variability. Overall, in the marginal seas,  
 193 sea level variability is dominated by sub-monthly and seasonal signals. In contrast, decadal sea level variability  
 194 dominates in the deep regions of the pan-Arctic Ocean. Besides, seasonal variability is also visible in the south-  
 195 ern periphery of the Beaufort Sea, indicating possible exchanges between the marginal seas and the Beaufort  
 196 Sea.

197 **4.1 High-frequency (<30 days) variability**

198 With a coarse resolution model simulation, Vinogradova et al. (2007) demonstrated that sea level variabil-  
 199 ity is coherent with and virtually equivalent to bottom pressure in the mid-latitude and subpolar regions at peri-  
 200 ods <100 days, reflecting the barotropic nature of high-frequency variability (Stammer et al., 2000). Here, we  
 201 revisit the high-frequency sea level variability in the pan-Arctic Ocean with high-resolution model simulations  
 202 and a transfer function (Vinogradova et al., 2007) of sea level and bottom pressure.



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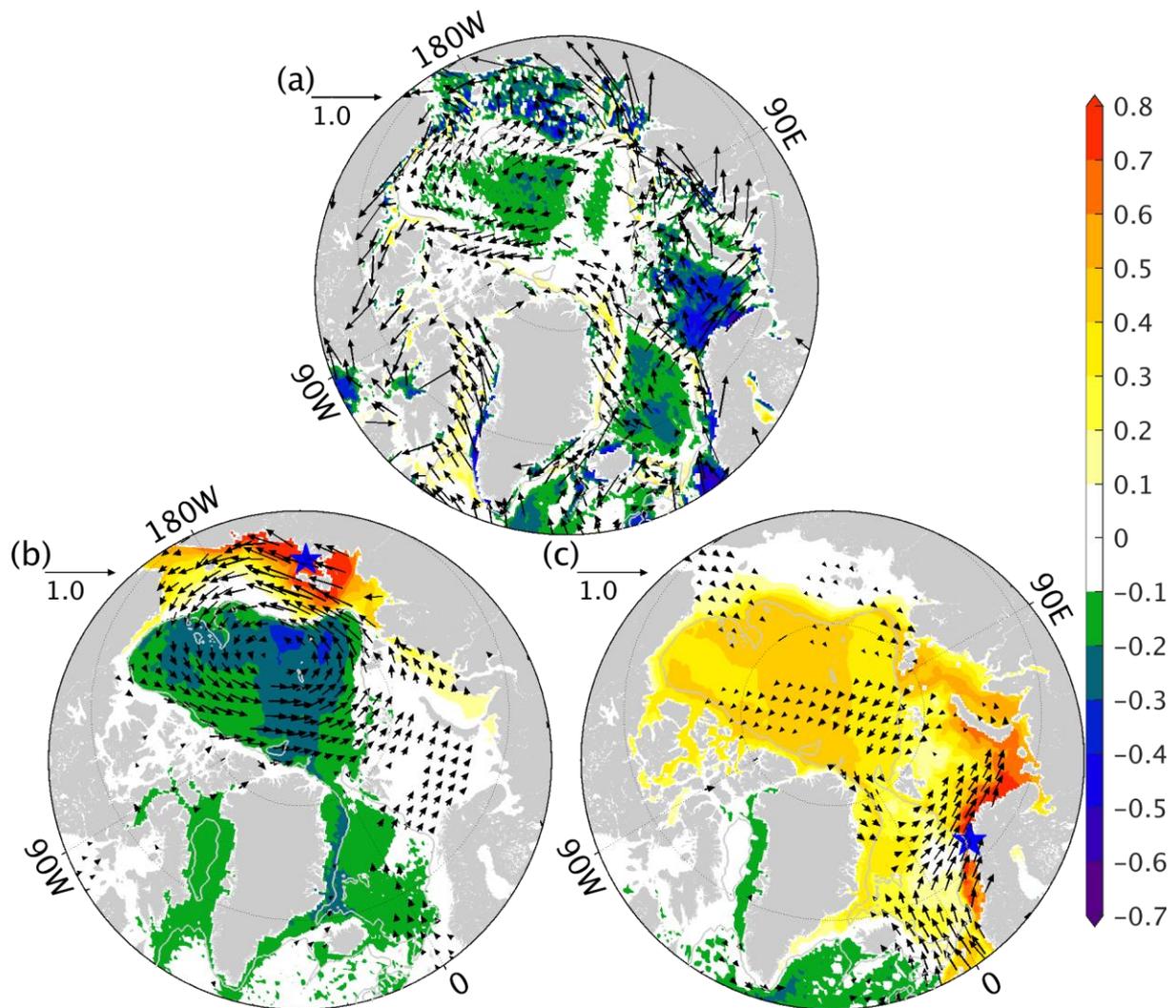
204 Figure 5. (a) Amplitude and (b) phase of the transfer function between sea level anomaly and bottom pressure  
205 anomaly at periods <30 days. Time series of sea level anomaly (blue lines), mass component (black lines), and  
206 steric component (red lines) averaged (c) in the East Siberian Sea (black box c in panel (a)) and (d) along the  
207 NwAC (black box d in panel (a)).

208 Except for the Norwegian Atlantic Current (NwAC) and the East/West Greenland Current (EGC/WGC),  
209 the amplitude of the transfer function between sea level and mass component is  $\sim 1$  (Fig. 5a) in most of the pan-  
210 Arctic regions. The phases (Fig. 5b) are  $\sim 0$  in the entire Arctic Ocean, indicating that the high-frequency sea  
211 level variability is mostly barotropic. However, in the strong current regions, including NwAC, EGC, and WGC,  
212 an amplitude of the transfer function of  $\sim 0.4$  is observed, revealing that both barotropic and baroclinic processes  
213 contribute to the high-frequency sea level variability.

214 Sub-regions in the East Siberian Sea (c in Fig. 5a) near the maximum RMS variability and along the  
215 NwAC (d in Fig. 5a) are used to reveal details of the high-frequency sea level variability. It is clear that the sea  
216 level anomaly in the East Siberian Sea (Fig. 5c) is almost equivalent to the bottom pressure anomaly, and the  
217 steric component contributes slightly to the seasonal timescale. Along the NwAC (Fig. 5d), pronounced steric  
218 height variability with timescales of 20-60 days is visible, which may be caused by baroclinic instability, and  
219 the mass component shows high-frequency variability.

220 The high-frequency sea level variability is mainly related to wind forcing (Fukumori et al., 1998) at high  
221 latitudes. Correlations to the wind forcing and sea level anomalies are used to explain the driving mechanisms of  
222 the high-frequency SLA variability. The negative correlations between high-frequency sea level variability and  
223 wind stress curl (shading in Fig. 6a) in the Canadian Basin and GIN seas ( $-0.3$ ) and in the marginal seas ( $-0.3$ ~  
224  $0.5$ ) reveal that local sea level increase/decrease is partially related to wind-induced convergence/divergence  
225 (vectors and shading in Fig. 6a) of water. In addition, the high correlations of SLA to wind stress (vectors in Fig.  
226 6a) along the coast reveal that cyclonic along-shore wind distributes water to the coast through Ekman transport,  
227 increasing sea level there.

228 To further explore the propagating features of the strong SLA variability along the coasts, we show  
229 correlations of SLA in sub-regions of the East Siberian Sea (blue pentagon Fig. 6b) and Norwegian coast (blue  
230 pentagon Fig. 6c) to SLA (shading), and wind stress (vectors). Fig. 6b demonstrates that anticyclonic wind  
231 stress distributes water to the coast through Ekman transport which interacts with topography, rising coastal sea  
232 levels. SLA in the Norwegian coast is also driven by along-shore wind (vectors in Fig. 6c) through Ekman  
233 transport, and the SLA signals propagate along the coast to the Barents Sea and the central Arctic Ocean  
234 (shading in Fig. 6c).



