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Arctic sea level variability from high-resolution model simulations and implications for the Arctic observing system

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

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1 Introduction

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

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

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

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

Our study systematically explores the Arctic sea level variability as function of timescale and geographic location using daily and monthly outputs of two high-resolution model simulations. Contributions from barotropic changes expressed in bottom pressure variations and baroclinic processes represented by thermo/halosteric changes are quantified at different timescales. We further discuss the existing Arctic Ocean observing system's capability to monitor the Arctic freshwater content variability.

The structure of the remaining paper is as follows: the numerical models and the observations from the bottom pressure sensor, GRACE, and satellite altimetry are described in Section 2, together with different components of sea level changes. We compare the model simulations against observations in Section 3. Section 4 analyzes sea level variability and associated mechanisms at high frequency (<30 days), at the seasonal cycle and at

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decadal timescales. The relations with bottom pressure and thermos/halosteric components are demonstrated, pointing out key regions and parameters we need to observe. Further, we analyze the ability of satellite altimetry, GRACE, and the in situ profiler system to monitor the Arctic freshwater content variability in Section 5. Section 6 provides a summary and conclusions.

2 Model Simulations and observations

2.1 Atlantic-Arctic simulations

This study relies on two ocean high-resolution numerical simulations using the MIT general circulation model (Marshall et al., 1997). A dynamic thermodynamic sea ice model (Hibler, 1979, 1980;Zhang and Rothrock, 2000), implemented by Losch et al. (2010), is employed to simulate sea ice processes. The model domain covers the entire Arctic Ocean north of the Bering Strait and the Atlantic Ocean north of 33°S. In the horizontal, the model uses a curvilinear grid with resolutions of ~8 km (ATLARC08km) and ~4 km (ATLARC04km). In the vertical, ATLARC08km and ATLARC04km have 50 and 100 vertical z-levels, respectively.

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

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

	Horizontal resolution	*Vertical grid*	*Surface forcing*	*periods*	*Output Frequency*
ATLARC08km	~8 km	50 z-levels	NCEP-RA1	1948-2016	monthly
				05.01.2003-01.12.2010	daily
ATLARC04km	~4 km	100 z-levels	ERA-Interim	01.01.2003-23.08.2012	daily

2.2 Satellite and in-situ observations

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

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The monthly altimetric sea level observations from Armitage et al. (2016) and GRACE measurements (Chambers and Bonin, 2012) are used in a comparison with the model simulations. For the very high-frequency variability, bottom pressure records supplied by the Beaufort Gyre Exploration Project (BGEP, M_a, M_b, M_c, and M_d in Fig. 1) and the North Pole Environmental Observatory (NPEO, M_{npeo} in Fig. 1, Morison et al., 2012) are used. Tidal signals are removed using the T_TIDE Matlab program (Pawlowicz et al., 2002), since the model did not include tidal forcing.

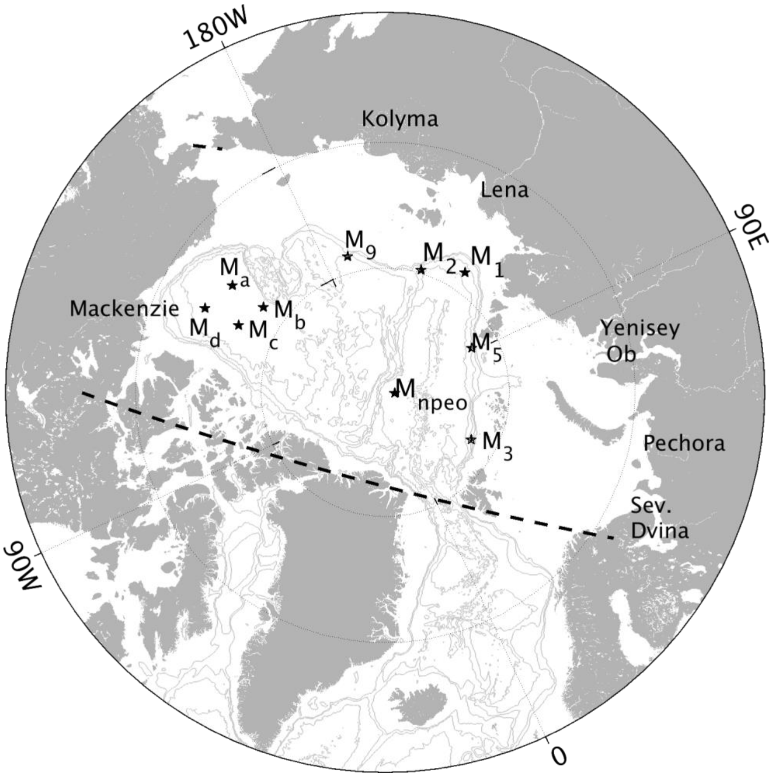

Figure 1. A map of the pan-Arctic Ocean presenting the locations of moorings deployed by the Nansen and Amundsen Basin Observational System (NABOS, black pentagrams labeled with M_1, M_2, M_3, M_5, and M_9), by the Beaufort Gyre Exploration Project (BGEP, black pentagrams marked with M_a, M_b, M_c, M_d), and by the North Pole Environmental Observatory (NPEO, black pentagram labeled with M_{npeo}). The black dashed lines enclose the Arctic regions used in Section 4.2. Bathymetry contours of 500, 1000, 2000, 3000, and 4000 m are drawn with grey lines. Main rivers are labeled with their names near the river mouths.

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

Following Ponte (1999) and Calafat et al. (2013), sea level anomaly η', can be separated into a steric component η_s' due to density change, an inverse barometer effect η_{IB}', and a mass (measured by bottom pressure observations) component η_m':

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

where g=9.8 m s^{-2} is the gravitational acceleration. The first term on the right-hand side represents the steric effect η_s', with ρ_0 being a reference density (1025.0 kg/m^3 in this study) and ρ' being the density change. The

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second term is the inverse barometer effect η'_{IB}: $\overline{P'_a}$ and P'_a represent air pressure anomalies average over the global ocean and at the observing location, respectively. The last term defines the mass component η'_m.

