



The role of Arctic gateways on sea ice and circulation in the Arctic and North Atlantic Oceans: a sensitivity study with an ocean-sea-ice model

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Abstract

The impact of changes in volume, heat and freshwater fluxes through Arctic gateways on sea ice, circulation and fresh water and heat contents of the Arctic and North Atlantic Oceans is not fully understood. To explore the role played by each gateway, we use a regional sea-ice ocean general circulation model with a fixed atmospheric forcing. We run sensitivity simulations with combinations of Bering Strait (BS) and Canadian Arctic Archipelago (CAA) open and closed inspired by paleogeography of the Arctic. We show that fluxes through BS influence the Arctic, Atlantic and Nordic Seas while the impact of the CAA is more dominant in the Nordic Seas. In the experiments with BS closed, there is a change in the surface circulation of the Arctic with a weakening of the Beaufort Gyre by about thirty percent. As a consequence, the Siberian river discharge is spread offshore to the west, rather than being directly advected away by the Transpolar Drift. This results in a decrease of salinity in the upper 50 m across much of the central Arctic and East Siberian and Chukchi Seas. We also find an increase in stratification between the surface and subsurface layers after closure of BS. Moreover, closure of the BS results in an upward shift of the relatively warm waters lying between 50 and 120 m, as well as a reorganization of heat storage and transport. Consequently, more heat is kept in the upper layers of the Arctic Ocean, thus increasing the heat content in the upper 50 m and leading to a thinner sea ice cover. The CAA closing has a large impact on sea ice, temperature and salinity in the subarctic North Atlantic with opposite responses in the Greenland-Iceland-Norwegian Seas and Baffin Bay. It is also found that CAA being open or closed strongly controls the sea ice export through the Fram Strait. In all our experiments, the changes in temperature and salinity of the Barents and Kara Seas, and in fluxes through Barents Sea Opening are relatively small, suggesting that they are likely controlled by the atmospheric processes. Our results demonstrate the need to take into consideration the fluxes through the Arctic gateways when addressing the ocean and climate changes during deglaciations as well as for predictions of future climate.

1 Introduction

The Arctic Ocean is a small sea-ice-covered basin with broad shallow shelves and significant freshwater input (e.g. Haine et al. 2015; Fig. 1a). It is connected to the Pacific Ocean through the Bering Strait (BS), and communicates with the Atlantic Ocean via Fram Strait, the Barents Sea Opening (BSO), and the Canadian Arctic Archipelago (CAA). The volume, heat and freshwater fluxes through these straits control the freshwater and heat balances in the Arctic, and also impact the North Atlantic Ocean. Changes in the volume fluxes at any of the Arctic gateways can affect the balance of potential vorticity in the Arctic and, therefore, its circulation (Yang 2005; Joyce and Proshutinsky 2007). Moreover, changes in the freshwater and heat fluxes through Arctic gateways can affect the distribution of the Arctic sea