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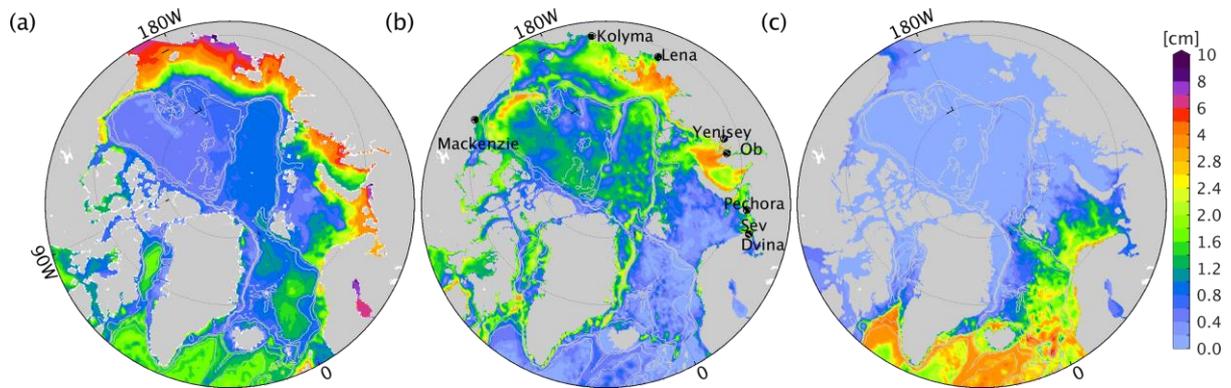
237 Figure 6. (a) Coefficients of the correlation between sea level anomalies and wind stress curl (shading), wind  
 238 stress (vectors) at periods  $< 30$  days. (b) Correlations of sea level anomalies in sub-regions of the East Siberian  
 239 Sea (blue pentagram in panel (b)) to wind stress (vectors) and sea level anomalies (shading). Panel (c) are the  
 240 same as panel (b) but for sea level anomalies in the Norwegian shelf (blue pentagram in panel (c)). Correlation  
 241 coefficients with 95% significance levels are plotted.

242 Overall, both the model simulations and the several bottom pressure records demonstrate high-frequency  
 243 bottom pressure variability in the Arctic Ocean (Fig. 3). The model simulations reveal that the high-frequency  
 244 variability is barotropic primarily in response to wind-induced Ekman transport and propagations of the baro-  
 245 tropic signals. In the strong current regions, steric effects also contribute to local sea level variability caused  
 246 by baroclinic processes.

#### 247 4.2 Seasonal variability

248 Seasonal sea level variability could be related to the redistribution of water from the deep ocean to the  
 249 marginal seas due to cyclonic/anticyclonic wind stress (Proshutinsky and Johnson, 1997), a seasonal variation of  
 250 the Arctic Ocean volume (Armitage et al., 2016). In addition, the steric effect due to warm Atlantic inflow and

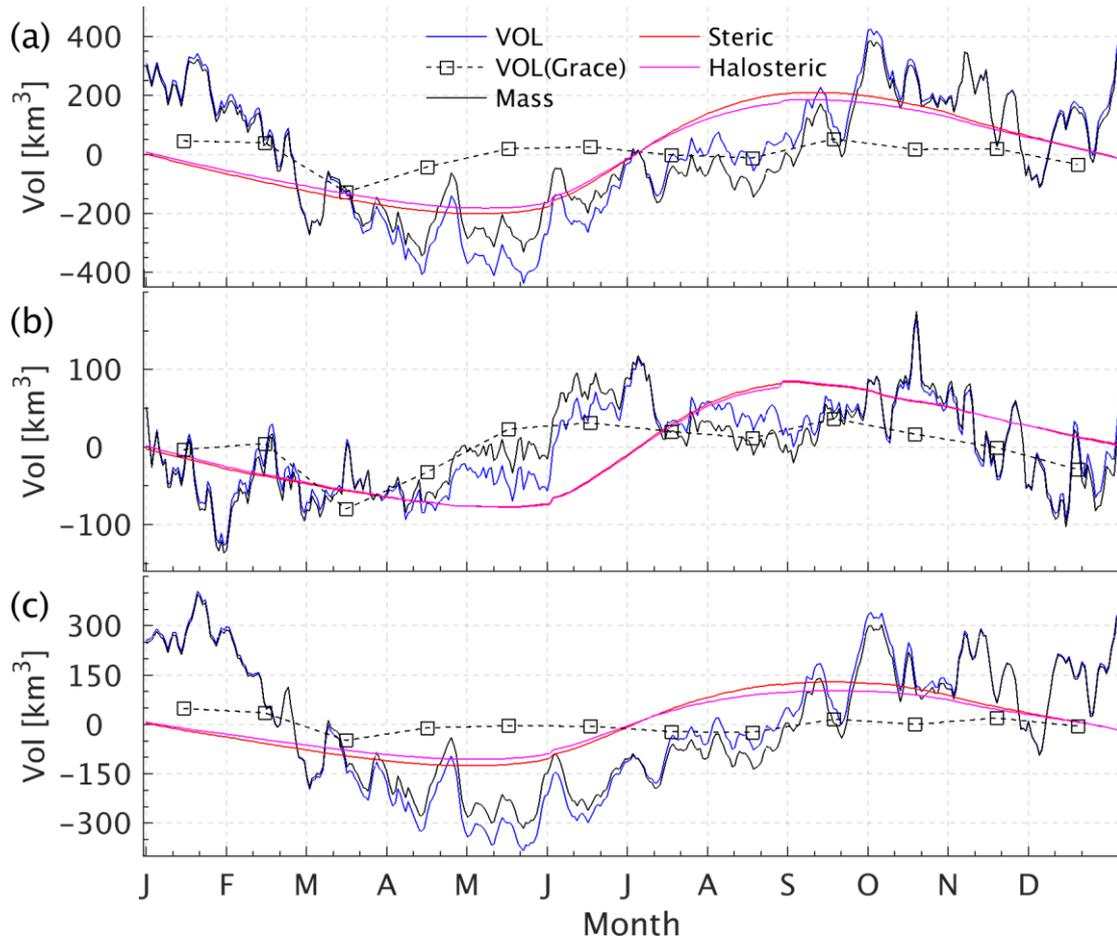
251 sea ice formation/melting contribute to regional sea level variability in the Barents Sea (Volkov and Landerer,  
252 2013). This section focuses on the spatial-varying Arctic sea level variability at seasonal periods and its mecha-  
253 nism.



254

255 Figure 7. RMS variability of (a) mass, (b) halosteric, and (c) thermosteric components at the seasonal periods.

256 Like the high-frequency sea level variability, the mass component still dominates the seasonal sea level  
257 variability in the marginal seas (Fig. 4b and Fig. 7a). Halosteric effects are significant near the river mouth,  
258 seasonal ice edge, and along the coast of Alaska (Fig. 7b), indicating the spreading of freshwater driven by oce-  
259 anic flows. Pathways of freshwater from rivers and marginal seas to the Makarov Basin and the periphery of the  
260 Beaufort Sea can also be inferred from the significant halosteric effect. The thermosteric effects dominate the  
261 ice-free region in the GIN seas and the Barents Sea, and it is remarkably weakened as it penetrates the ice-  
262 covered Arctic Ocean (Fig. 7c).



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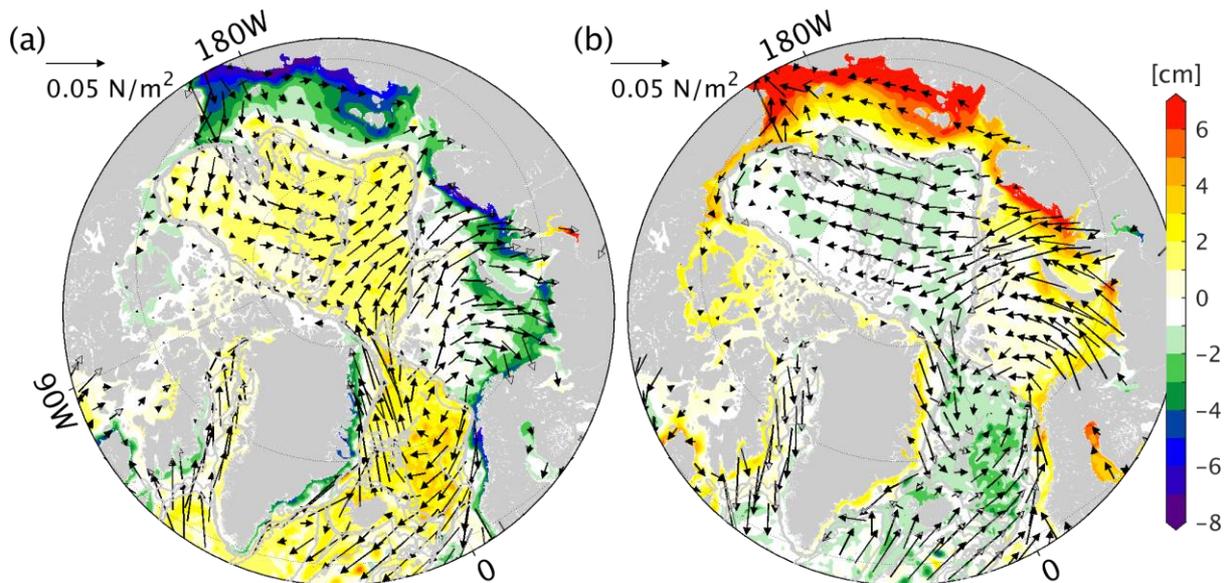
264 Figure 8. (a) Time series of total volume (VOL) anomaly in the Arctic Ocean (see Fig. 1 for the regions) and the  
 265 contributions from mass changes (Mass) and steric effects (Steric). The halosteric component (Halosteric) and  
 266 the GRACE-observed mass component are also shown. Panels (b) and (c) show the corresponding values in (b)  
 267 the deep basin (>500 m) and (c) the shallow water (<500m).