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

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

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

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

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

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

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

3 Testing simulations against observations

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

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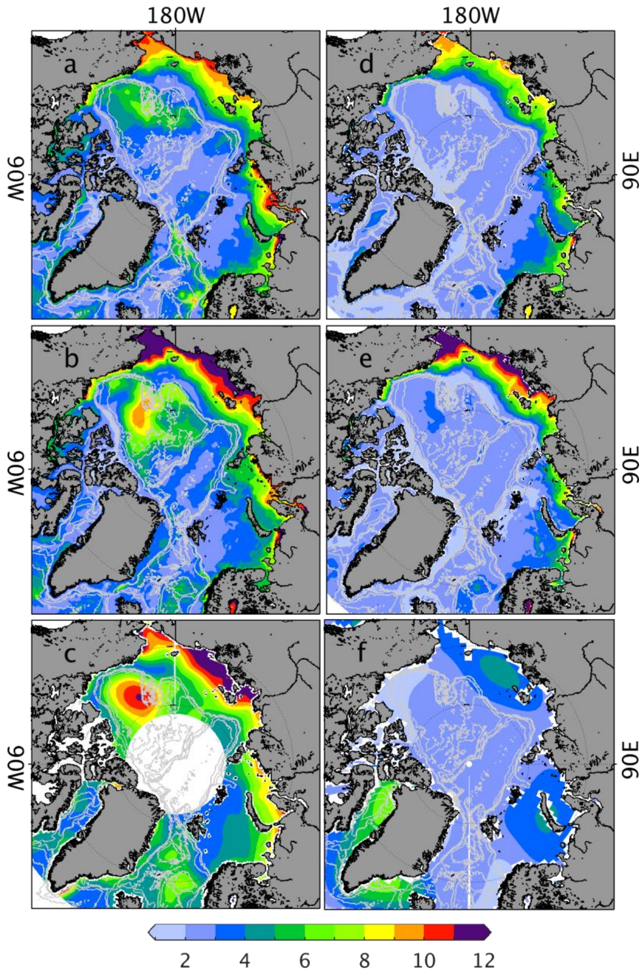

Figure 2. RMS variability of (a-c) sea level and (d-f) bottom pressure in (a, d) ATLARC08km, (b, e) ATLARC04km, (c) satellite altimetry, and (f) GRACE. We computed the RMS variability using monthly data from January 2003 to December 2011. Bathymetry contours of 500, 1000, 2000, and 3000 m are drawn with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

Both the model simulations (Fig. 2a, b) and satellite altimetry (Fig. 2c) reveal pronounced sea level variability in the Canadian Basin and along the coast, which could be attributed to the redistribution of water due to the shifting of basin-scale cyclonic/anticyclonic wind (Proshutinsky and Johnson, 1997) and to the discharge and transport of river runoff along the coast (Proshutinsky et al., 2007). ATLARC04km simulates more significant sea level variability than ATLARC08km, especially in the East Siberian Sea and the Canadian Basin, and matches better with the observed sea level variability. Bottom pressure also shows significant variability in the Arctic marginal seas (Fig. 2d-f), especially in the East Siberian Sea. However, due to the smoothing process applied on GRACE measurements (a 500 km Gaussian filter), both the model simulations simulate much more substantial RMS variability of bottom pressure.

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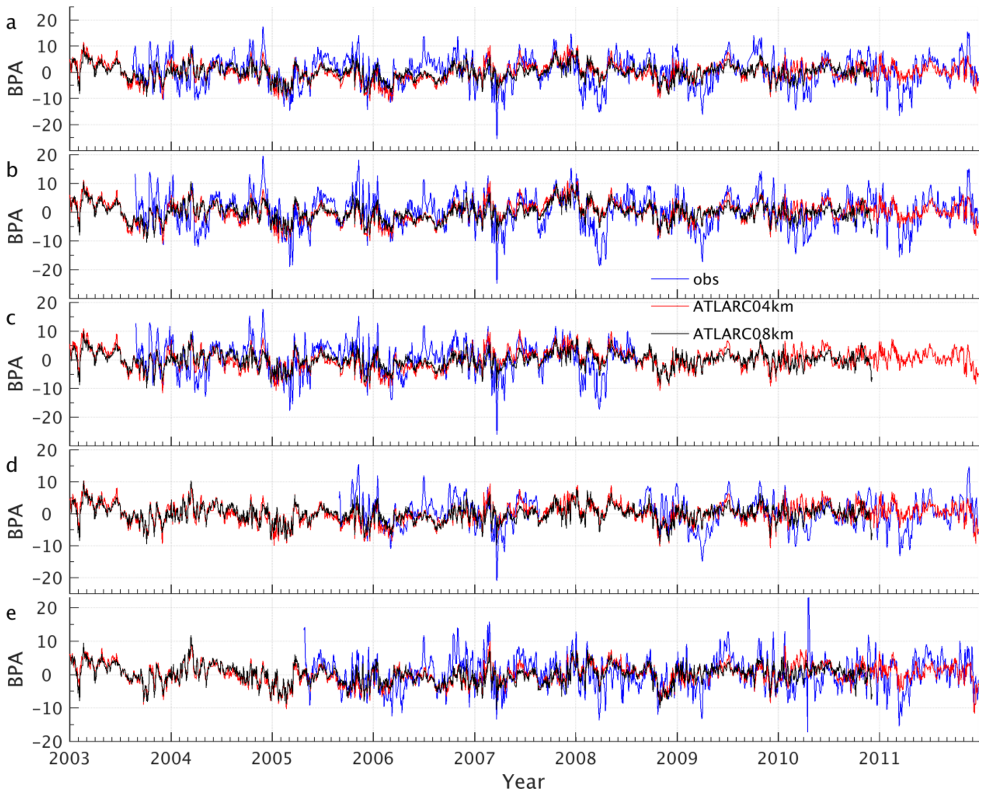

Figure 3. Time series of bottom pressure anomalies in ATLARC08km, ATLARC04km, and in-situ observations. Observations are derived from (a) mooring A, (b) mooring B, (c) mooring C, and (d) mooring D of BGEP. Panel (e) is from the NEPO moorings. Mooring locations are marked in Fig. 1.

Besides monthly to decadal variability of bottom pressure, both the model simulations and the in-situ observations also demonstrate significant high-frequency bottom pressure anomalies (Fig. 3). Both model simulations correlate well with the observations (~0.45-0.55) in the five shown locations, but ATLARC04km and ATLARC08km underestimate the RMS variability by ~30-50%, with ATLARC04km showing relative more significant RMS variability.

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

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

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

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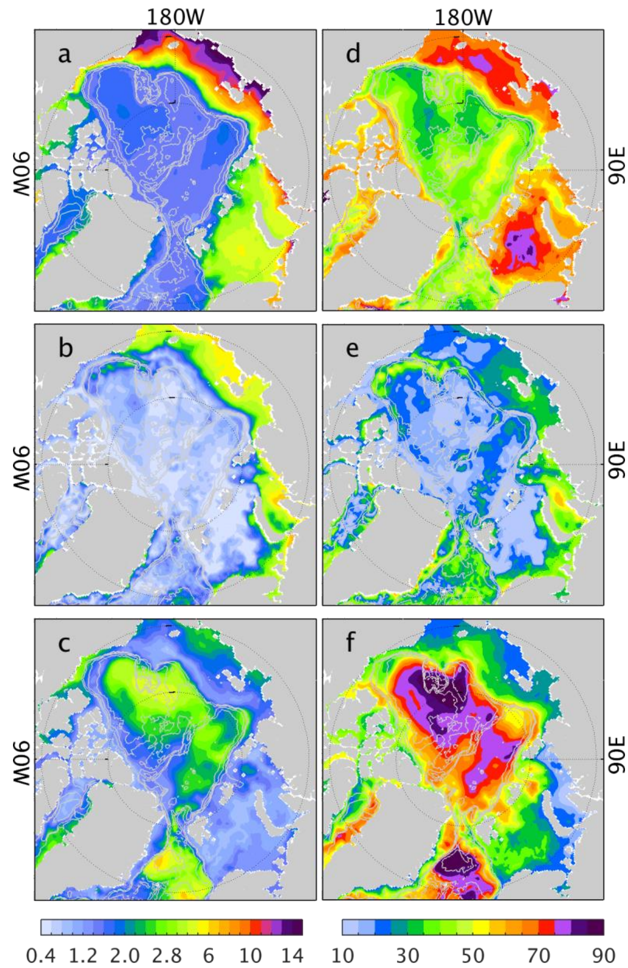

Figure 4. RMS variability (cm) of sea level (a) in the high-frequency band (<30 days), (b) at the seasonal cycle, and (c) at decadal periods (>4 years). Panels (d)-(f) are the corresponding ratios (%) to the total sea level variance that panels (a)-(c) explained. The grey lines denote bathymetry contours of 500, 1000, 2000, and 3000 m.