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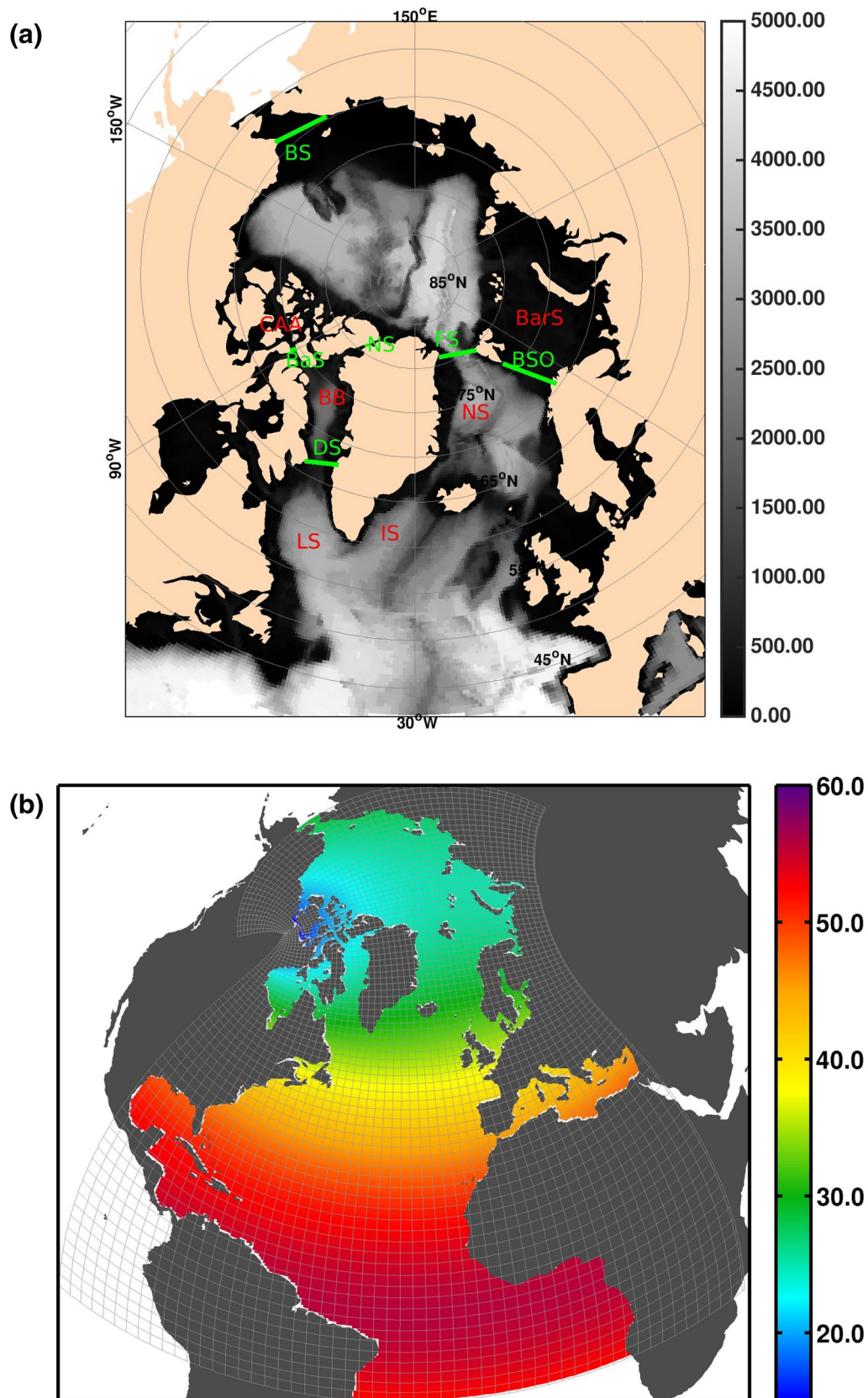


Fig. 1 **a** Map showing model bathymetry as well as main Arctic Gateways (in green) as well as other geographical regions mentioned in the text. *BS* Bering Strait, *FS* Fram Strait, *BSO* Bering Strait Opening, *DS* Davis Strait, *NS* Nares Strait, *BS* Barrow Strait, *LS* Labrador Sea, *BB* Baffin Bay, *CAA* Canadian Arctic Archipelago, *IS* Irminger Sea, *NS* Nordic Seas, *BS* Barents Sea; **b** ANHA2 configuration mesh, with colours showing model horizontal resolution in km