268 Sea level changes reflect total volume changes. The Arctic volume anomalies, dominated by mass compo-  
 269 nent, shows a clear seasonal cycle overlaid with sub-seasonal variability (Fig. 8a). Since the surface freshwater  
 270 flux is treated as a virtual salt flux, river runoff and evaporation minus precipitation do not change the total  
 271 volume directly. The seasonal volume variability, especially the mass component, is driven by volume exchang-  
 272 es with the Pacific and Atlantic Oceans. The steric component (red lines in Fig. 8), especially the halosteric  
 273 component (magenta lines in Fig. 8), causes the volume to decrease in the winter season and increase in the  
 274 summer season due to the sea ice formation/melting.

275 The model simulates more substantial seasonal mass variability than the GRACE measurement. Still, it  
 276 fails to reproduce the secondary peak from May to July (Fig. 8a), which may relate to river discharge in the  
 277 marginal seas. Splitting the total Arctic volume changes into contributions from the deep basin and coastal seas,  
 278 we note that the secondary peak is related to volume changes in the deep basin from May to July (Fig. 8b) in  
 279 both the model simulation and the GRACE observations. At the same time, volume anomalies are negative in  
 280 the marginal seas. This antiphase of the volume anomalies in the deep basin and marginal seas seems to be driv-  
 281 en by the cyclonic/anticyclonic wind pattern in the summer/winter season (Proshutinsky and Johnson, 1997).

282 Mean sea level anomalies from June to August (Fig. 9a) and from December to February (Fig. 9b) further reveal  
 283 the antiphase of the sea level changes between the deep basin and the shallow waters. The mean pattern of wind  
 284 stress anomalies (vectors in Fig. 9) indicates that wind-driven Ekman transport drives the water toward/away  
 285 from the marginal seas, resulting in the antiphase of seasonal sea level variability in the deep basin and shallow  
 286 waters.

287 The model simulation demonstrates the critical importance of exchanges with the Pacific and Atlantic  
 288 Oceans for the Arctic volume changes at seasonal periods. The wind stress will further redistribute water in the  
 289 Arctic Ocean, resulting in the antiphase pattern of sea level changes in the shallow waters and deep basins. Us-  
 290 ing a one-dimensional model, Peralta-Ferriz and Morison (2010) demonstrated that river runoff and evaporation  
 291 minus precipitation (EmP) drive the basin-scale seasonal mass variation of the Arctic Ocean. This process is not  
 292 included in our model simulations due to the virtual salt flux parameterization. But it should be noted that either  
 293 input from river runoff and EmP (Peralta-Ferriz and Morison, 2010) or exchanges with the Pacific and Atlantic  
 294 Oceans is large enough to drive the Arctic volume changes. Moreover, the wind stress will further redistribute  
 295 the water to different regions. It is also expected that volume input from the rivers ( $\sim 700 \text{ km}^3$ ) could signifi-  
 296 cantly alleviate the negative volume anomalies from May to August in the marginal seas.



297

298 Figure 9. SLA (shading) and wind stress anomalies (vectors) to the climatology averaged from (a) June to Au-  
 299 gust and (b) December to February. The daily output of ATLARC4km is used.

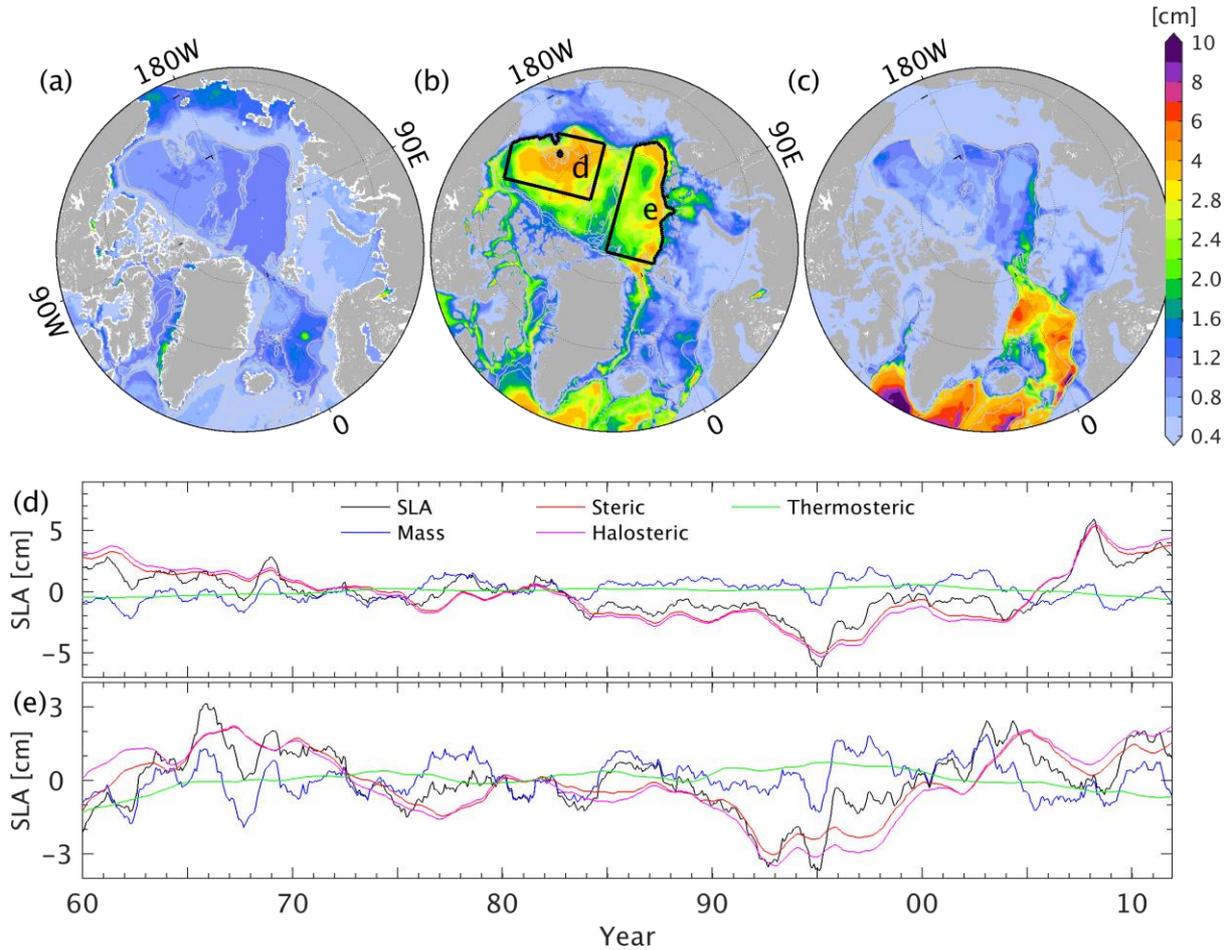
### 300 4.3 Decadal variability

301 The Arctic sea level shows significant decadal variability driven by cyclonic/anticyclonic wind patterns  
 302 (Proshutinsky and Johnson, 1997), accompanied by freshwater content changes (Häkkinen and Proshutinsky,  
 303 2004; Köhl and Serra, 2014). Satellite altimetry observations were used to infer Arctic freshwater content in-  
 304 creases (Armitage et al., 2016; Giles et al., 2012; Proshutinsky et al., 2019; Rose et al., 2019) and complement

305 freshwater content estimate using in-situ observations (Haine et al., 2015;Polyakov et al., 2020;Rabe et al.,  
306 2014;Rabe et al., 2011). This section examines the spatial variability of Arctic decadal sea level and addresses  
307 its relation to the mass, halosteric, and thermosteric components.