At period <30 days, RMS variability of sea level up to 14 cm appears in the marginal seas and along the coasts (Fig. 4a), accounting for 60%~80% of the local sea level variance (Fig. 4d). The seasonal sea level variability is pronounced in the marginal seas and southern edge of the Beaufort Sea, and it explains 20%-40% of the total sea level variance. In the deep regions of the pan-Arctic Ocean, the decadal variability dominates the sea level variability, and it explains more than 70%~90% of the sea level variability. Overall, in the marginal seas, sea level variability is dominated by sub-monthly and seasonal signals. In contrast, decadal sea level variability dominates in the deep regions of the pan-Arctic Ocean. Besides, seasonal variability is also visible in the southern periphery of the Beaufort Sea, indicating possible exchanges between the marginal seas and the Beaufort Sea.

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4.1 High-frequency (<30 days) variability

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

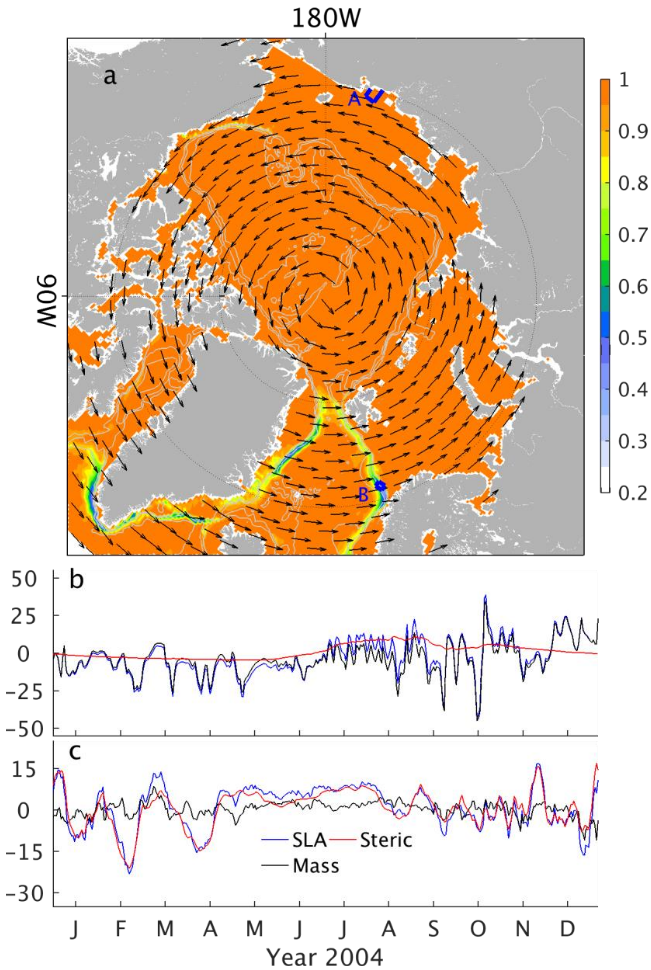

Figure 5. (a) Amplitude (shading) and phase (black vectors) of the transfer function between sea level anomaly and bottom pressure anomaly at periods <30 days. Time series of sea level anomaly (blue lines), mass component (black lines), and steric component (red lines) averaged in (b) the East Siberian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

Except for the Norwegian Atlantic Current (NwAC) and the East/West Greenland Current (EGC/WGC), the amplitude of the transfer function between sea level and mass component is ~1 (Fig. 5a) in most of the pan-

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Arctic regions. The phases (vectors in Fig. 5a) are ~0 in the entire Arctic Ocean, indicating that the high-frequency sea level variability is mostly barotropic. However, in the strong current regions, including NwAC, EGC, and WGC, an amplitude of the transfer function of ~0.4 is observed, revealing that both barotropic and baroclinic processes contribute to the high-frequency sea level variability.

Subregions in the East Siberian Sea (A in Fig. 5a) and along the NwAC (B in Fig. 5a) are used to reveal details of the high-frequency sea level variability. It is clear that the sea level anomaly in the East Siberian Sea (Fig. 5b) is almost equivalent to the bottom pressure anomaly, and the steric component contributes slightly to the seasonal timescale. Along the NwAC (Fig. 5c), pronounced steric height variability with timescales of 20-60 days is visible, which may be caused by baroclinic instability, and the mass component shows high-frequency variability.

The high-frequency sea level variability is mainly related to wind forcing (Fukumori et al., 1998) at high latitudes. Correlations to the wind forcing and sea level anomalies are used to explain the driving mechanisms of the high-frequency sea level variability. The negative correlations between high-frequency sea level variability and wind stress curl (shading in Fig. 6a) in the Canadian Basin and GIN seas (-0.3) and in the marginal seas (-0.3~-0.5) reveal that local sea level increase/decrease is partially related to convergence/divergence of Ekman transport. Positive correlations (0.2~0.3) are visible along the 1000 m isobath where strong currents exist and stratification is strong. A plausible explanation is that wind stress curl anomalies may likely result in baroclinic instabilities, resulting in the baroclinic component of sea level variability along NwAC, EGC, and WGC (e.g., Fig. 5). In the coastal regions, the pronounced correlation of the along-shore wind stress and sea level anomaly at the high-frequency band indicates that the along-shore wind is essential to produce the significant sea level variability (vectors in Fig. 6a).

Correlations of sea level anomalies in regions A (Fig. 6b) and B (Fig. 6c) to the sea level anomalies (contour), wind stress (vectors), and wind stress curl (shading) demonstrate that the along-shore wind drives water towards the coast through Ekman transport which interacts with topography, rising sea level along the coast. And the sea level anomalies could propagate along the coast.