308 The ATLARC08km simulation revealed that the pronounced decadal sea level variability in the Canadian  
309 and Eurasian Basins (Fig. 4c) is mainly due to the halosteric effect (Fig. 10b), with the mass components ac-  
310 counting for 20-30%. The thermosteric effect dominates in the GIN seas, mainly relating to the convection pro-  
311 cesses (Brakstad et al., 2019;Ronski and Budéus, 2005). Brakstad et al. (2019, see their Fig. A1) demonstrated  
312 that a change from shallow convection to deep convection can lead to temperature changes of more than  $-0.2\text{ }^{\circ}\text{C}$   
313 over the upper 600 m and salinity changes of 0.02 PSU over the upper 200 m, resulting in a significant thermo-  
314 steric effect. In the north Atlantic Ocean, the thermosteric effect dominates. At the same time, the halosteric  
315 effect compensates for the thermosteric effect in this region, rendering more considerable thermosteric height  
316 variability than decadal total sea level variability.

317 Timeseries of sea level anomalies and their different components confirm that sea level variability is most-  
318 ly halosteric in the Canadian (Fig. 10d, Armitage et al., 2016;Giles et al., 2012;Morison et al., 2012) and Eura-  
319 sian basins (Fig. 10e) and that the thermosteric component contributes with a linear trend (not shown here). In  
320 addition, the mass components contribute to the interannual sea level variability (blue lines in Fig. 10d and e) in  
321 both the basins. We note that the mass changes are highly correlated in the Canadian and Eurasian basins  
322 ( $r>0.98$  with 95% significance level). They are positively correlated to the mass changes in the deep basin of the  
323 GIN seas and the Arctic Ocean and are negatively correlated to mass changes in the Arctic marginal shelves,  
324 especially in the East Siberian Sea, representing a barotropic response of sea level to changes of the intensity  
325 and locations of the Icelandic low and the East Siberian high (e.g., Proshutinsky and Johnson, 1997). The halo-  
326 steric component shows clearly decadal variability and is in phase with that in the Canadian Basin. The thermo-  
327 steric component slightly compensates for the halosteric component.



328

329 Figure 10. RMS variability at the decadal period of (a) bottom pressure anomaly, (b) the halosteric component,  
 330 and (c) the thermosteric component. Panels (d) and (e) show the time series of sea level anomaly and mass,  
 331 steric, and thermo/halosteric components in the Canadian and the Eurasian basins (see the regions in panel (b)),  
 332 respectively. **Linear trends in all the time series are removed in panels (d) and (e).**

### 333 5 Capability of the observing system to monitor freshwater content variability

334 Observing Arctic freshwater content changes remains challenging (Proshutinsky et al., 2019). The results  
 335 above and previous studies (Giles et al., 2012; Morison et al., 2012; Proshutinsky et al., 2019) have indicated that  
 336 satellite altimetry could infer freshwater content changes. International efforts try to enhance the profiles ob-  
 337 serving system, including ice-tethered profilers (ITPs, Toole et al. (2016), doi:10.7289/v5mw2f7x), shipboard  
 338 observations, and moorings. Here, we test their capability to monitor the freshwater changes in an idealized  
 339 setting in which 1) we do not consider influences of observational errors and 2) **we assume the profiles sample**  
 340 **the top 800 m and the moorings sample from 65-800 m.** Freshwater inventory is defined, as in Rabe et al. (2011)  
 341 and Schauer and Losch (2019), as the freshwater fractions relative to a conventional reference salinity  $S_0 = 34.8$   
 342 PSU integrated over depth, and freshwater content is the total freshwater inventory over a region:

$$343 \quad FWC = \int FWI dA = \int \int_H^0 \frac{S_0 - S}{S_0} dz dA \quad (5)$$

344 with  $H$  being the depth of the 34.8 isohaline. The reference salinity indicates the mean salinity within the Arctic  
 345 Ocean and can differ slightly in previous studies, which mainly impacts the mean state of freshwater content.

## 346 5.1 Satellite altimetry and GRACE measurements

347 Giles et al. (2012) used altimetric sea level observations, GRACE-based bottom pressure, and a static 1.5-  
348 layer model to infer freshwater changes in the Canadian Basin. They assumed that freshwater changes lead to  
349 sea level and isopycnal changes simultaneously, changing the water column's layer thickness and total mass. In  
350 this case, freshwater change in the water column is estimated as follows:

$$351 \Delta FW = \frac{S_2 - S_1}{S_2} \cdot \Delta h = \frac{S_2 - S_1}{S_2} \cdot \left( \eta' \cdot \left( 1 + \frac{\rho_1}{\rho_2 - \rho_1} \right) - \frac{\Delta m}{\rho_2 - \rho_1} \right) \quad (6),$$

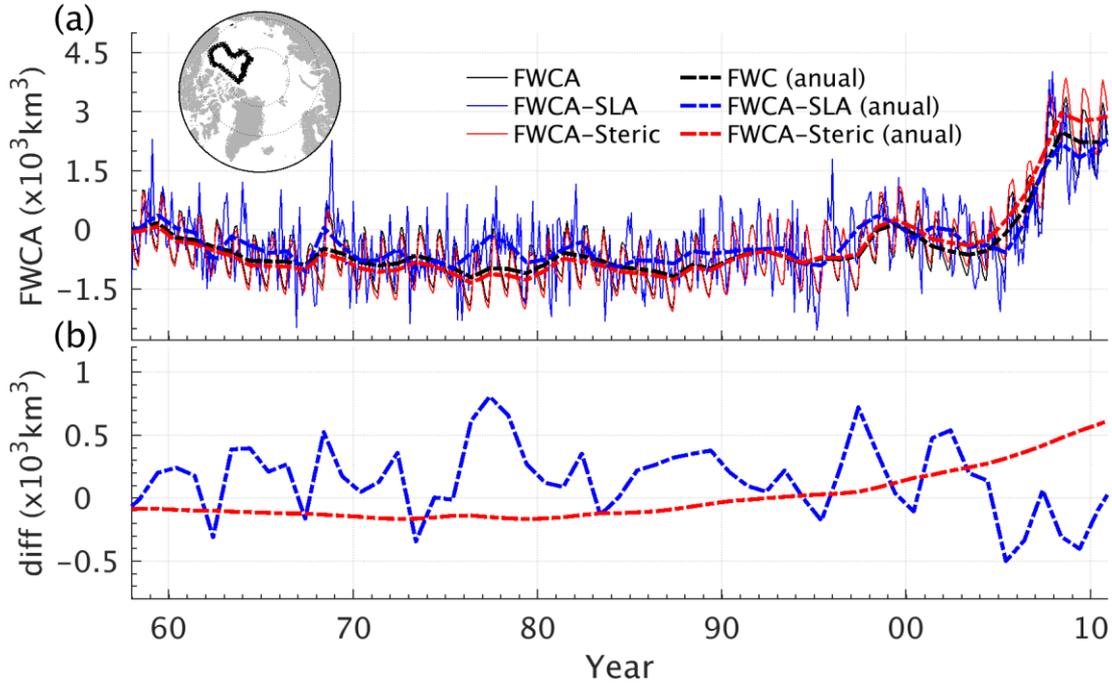
352 where  $\rho_1 = 1025.0 \text{ kg m}^{-3}$  and  $\rho_2 = 1028.0 \text{ kg m}^{-3}$  are the mean density in the top and bottom layers.  $S_1 = 33.0 \text{ PSU}$   
353 is the mean salinity in the top layer, and  $S_2 = 34.8 \text{ PSU}$  is a reference salinity.  $\eta'$  and  $\Delta m$  are the sea level anomaly  
354 and bottom pressure anomalies observations. Morison et al. (2012) suggest that freshwater changes depend on  
355 steric height changes linearly and could be approximated by:

$$356 \Delta FW = \alpha \cdot \eta'_s \quad (7),$$

357 where  $\alpha$  is an empirical constant estimated from in-situ profile observations and is set to 35.6 following  
358 Morison et al. (2012). **The choice of  $\alpha$  just contributes a static offset to freshwater content estimation in Eq. (7).**

359 **In the Canadian Basin, freshwater content changes and the two estimates show similar decadal variabilities,**  
360 **but differences remain in the seasonal and long-term trends (Figs. 11a and b).** Since the halosteric effect domi-  
361 nates the steric effect, estimation using Eq. (7) matches the seasonal freshwater cycle well (red and black lines),  
362 considering the amplitude and phase. However, it overestimates the long-term trend (Fig. 11b) since Eq. (7)  
363 attributes the thermosteric effect to freshwater changes. Eq. (6) infers a much more substantial seasonal variabil-  
364 ity of freshwater content, and the phase does not always match the real freshwater content changes (blue and  
365 black lines).