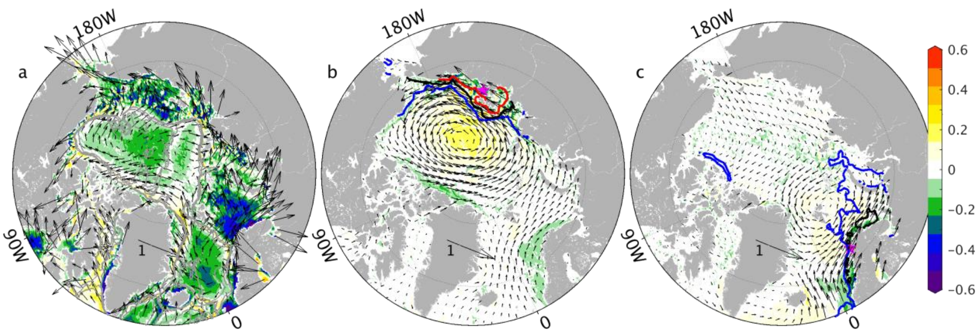

Figure 6. (a) Correlations of sea level anomalies to wind stress curl (shading) and wind stress (vectors) at periods <30 days. Correlations of sea level anomalies in subregions of (a) the East Siberian Sea and (b) along the NwAC (see magenta pentagrams in panels (b) and (c)) to wind stress (vectors), wind stress curl (shading), and sea level anomalies (contours). The blue, black, and red contours denote correlation levels of 0.3, 0.5, and 0.7.

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

4.2 Seasonal variability

Seasonal sea level variability could be related to the redistribution of water from the deep ocean to the marginal seas due to cyclonic/anticyclonic wind stress (Proshutinsky and Johnson, 1997), a seasonal variation of the Arctic Ocean volume (Armitage et al., 2016). In addition, the steric effect due to warm Atlantic inflow and sea ice formation/melting contribute to regional sea level variability in the Barents Sea (Volkov and Landerer, 2013). This section focuses on the spatial-varying Arctic sea level variability at seasonal periods and its mechanism.

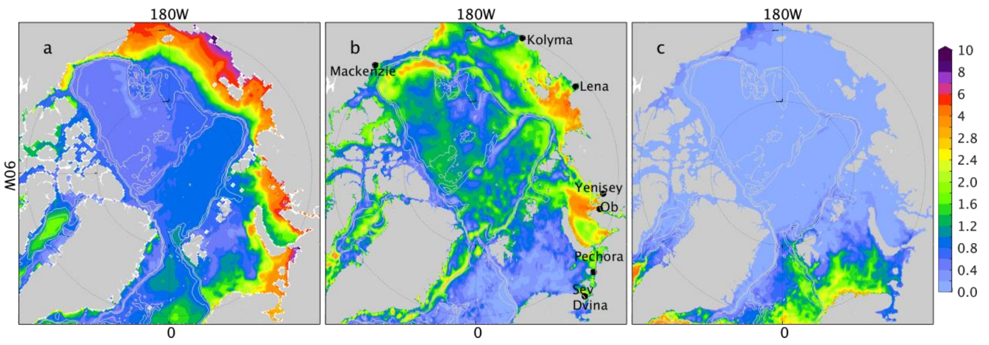

Figure 7. RMS variability of (a) mass, (b) halosteric, and (c) thermosteric components at the seasonal periods.

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

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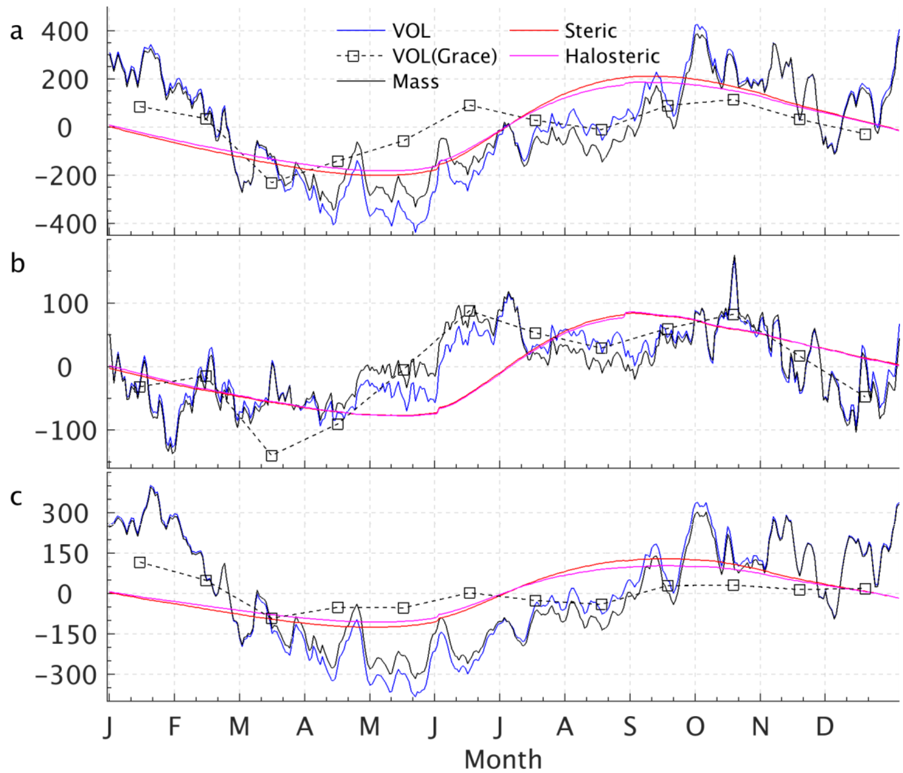

Figure 8. (a) Time series of total volume (VOL) anomaly in the Arctic Ocean (see Fig. 1 for the regions) and the contributions from mass changes (Mass) and steric effects (Steric). The halosteric component (Halosteric) and the GRACE-observed mass component are also shown. Panels (b) and (c) show the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

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

The model simulates more substantial seasonal mass variability than the GRACE measurement. Still, it fails to reproduce the secondary peak from May to July (Fig. 8a), which may relate to river discharge in the marginal seas. Splitting the total Arctic volume changes into contributions from the deep basin and coastal seas, we note that the secondary peak is related to volume changes in the deep basin from May to July (Fig. 8b) in both the model simulation and the GRACE observations. At the same time, volume anomalies are negative in the marginal seas. This antiphase of the volume anomalies in the deep basin and marginal seas seems to be driv-

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en by the cyclonic/anticyclonic wind pattern in the summer/winter season (Proshutinsky and Johnson, 1997). Mean sea level anomalies from June to August (Fig. 9a) and from December to February (Fig. 9b) further reveal the antiphase of the sea level changes between the deep basin and the shallow waters. The mean pattern of wind stress anomalies (vectors in Fig. 9) indicates that wind-driven Ekman transport drives the water toward/away from the marginal seas, resulting in the antiphase of seasonal sea level variability in the deep basin and shallow waters.