366  
367



368

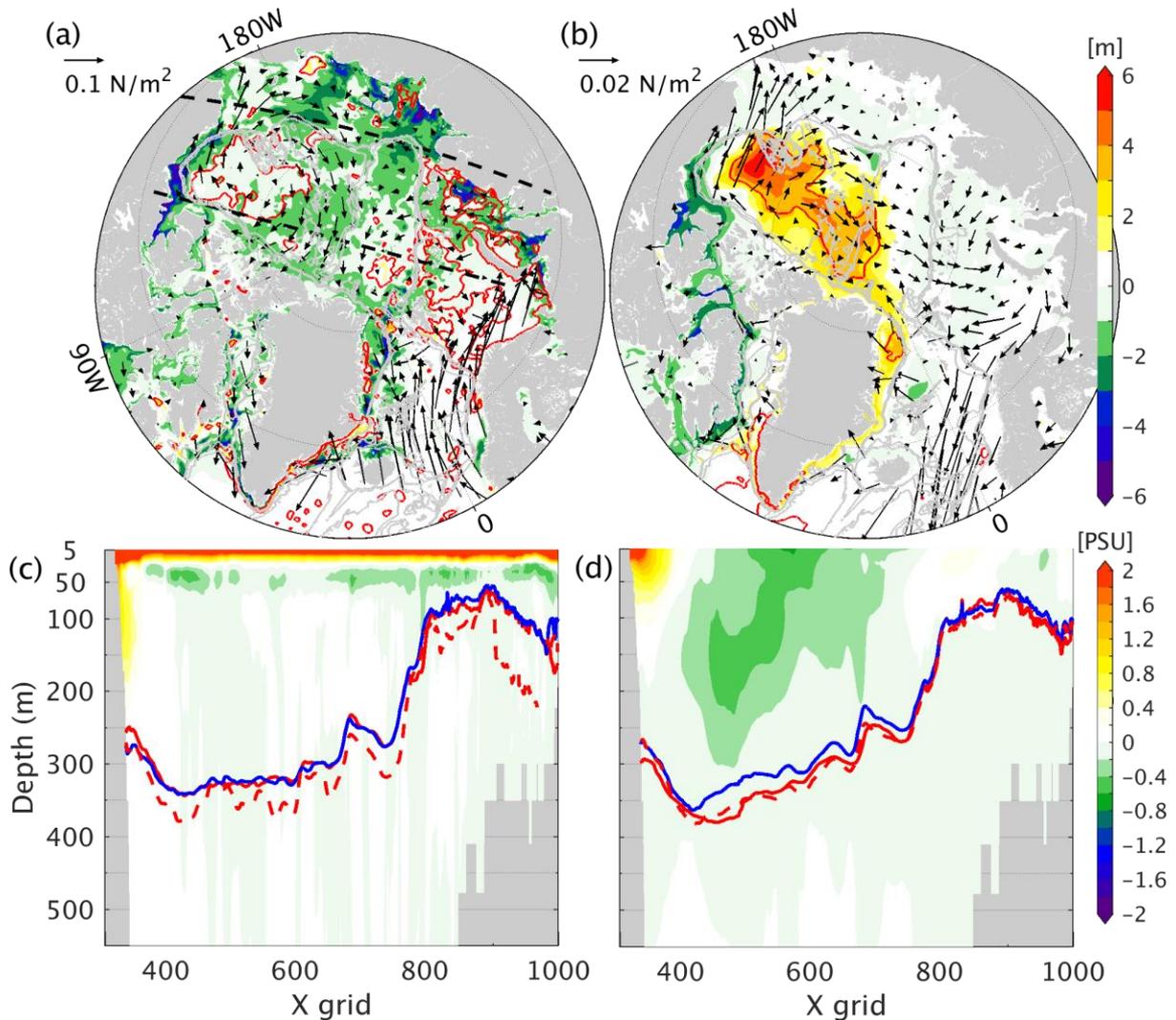
369 **Figure 11.** (a) Freshwater content anomalies ( $10^3 \text{ km}^3$ ) and approximated based on Eq. (6) in blue and Eq. (7) in  
 370 red using the monthly output of ATLARC08km. The thick dashed lines are the annual mean values. (b) The  
 371 differences of the approximated annual mean freshwater content anomalies based on Eq. (6) in blue and Eq. (7)  
 372 in red to the annual mean freshwater content anomalies.

373 Eq. (6) assumes the **upper layer** adjusts simultaneously with sea level anomaly, which may not apply in the  
 374 presence of baroclinic effects. To illustrate the limitation of Eq. (6), we take the differences between Feb. 2003  
 375 and Sep. 2002 (in which Eq. (6) fails to reproduce the phase and the amplitude of freshwater content changes)  
 376 and between 2008-2010 and 1994-1996 (when Eq. (6) reproduces the freshwater changes well).

377 From Sep. 2002 to Feb. 2003 (Fig. 12a), anticyclonic wind stress anomalies occur in the Beaufort Sea, re-  
 378 sulting in positive SLA through Ekman transport. However, freshwater content is reduced during this period.  
 379 The salinity difference averaged over the central Arctic Ocean reveals that salinity increases in the top 30 m  
 380 **were** caused by ice formation. At the same time, the isopycnal ( $27.9 \text{ kg m}^{-3}$ ) **did** not deepen (Fig. 12c) as pre-  
 381 dicted by Eq. (6). The assumption that freshwater content changes are captured by freshwater column thickness  
 382 changes  $\eta \cdot \left(1 + \frac{\rho_1}{\rho_2 - \rho_1}\right)$  (red dashed lines in Fig. 12c) fails to infer freshwater content changes in this case.

383 From 1994-1996 to 2008-2010, anticyclonic wind stress anomalies appeared in the Canadian Basin, ac-  
 384 companied by positive SLA and freshwater content anomalies (Fig. 12b). During that period, Ekman pumping  
 385 deepens the isopycnals (blue and red lines in Fig. 12), accumulating more freshwater and reducing the local  
 386 salinity over the top 300 m (Fig. 12d). In this scenario, the water column thickness change dominates the fresh-  
 387 water content variability, which is approximated by  $\eta \cdot \left(1 + \frac{\rho_1}{\rho_2 - \rho_1}\right)$  (red dashed lines in Fig. 12d). Therefore, Eq.  
 388 (6) captures the interannual freshwater content changes using satellite altimetric observations. Caution needs to  
 389 be taken when inferring Arctic Ocean freshwater content changes using satellite altimetry observations and

390 GRACE measurements. In addition, Figs. 12b and 12d indicate that Eq. (6) can be only used in the Canadian  
 391 Basin where wind drives the sea level changes and the deepening/shoaling of the isopycnals.

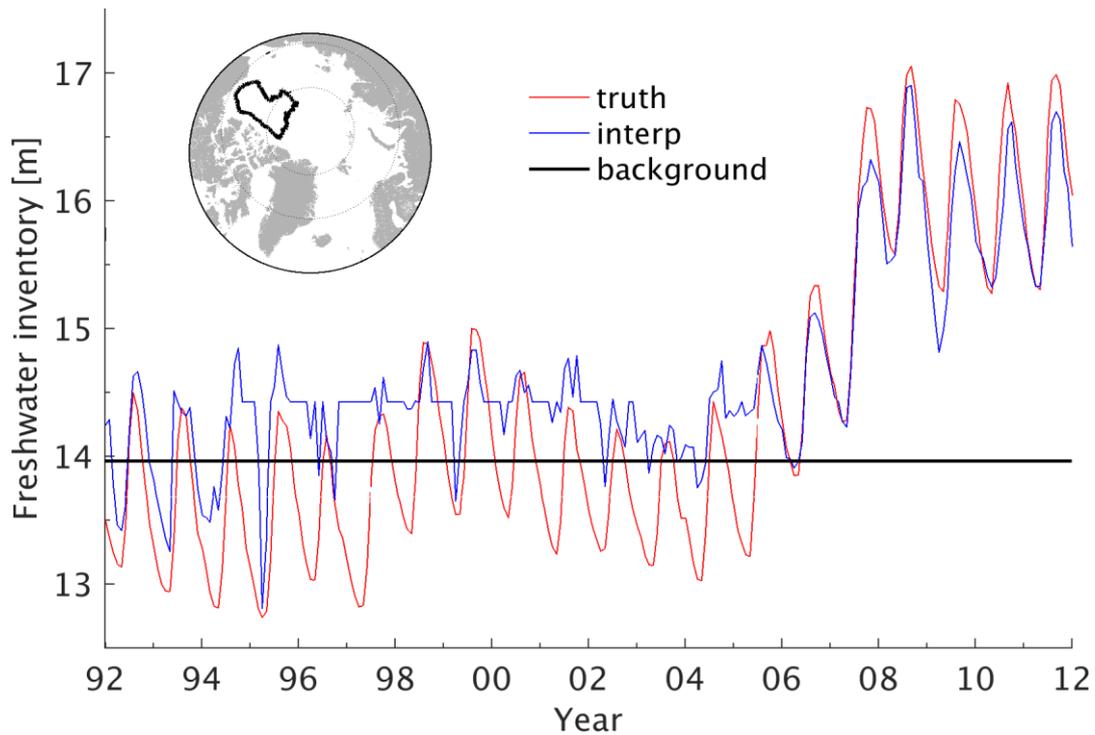


392 Figure 12. The differences of freshwater inventory in meters (shading), sea level anomaly (0.15 m contour,  
 393 black lines), and wind stress(vectors) between (a) Feb. 2003 and Sep. 2002, and (b) 2008-2010 and 1994-1996.  
 394 Panels (c) and (d) are the corresponding salinity differences (shading) average over the central Arctic Ocean  
 395 (black dashed lines in panel (a)). The blue lines denote the 27.9 kg m<sup>-3</sup> isopycnal in Sep. 2002 and 1994-1996,  
 396 respectively. The red lines and red dashed lines are the 27.9 kg m<sup>-3</sup> isopycnal and the diagnosed 27.9 kg m<sup>-3</sup>  
 397 isopycnal with SLA and Eq. (5) in Feb. 2003 and 2008-2010, respectively.

## 398 5.2 In-situ profilers

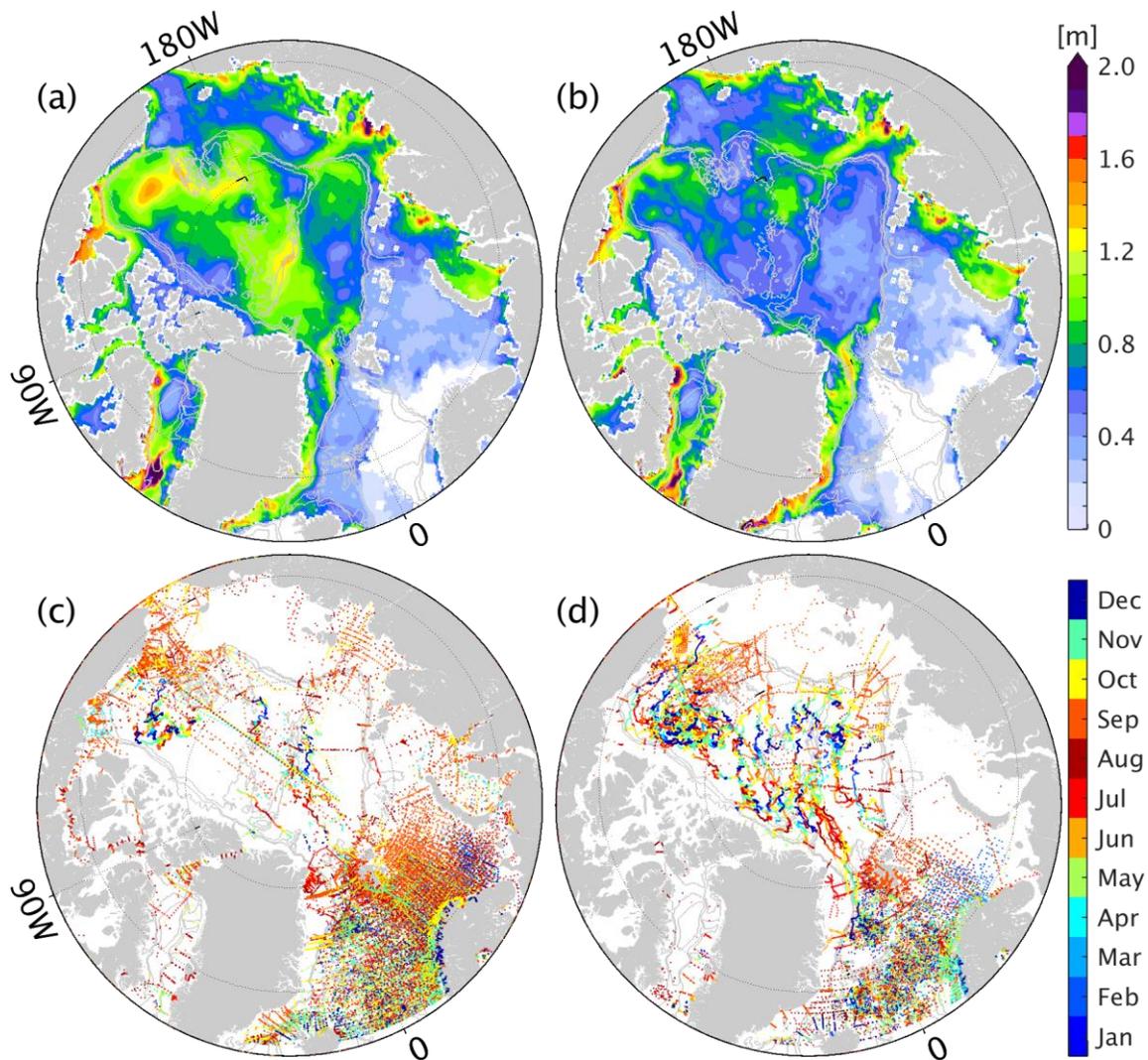
399 In-situ profilers measure salinity directly, but they are limited by sea ice and distributed unevenly in time  
 400 and space. Over the past decades, the endeavor of polar expeditions and the evolving measurement techniques  
 401 (e.g., ITP) have generated a large number of hydrographic data in the central Arctic and subarctic seas (e.g.,  
 402 Behrendt et al., 2018). Using historical hydrographic observations and objective mapping techniques, previous  
 403 studies (e.g., Haine et al., 2015; Polyakov et al., 2008; Rabe et al., 2014; Rabe et al., 2011) have explored Arctic  
 404 freshwater content changes and the mechanisms on multi-year periods. However, the interpolated products suffer  
 405 from high uncertainties at timescales shorter than multi-year periods (e.g., Fig. 4 in the supplement of Rabe

406 et al., 2014), indicating observational gaps on resolving the seasonal to interannual freshwater content changes.  
 407 Besides, spatial observational gaps are observed (e.g., Fig. 7 in Rabe et al., 2011) but not explored yet. This  
 408 section examines how existing hydrographic observations could help reveal Arctic freshwater content changes  
 409 and identify observational gaps in time and space. Based on the spatiotemporal distribution of profiles compiled  
 410 by Behrendt et al. (2018) and an ensemble optimal interpolation (EnOI) scheme (Evensen, 2003;Lyu et al.,  
 411 2014), we test to what extent the generated synthetic profiles could help to reconstruct the "true" state (here the  
 412 ATLARC08km simulation) during the periods 1992 to 2012. Details of the EnOI scheme are given in Appendix  
 413 A.