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

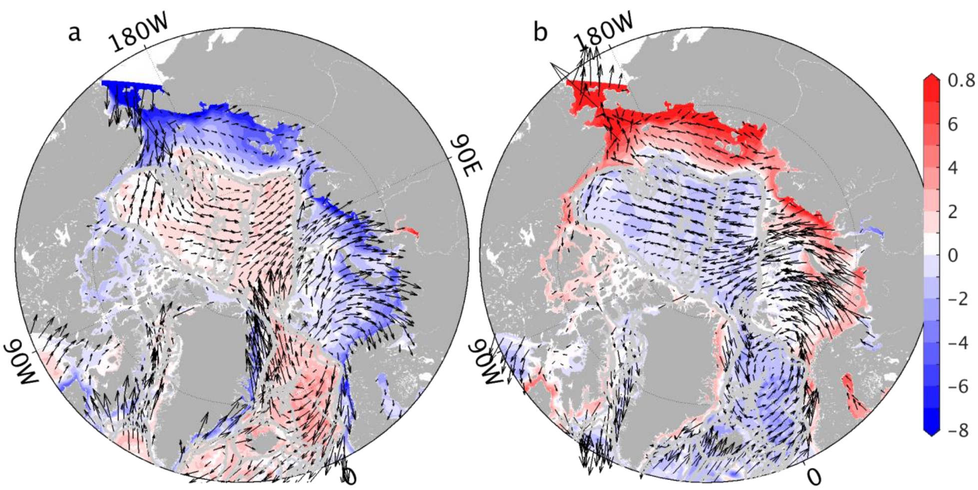

Figure 9. Sea level anomalies (shading) and wind stress anomalies (vectors) averaged from (a) June to August and (b) December to February.

4.3 Decadal variability

The Arctic sea level shows significant decadal variability driven by cyclonic/anticyclonic wind patterns (Proshutinsky and Johnson, 1997), accompanied by freshwater content changes (Häkkinen and Proshutinsky, 2004;Köhl and Serra, 2014). Satellite altimetry observations were used to infer Arctic freshwater content increases (Armitage et al., 2016;Giles et al., 2012;Proshutinsky et al., 2019;Rose et al., 2019). This section examines the spatial variability of Arctic decadal sea level and addresses its relation to the mass, halosteric, and thermosteric components.

It is revealed that the pronounced decadal sea level variability in the Canadian and Eurasian Basins (Fig. 4c) is mainly due to the halosteric effect (Fig. 10b), with the mass components accounting for 20-30%. The thermosteric effect dominates in the GIN Seas since a change from shallow convection to deep convection can lead to temperature changes of more than -0.2 °C over the upper 600 m and salinity changes of 0.02 PSU over the upper 200 m (see Fig. A1 in Brakstad et al., 2019). In the north Atlantic Ocean, the thermosteric effect dominates. At the same time, the halosteric effect compensates for the thermosteric effect, rendering more considerable thermosteric height variability than decadal total sea level variability.

Timeseries of sea level anomalies and its different components in Fig. 10d confirm that sea level variability in the Canadian Basin is mostly halosteric (Armitage et al., 2016;Giles et al., 2012;Morison et al., 2012), and that the thermosteric component contributes with a linear trend (not shown here). In the Eurasian Basin, the mass component, which is likely related to volume exchanges with the Atlantic Ocean and the Barents Sea, also contributes to the interannual sea level variability. The halosteric component shows clearly decadal variability and is in phase with that in the Canadian Basin. The thermosteric component slightly compensates for the halosteric component.

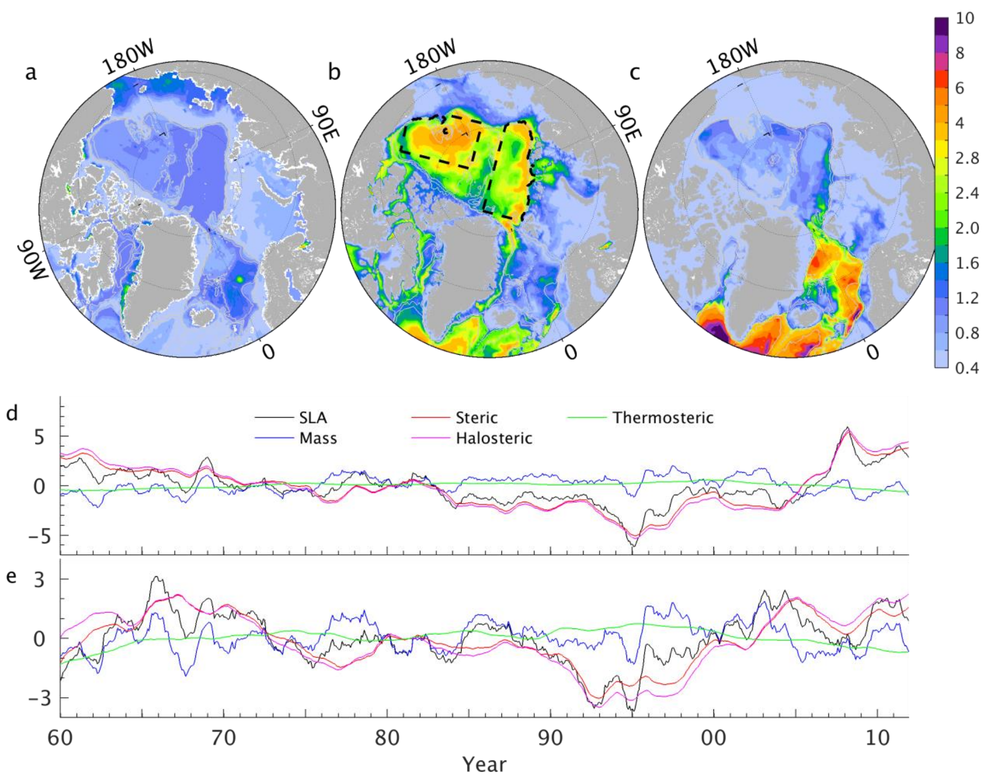

Figure 10. RMS variability at the decadal period of (a) bottom pressure anomaly, (b) the halosteric component, and (c) the thermosteric component. Panels (d) and (e) show the time series of sea level anomaly and mass,

steric, and thermo/halosteric components in the Canadian and in the Eurasian Basins (see the regions in panel (b)), respectively. Linear trends are removed.

5 Capability of the observing system to monitor freshwater content variability

Observing Arctic freshwater content changes remains challenging (Proshutinsky et al., 2019). The results above and previous studies (Giles et al., 2012;Morison et al., 2012;Proshutinsky et al., 2019) have indicated that satellite altimetry could infer freshwater content changes. International efforts try to enhance the profiles observing system, including ice-tethered profilers (ITP), shipboard observations, and moorings. Here, we test their capability to monitor the freshwater changes in an idealized setting in which 1) we do not consider influences of observational errors and 2) we assume the profiles sample the top 800 m and the moorings sample from 65-800 m.