414  
 415 Figure 13. Mean freshwater **inventory** (in meter) in the Canadian Basin (enclosed by the black line in the top  
 416 subplot) from the background state, the "truth", and the optimal interpolation reconstructed state (see legend).

417 As shown in Fig. 13, the sparse in-situ profiles help bring the freshwater inventory in the background state  
 418 close to the "truth" state. However, it was not until 2007 that the reconstructed state reproduces the seasonal to  
 419 inter-annual freshwater inventory variability in the Canadian Basin, benefiting from the increasing number of  
 420 research activities and international collaborations. We further examined RMS errors of freshwater inventory  
 421 from 1992-2006 (Fig. 14a), 2007-2012 (Fig. 14b), and the corresponding profile locations (Fig. 14c and d). The  
 422 lack of in-situ profiles in the Arctic shelves (Figs. 14c and d) and in the deep basin from 1992-2006 (Fig. 14c)  
 423 results in pronounced errors. The ITP profiles (trajectories in Fig. 14d) enhanced the capability to observe the  
 424 Arctic freshwater changes in the deep basin and the winter season, significantly reducing freshwater inventory  
 425 uncertainties (Fig. 14b). Additionally, high errors remain in regions with high variability (e.g., EGC/WGC), in  
 426 the Laptev Sea and the Alaskan coast, extending from the coasts to the deep basin, underlining the observing  
 427 requirements.



428

429 Figure 14. Root mean square errors of **freshwater inventory** (in meter) between the reconstructed state and the  
 430 "truth" from (a) 1992-2006 and (b) 2007-2012. **Panels (c) and (d) are the corresponding profile locations in**  
 431 **different months (color bar).**

432

433

The above results highlight that the increase of hydrographic observations has enhanced our ability to re-  
 434 construct the changes in Arctic freshwater content since 2007. A lack of hydrographic observations in the  
 435 coastal areas results in significant errors in the marginal seas, which require extensive international collabora-  
 436 tions.

## 437 6 Summary and conclusions

438 Sea level variability reflects changes in ocean dynamics, atmospheric forcing, and terrestrial runoff pro-  
 439 cesses (Stammer et al., 2013). In particular, sea level observations have been applied to infer freshwater content  
 440 changes (Armitage et al., 2016; Giles et al., 2012; Proshutinsky et al., 2019) in the central Arctic Ocean. To com-  
 441 plement our understanding of the Arctic sea level variability and its mechanisms, we use two high-resolution  
 442 ATLARC model simulations to investigate the Arctic sea level variability at different timescales and the relation  
 443 with bottom pressure and thermo/halosteric effects, identifying critical observational gaps that need to be filled.

444 Both the model simulations and mooring observations reveal very high-frequency bottom pressure varia-  
445 tions. The model simulations confirm that the bottom pressure anomaly is equivalent to sea level anomaly in  
446 most areas of the Arctic Ocean at periods <30 days, reflecting the barotropic nature of this high-frequency vari-  
447 ability. Correlation analyses show that the high-frequency sea level variability is caused by wind-driven Ekman  
448 transport and propagations of these barotropic signals.

449 The seasonal sea level variability is dominated by volume exchanges with the Pacific and Atlantic Oceans  
450 and the redistribution of the water by wind stress. Halosteric effects due to river runoff and ice melt-  
451 ing/formation are also pronounced in the marginal seas and seasonal sea ice extent regions. Peralta-Ferriz and  
452 Morison (2010) demonstrated that river runoff and EmP drive the seasonal cycle of the Arctic bottom pressure.  
453 **Although the virtual salt flux parameterization could not mimic the influences of volume input from rivers and**  
454 **surface fluxes, the model simulations still simulate much stronger seasonal mass anomalies than the observa-**  
455 **tions from GRACE.** Either volume exchanges with the Pacific and Atlantic Oceans or volume input from river  
456 runoff and EmP are large enough to cause the Arctic Ocean's seasonal volume variability. They should work  
457 together, resulting in the Arctic seasonal volume variability. **We speculate that using river runoff and EmP as**  
458 **volume flux, rather than the virtual salt flux, could likely improve the volume and sea level variability in the**  
459 **marginal seas from April to July since the volume inputs from river runoff could alleviate the negative volume**  
460 **anomalies in the marginal seas caused by wind.**

461 At decadal timescales, the model simulations further confirm that the pronounced sea level variability in the  
462 central Arctic Ocean, especially in the Canadian Basin, is mainly a halosteric effect. Using the satellite altimet-  
463 ric observations and GRACE observations, the method of Giles et al. (2012) could infer the freshwater content  
464 changes in the Canadian Basin reasonably at timescales longer than one year since **the upper layer (indicated by**  
465 **the 27.9 kg m<sup>-3</sup> isopycnal in this study)** requires time to adjust to sea level changes. Inferring freshwater content  
466 changes using a linear relation of freshwater content and steric height (Morison et al., 2012) reveals both the  
467 interannual and the seasonal variability of freshwater content. However, **caution needs** to be taken since the  
468 method attributes the thermosteric effects to halosteric effects, resulting in an additional linear trend. In addition,  
469 uncertainties in the satellite altimetric and GRACE measurements make the estimation more complicated and  
470 introduce significant uncertainties in the steric effects and freshwater content estimation (Ludwigsen and  
471 Andersen, 2021).

472 The increasing number of international collaborations and new measurement techniques have generated a  
473 large number of profiles. Previous studies have applied **different objective mapping methods (Haine et al.,**  
474 **2015;Polyakov et al., 2008;Rabe et al., 2014;Rabe et al., 2011)** to reconstruct the Arctic freshwater content  
475 changes and budget. **However, the interpolated products still show high errors for the annual mean estimate of**  
476 **freshwater content, indicating potential observational gaps in resolving the seasonal freshwater content cycle.**  
477 **We further examined the observational gaps in time and space using monthly output from ATLARC08km.**  
478 Through reconstructing the salinity with synthetic observations, we note that the in-situ profile system seems to  
479 capture the seasonal freshwater variability since the year 2007, encouraging further Arctic data synthesis studies  
480 (Behrendt et al., 2018;Cheng and Zhu, 2016;Steele et al., 2001) with more complicated interpolation methods.  
481 In addition, international collaborations need to be enhanced to fill in the observational gaps in the marginal seas.

482 Further observing system simulation experiments (e.g., Lyu et al., 2021;Nguyen et al., 2020) should be per-  
483 formed **in a coordinated fashion** to develop an autonomous Arctic observing system (Lee et al., 2019;Sandu et  
484 al., 2012) **to meet the societal and scientific needs.**

## 485 **7 Data availability**

486 The data used to create the plots in the paper are available at Pangaea ([https://issues.pangaea.de/browse/PDI-](https://issues.pangaea.de/browse/PDI-22940)  
487 [22940](https://issues.pangaea.de/browse/PDI-22940)). To access the results of the two high-resolution ATLARC model simulations, please contact Dr. Nuno  
488 Serra at <https://www.ifm.uni-hamburg.de/en/institute/staff/serra.html>. **The Beaufort Gyre Exploration Program**  
489 **data were collected and made available by the Woods Hole Oceanographic Institution**  
490 **(<https://www2.whoi.edu/site/beaufortgyre/>)** in collaboration with researchers from Fisheries and Oceans Canada  
491 **at the Institute of Ocean Sciences and were derived from**  
492 **<https://www2.whoi.edu/site/beaufortgyre/data/mooring-data/>.** The North Pole Environmental Observatory data  
493 **were derived from <http://psc.apl.washington.edu/northpole/Mooring.html>.** The satellite altimetric and GRACE  
494 **measurements were retrieved via [http://www.cpom.ucl.ac.uk/dynamic\\_topography](http://www.cpom.ucl.ac.uk/dynamic_topography) and**  
495 **[https://podaac.jpl.nasa.gov/announcements/2021-06-11-GRACE-and-GRACE-FO-L3-Monthly-Ocean-and-](https://podaac.jpl.nasa.gov/announcements/2021-06-11-GRACE-and-GRACE-FO-L3-Monthly-Ocean-and-Land-Mass-Anomaly-RL06-04-Dataset-Release)**  
496 **[Land-Mass-Anomaly-RL06-04-Dataset-Release](https://podaac.jpl.nasa.gov/announcements/2021-06-11-GRACE-and-GRACE-FO-L3-Monthly-Ocean-and-Land-Mass-Anomaly-RL06-04-Dataset-Release).** We gratefully acknowledge the Ice-Tethered-Profiler Pro-  
497 **gram based at the Woods Hole Oceanographic Institution (<https://www.whoi.edu/itp>)** and the Unified Database  
498 **for Arctic and Subarctic Hydrography (<https://doi.pangaea.de/10.1594/PANGAEA.872931>)** collected and com-  
499 **plied by the Alfred Wegener Institute.**

500

501 *Author contribution.* G. Lyu performed the analysis and wrote the paper. N. Serra performed the model simula-  
502 tions. D. Stammer proposed this study, and M. Zhou provided advice on the analysis.

503 *Competing interests.* The authors declare that they have no conflict of interest.

504

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## 515 Appendix A

### 516 An EnOI Scheme

517 We use an EnOI scheme (Cheng and Zhu, 2016) to reconstruct the salinity in the Arctic Ocean using syn-  
518 thetic observations. At one grid (denoted by subscript  $g$ ), the analysis state  $\varphi_g^a$  is a linear combination of a back-  
519 ground field  $\varphi_g^b$  and surrounding in-situ observations  $d$ :

$$520 \varphi_g^a = \varphi_g^b + K(d - H\varphi_g^b) \cdot e^{-\frac{x^2}{2\sigma^2}} \quad (\text{A1}),$$

521 where  $H$  is a transfer matrix that maps model state from model space to observation points. In this study, the  
522 background state of salinity  $\varphi_g^b$  is taken as the mean salinity at each grid over the period 1992-2012.  $K$  is the  
523 Kalman gain, calculated as:

$$524 K = \varphi'_g \varphi'^T_g H^T (H \varphi'_g \varphi'^T_g H^T + \gamma \gamma^T)^{-1} \quad (\text{A2}).$$

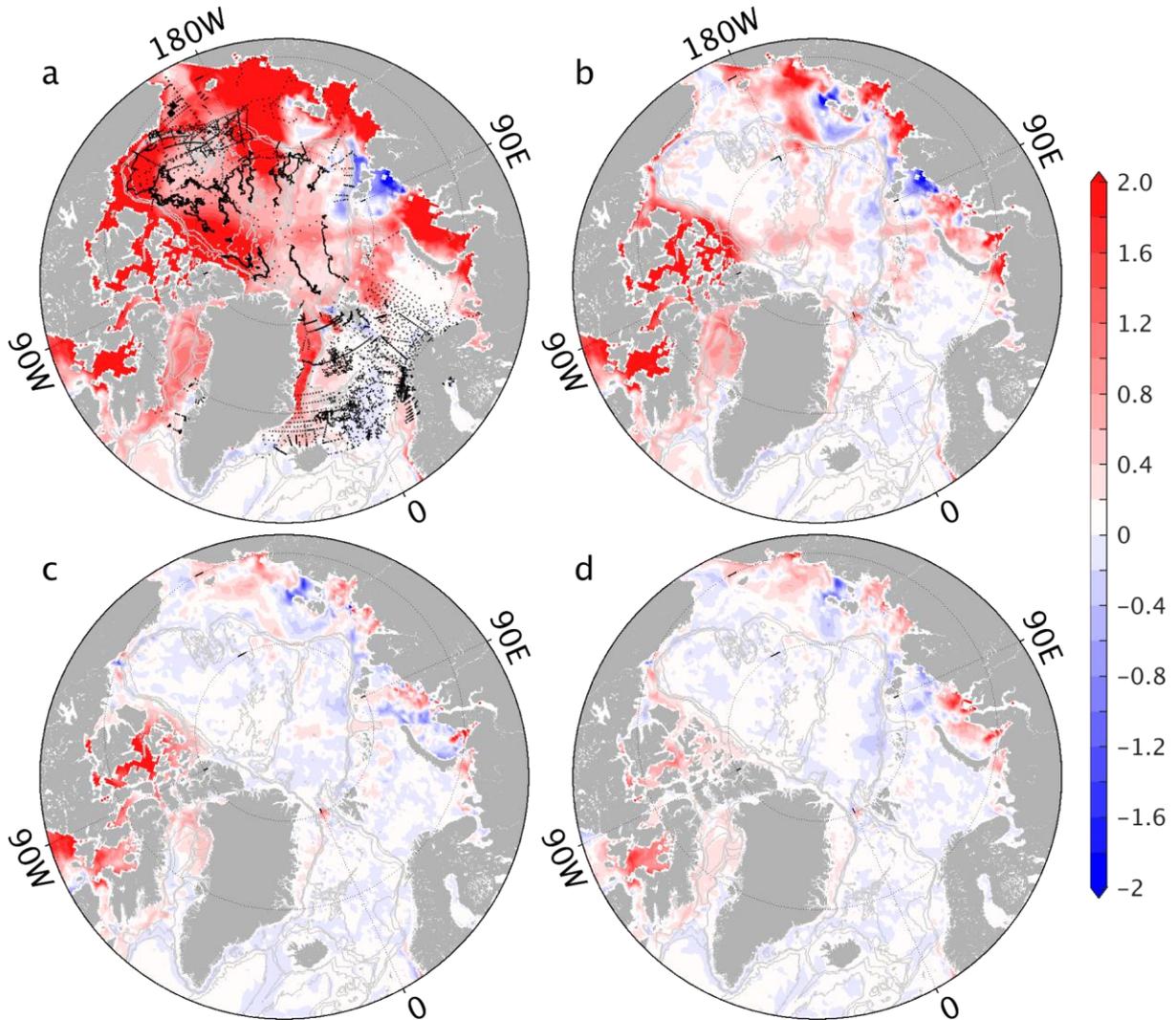
525 The superscript  $T$  denotes matrix transposition. In this formulation, we use  $\varphi'_g$ , the salinity deviation from  
526 the mean salinity, to compute the error covariance of the background state ( $\varphi'_g \varphi'^T_g$ ). We use monthly data from  
527 the year 1992 to 2012 to compute  $\varphi'$ , resulting in a total of 252 ensemble members. For simplicity, we assume  
528 the representation errors  $\gamma$  only depend on depth, ranging from 0.09 PSU at the surface to 0.02 PSU in the deep  
529 ocean, and are not correlated.

530 The use of ensemble members to approximate the background error covariance ( $\varphi'_g \varphi'^T_g$ ) will inevitably  
531 introduce long-distance correlations and propagate the observational information incorrectly over a much longer  
532 distance. Therefore, we introduce a Gaussian function depending on the distance between observational  
533 locations and the model grid ( $x$  in Eq. A1) and a decorrelation radius ( $\sigma$  in Eq. A1) to ensure that only  
534 observations within the decorrelation radius  $\sigma$  of a model grid point could modify the analysis state.

535 Taking the "true" salinity state from August 1992 and observation locations from 2008 (black dots in Fig.  
536 A1a), we test the impacts of the decorrelation radius on the analysis field. The background state is more saline  
537 than the truth (Fig. A1a). With a 300 km decorrelation radius (Fig. A1b), the analysis state reduces the errors  
538 near the observations while significant errors remain in regions far from observations. Increasing the

539 **decorrelation** radius to 1000 km, we see that salinity errors in the marginal seas, north pole areas, and the Baffin  
540 bay are reduced (Fig. A1c). A 2400 km **decorrelation** radius further reduces salinity error in the Canadian Arctic  
541 Archipelago (Fig. A1d). However, only slight improvements are observed in the central Arctic Ocean, and  
542 errors in the Kara Sea are slightly increased. Since we focus on the Arctic freshwater content variability, we use  
543 a 1000 km **decorrelation** radius throughout this study.

544



545

546 Figure A1. Example of sea surface salinity difference between (a) the background and the truth, (b) the analysis  
547 with a **decorrelation** radius of 300 km and the truth, (c) the analysis with a **decorrelation** radius of 1000 km and  
548 the truth, and (d) the analysis with a **decorrelation** radius of 2400 km and the “truth.” Black dots in panel (a)  
549 denote the locations of synthetic observations, sampled using sites of the observations from the year 2008.  
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