5.1 Satellite altimetry and GRACE measurements

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

$$\Delta FW = \frac{S_2 - S_1}{S_2} \cdot \Delta h = \frac{S_2 - S_1}{S_2} \cdot \left(\eta' \cdot \left(1 + \frac{\rho_1}{\rho_2 - \rho_1} \right) - \frac{\Delta m}{\rho_2 - \rho_1} \right) \tag{5}$$

where ρ_1=1025.0 kg m^{-3} and ρ_2 =1028.0 kg m^{-3} are the mean density in the top and bottom layers. S_1=33.0 PSU is the mean salinity in the top layer, and S_2=34.8 PSU is a reference salinity. η' and Δm are the sea level anomaly and bottom pressure anomalies observations. Morison et al. (2012) suggest that freshwater changes depends on steric height changes linearly and could be approximated by:

$$\Delta FW = \alpha \cdot \eta'_s \tag{6}$$

where α is an empirical constant estimated from in-situ profile observations and is set to 35.6.

As shown in Fig. 11, freshwater content changes and the two estimates show similar decadal variabilities, but differences remain in the seasonal and long-term trends. Since the halosteric effect dominates the steric effect, estimation using Eq. (6) matches the seasonal freshwater cycle very well (red and black lines), considering the amplitude and phase. However, it overestimates the long-term trend (the difference between the black and red dashed lines) since Eq. (6) attributes thermosteric effect (mainly a linear trend) to freshwater changes. Eq. (5) infers a much more substantial seasonal variability of freshwater content, and the phase does not always match the real freshwater content changes (blue and black lines).

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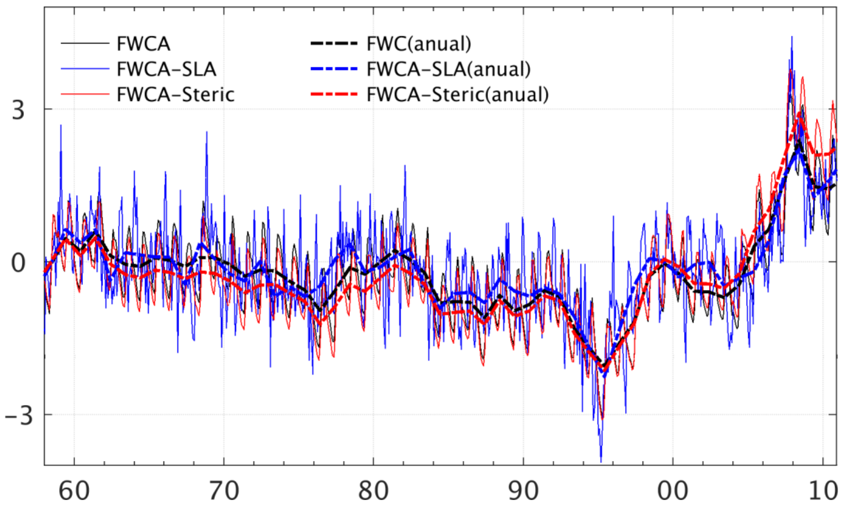

Figure 11. Freshwater content anomalies (10^3 km^3) and approximated based on Eq. (5) in blue and Eq. (6) in red using the monthly output. The thick dashed lines are the annual mean values.

Eq. (5) assumes the isopycnal adjusts simultaneously with sea level anomaly, which may not apply in the presence of baroclinic effects. In order to illustrate the limitation of Eq. (5) we take the differences between Feb. 2003 and Sep. 2002 (in which Eq. (5) fails to reproduce the phase and the amplitude of freshwater content changes) and between 1994-1996 and 2008-2010 (when Eq. (5) reproduces the freshwater changes well).

From Sep. 2002 to Feb. 2003 (Fig. 12a), anticyclonic wind stress anomalies occur in the Beaufort Sea, resulting in positive SLA through Ekman transport. However, freshwater content is reduced during this period. The salinity difference averaged over the central Arctic Ocean reveals that salinity increases in the top 30 m caused by ice formation. At the same time, the isopycnal (27.9 kg m^{-3}) does not deepen (Fig. 12c) as predicted by Eq. (5). The assumption that freshwater content changes are captured by freshwater column thickness 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.

From 1994-1996 to 2008-2010, anticyclonic wind stress anomalies appear in the Canadian Basin, accompanied by positive SLA and freshwater content anomalies (Fig. 12b). During that period, Ekman pumping deepens the isopycnals (blue and red lines in Fig. 12), accumulating more freshwater and reducing the local salinity over the top 300 m (Fig. 12d). In this scenario, the water column thickness change dominates the freshwater 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. (5) captures the interannual freshwater content changes using the satellite altimetry observations. Therefore, caution needs to be taken when inferring Arctic Ocean freshwater content changes using satellite altimetry observations and GRACE measurements. In addition, Figs. 12b and 12c also indicate that Eq. (5) can be only used in the deep basin of the Canadian Basin where wind drives the sea level changes and the deepening/shoaling of the isopycnals.

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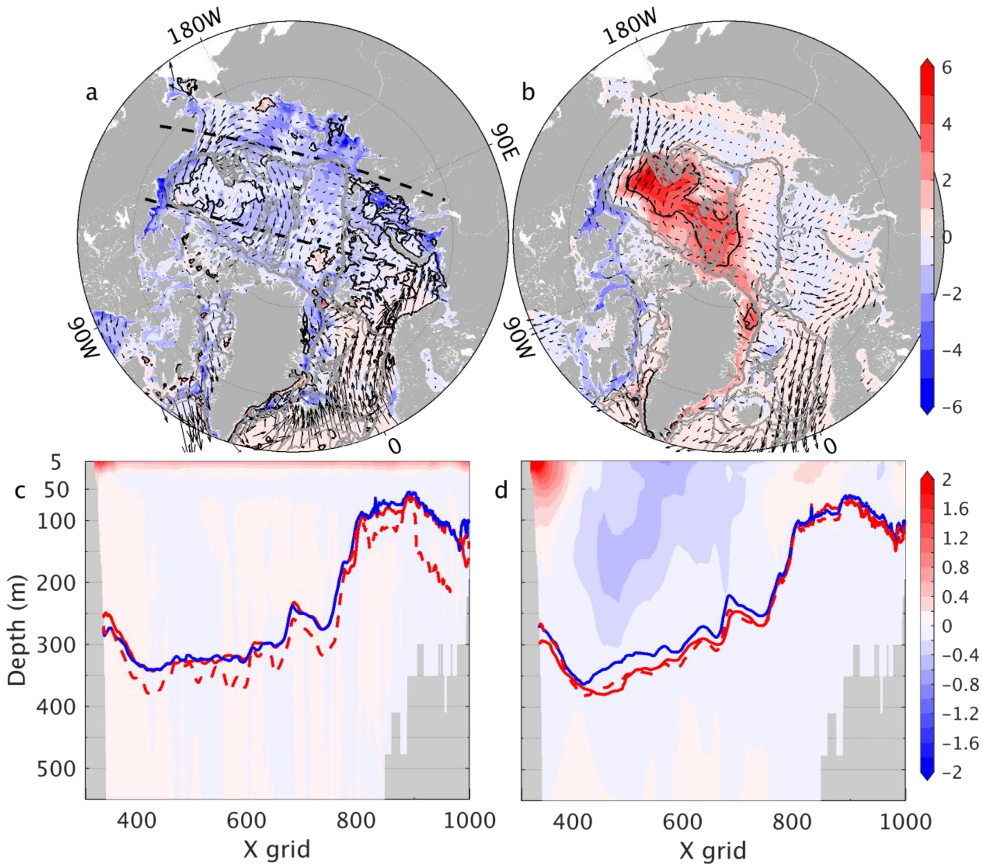

Figure 12. The differences of freshwater content (shading), sea level anomaly (0.15 m contour, black lines), and wind stress(vectors) between (a) Feb. 2003 and Sep. 2002, (b) 1994-1996 and 2008-2010. Panels (c) and (d) are the corresponding salinity differences (shading) average over the central Arctic Ocean (black dashed lines in panel (a)). The blue lines denote the 27.9 kg m^{-3} isopycnal in Sep. 2002 and 1994-1996. The red lines and red dashed lines are the 27.9 kg m^{-3} isopycnal and the diagnosed lines with SLA and Eq. (5) in Feb. 2003 and 2008-2010.

5.2 In-situ profilers

In-situ profilers measure salinity directly, but they are limited by sea ice presence. The endeavor of polar expeditions and the evolving measurement techniques (e.g., ITP) have generated a large number of hydrographic data in the central Arctic and subarctic seas (e.g., Behrendt et al., 2018). This section examines to what extent existing hydrographic observations could help reveal Arctic freshwater content changes and identify observational gaps. Based on the spatiotemporal distribution of profiles in the study of Behrendt et al. (2018) and an ensemble optimal interpolation (EnOI) scheme (Evensen, 2003;Lyu et al., 2014), we test to what extent existing profiles could help to reconstruct the "true" state (here the ATLARC08km simulation) during the periods 1992 to 2012. Details of the EnOI scheme are given in Appendix A.

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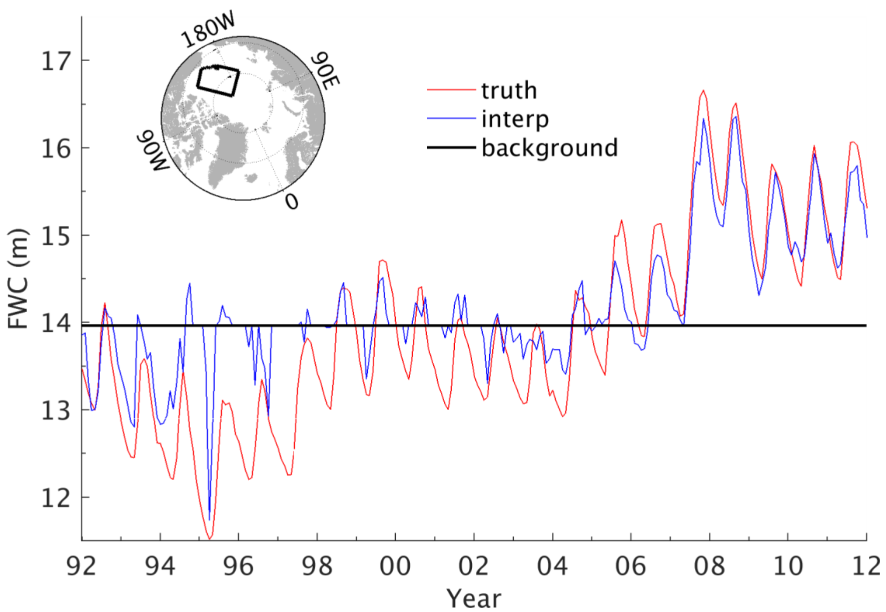

Figure 13. Mean freshwater content (in meter) in the Canadian Basin (enclosed by the black line in the top sub-plot) from the background state, the "truth", and the optimal interpolation reconstructed state (see legend).

As shown in Fig. 13, the sparse in-situ profiles help bring the freshwater content in the background state close to the "truth" state. However, it is not until 2007 that the reconstructed state reproduces the seasonal to inter-annual freshwater content variability in the Canadian Basin, benefiting from the increasing number of research activities and international collaborations. In Fig. 14, we further examined RMS errors of freshwater content depending on geographic locations from 2007 to 2012. Besides the Barents Sea, more significant errors remain in coastal areas due to the lacking of in-situ profiles. In the Laptev Sea and the Alaska coast, we note pronounced errors extending from the coasts to the deep basin, underlining the observing requirements.

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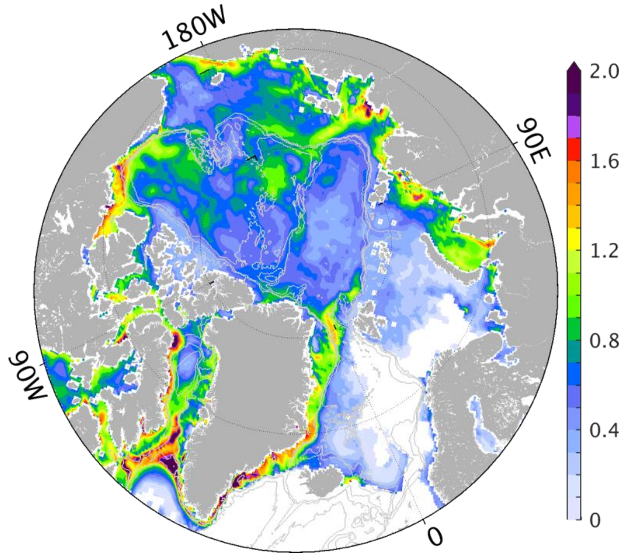

Figure 14. Root mean square errors of freshwater content between the reconstructed state and the "truth".

The above results highlight that the increase of hydrographic observations have enhanced our ability to reconstruct the Arctic freshwater content changes since 2007. A lack of hydrographic observations in the coastal areas results in significant errors in the marginal seas, which require extensive international collaborations.

6 Summary and conclusions

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

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

The seasonal sea level variability is dominated by volume exchanges with the Pacific and Atlantic Oceans and the redistribution of the water by wind stress. Halosteric effects due to river runoff and ice melting/formation are also pronounced in the marginal seas and seasonal sea ice extent regions. Peralta-Ferriz and Morison (2010) demonstrated that river runoff and EmP drive the seasonal cycle of the Arctic bottom pressure. Although the virtual salt flux parameterization could not mimic the influences of volume input from rivers and surface fluxes, the model simulations still simulate much stronger seasonal mass anomalies than the observa-

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tions from GRACE. Either volume exchanges with the Pacific and Atlantic Oceans or volume input from river runoff and EmP are large enough to cause the Arctic Ocean's seasonal volume variability. They should work together, resulting in the Arctic seasonal volume variability. We speculate that using river runoff and EmP as volume flux, rather than the virtual salt flux, could likely improve the volume and sea level variability in the marginal seas from April to July, since the volume inputs from river runoff could alleviate the negative volume anomalies in the marginal seas caused by wind.

At decadal timescales, the model simulations further confirm that the pronounced sea level variability in the central Arctic Ocean, especially in the Canadian Basin, is mainly a halosteric effect. Using the satellite altimetric observations and GRACE observations, the method of Giles et al. (2012) could infer the freshwater content changes in the Canadian Basin very well at timescales longer than one year since isopycnal requires time to adjust to sea level changes. Inferring freshwater content changes using a linear relation of freshwater content and steric height (Morison et al., 2012) reveals both the interannual and the seasonal variability of freshwater content. However, cautions need to be taken since the method also attributes the thermosteric effects to halosteric effects, resulting in an additional linear trend. In addition, uncertainties in the satellite altimetric and GRACE measurements make the estimation more complicated and introduce significant uncertainties in the steric effects and freshwater content estimation (Ludwigsen and Andersen, 2021).

The increasing number of international collaborations and new measurement techniques have generated a large number of profiles. From reconstructing the salinity with synthetic observations, we note that the insitu profile system seems to capture the seasonal freshwater variability since the year 2007, encouraging further Arctic data synthesis studies (Behrendt et al., 2018;Cheng and Zhu, 2016;Steele et al., 2001) with more complicated interpolation methods. In addition, international collaborations need to be enhanced to fill in the observational gaps in the marginal seas. Further, observing system simulation experiments (e.g., Lyu et al., 2021;Nguyen et al., 2020) should be performed coordinately to develop an autonomous observing system in the Arctic Ocean.

7 Data availability

The data used to create the plots in the paper are available at Pangaea (https://issues.pangaea.de/browse/PDI-22940). To access results of the two high resolution ATLARC model simulations, please contact Dr. Nuno Serra at https://www.ifm.uni-hamburg.de/en/institute/staff/serra.html. Observational data were retrieved from publicly available sources and are listed in the text.

Author contribution. G. Lyu performed the analysis and wrote the paper. N. Serra performed the model simulations. D. Stammer proposed this study and M. Zhou provided advice on the analysis.

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

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Appendix A

An EnOI Scheme

We use an EnOI scheme (Cheng and Zhu, 2016) to reconstruct the salinity in the Arctic Ocean using synthetic observations. The analysis state φ^a is a linear combination of a background field φ^f and in-situ observations *d*:

$$\varphi^a = \varphi^f + K(d - H\varphi^f) \quad \text{(A1)},$$

where H is a transfer matric that maps model state from model space to observation space. In this study, the background field of salinity φ^f is taken as the mean salinity over the period 1992-2012. K is the Kalman gain, calculated as:

$$K = A'A'^T H^T (HA'A'^T H^T + \gamma\gamma^T)^{-1} \quad \text{(A2)}.$$

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

The use of ensemble members to approximate the background error covariance ($A'A'^T$) will inevitably introduce long-distance correlations and propagate the observational information incorrectly over a much longer distance. Therefore, we introduce a Gaussian filter as a function of the distance between observational locations and the model grid and an influencing radius to ensure that only observations within the influencing radius of a model grid point could modify the analysis state.

Taken the "true" salinity state from August 1992 and observation locations from the year 2008 (black dots in Fig. A1a), we test the impacts of the influencing radius on the analysis field. The background state is more saline than the truth (Fig. A1a). With a 300 km influencing radius (Fig. A1b), the analysis state reduces the errors near the observations while significant errors remain in regions far away from observations. Increasing the influencing radius to 1000 km, we see that salinity errors in the marginal seas, north pole areas and the Baffin bay are reduced (Fig. A1c). A 2400 km influencing radius further reduces salinity error in the Canadian

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Arctic Archipelago (Fig. A1d). However, only slight improvements are observed in the central Arctic Ocean, and errors in the Kara Sea are slightly increased. Since we focus on the Arctic freshwater content variability, we use a 1000 km influencing radius throughout this study.

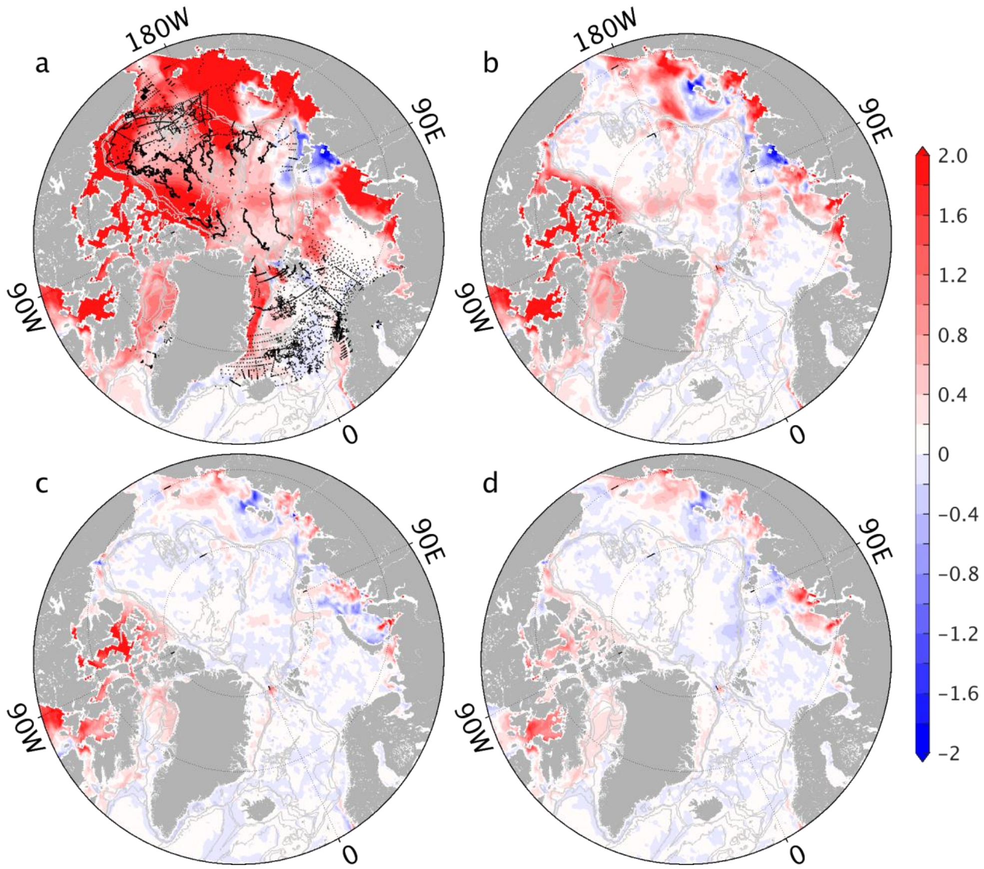

Figure A1. Example of sea surface salinity difference between (a) the background and the truth, (b) the analysis with an influencing radius of 300 km and the truth, (c) the analysis with an influencing radius of 1000 km and the truth, and (d) the analysis with an influencing radius of 2400 km and the "truth". Black dots in panel (a) denote the locations of synthetic observations, sampled using sites of the observations from year 2008.

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