ice (e.g. Belkin et al. 1998; Woodgate et al. 2010) and alter the surface water salinity and density, thus the rate of dense-water formation in both the Greenland-Iceland-Norwegian (GIN) Seas and the Labrador Sea and the strength of the Atlantic meridional overturning circulation (AMOC; e.g. Belkin et al. 1998; Wadley and Bigg 2002; Hu et al. 2010; Yang et al. 2015). For instance, enhanced freshwater export through Fram Strait and the CAA was suggested to play a role in enhancing the 1970s and 1980s Great Salinity Anomalies of the subpolar North Atlantic (Belkin et al. 1998). Examining the evolution of the climate system in a coupled climate model study, Koenigk et al. (2007) predicted a strong reduction in the deep convection of the Labrador Sea due to enhanced freshwater export from the CAA. Exchanges between the Arctic and Atlantic Oceans take place through the Nordic Seas (GIN seas and the Barents Sea) and Baffin Bay which are not just passive intermediaries, but they also play roles in controlling the dynamics and the properties of water masses being exchanged (Eldevik and Nilsen 2013; Münchow et al. 2015; Grivault et al. 2017).

A fundamental response of changing climate includes sea level. Sea level has been rising in concert with increasing temperatures over the last century, with forecasts of continued rise over the next centuries (IPCC 2013). Much more rapid sea level change may occur with the loss of terrestrial ice sheets. For example, there is about 7.36 m in sea level rise potential from the Greenland Ice Sheet melt (Vaughan et al. 2013). The geological past has shown that changes in the sea level and/or the gateway geometry (width and depth) related to both eustatic sea level and isostatic adjustments also altered the gateway fluxes. Examples include the global sea level drop associated with the growth of ice sheets, in addition to ice dams between Greenland and Ellesmere, which closed off the shallow and narrow gateways of the Arctic (e.g., England et al. 2006). They also include isostatic depressions related to ice load, which result in large amplitude regional sea level changes when ice vanishes due to both global sea level rise with the melting of ice sheets and the delays of crustal adjustments, which may last long after deglaciation (e.g., Khan et al. 2015). Hence BS, being only 50 m deep, and the CAA channels with depths of between 100 and 200 m were affected during the glacial–interglacial transitions. For example, during the Last Glacial Maximum (LGM; ~23,000 to 19,000 years before the present; Mix et al. 2001), the global sea level was lower than at present by around 120 m. After the deglaciation, the BS was closed

until about 11,000 years ago (Elias et al. 1996; Clark et al. 2014; Jakobsson et al. 2017) and remained shallower than at present by more than 10 m until about 8000 years ago (Manley 2002). The narrow and shallow channels of CAA were also closed as they were mostly covered by the Innuitian ice sheet during the last glacial stage and occupied by ice streams until about 8500 years before the present (Dyke 2004; England et al. 2006). Even Nares Strait which is the deepest channel of CAA (~200 m) was either nearly closed (Zreda et al. 1999) or closed (Jennings et al. 2011). The timing of BS and CAA opening and the flow related to sea level rise during the final phase of the deglaciation are not precisely known, but their impact deserves to be explored as they probably account for a large part of the postglacial reorganization of the North Atlantic Ocean circulation (e.g., Hu et al. 2010).

Finding the relative importance of each of the Arctic gateways in transporting heat and freshwater, and their respective impact on the Arctic and North Atlantic Ocean, poses a fundamental question for both paleo and future climate studies. Moreover, most global climate models have their shortcomings in simulating the changes in the Arctic fluxes due to their coarse resolution and poor representation of the fluxes through BS and the CAA (e.g., Fuentes-Franco and Koenigk 2019). Here we intend to provide a fundamental understanding of the role played by BS and the CAA in the ocean dynamics that governs the Arctic and the North Atlantic Ocean. For this, we perform sensitivity experiments with various gateway configurations for the BS and CAA inspired by the paleogeography of the Arctic. Our goal is to better understand the impact of modulating BS and CAA fluxes on circulation, sea ice properties, surface salinity and temperature, freshwater and heat fluxes in the Arctic and the Atlantic Ocean. This will allow us to have a better founded understanding of the deglaciation as well as predictions for the future with regards to the changes in the gateway fluxes.

The impact of closing the BS and CAA has been the subject of earlier studies (e.g. Wadley and Bigg 2002; Hu et al. 2015), suggesting that blocking the BS freshwater inflow and/or closing the CAA strengthened the AMOC. However, none of these studies have considered the importance of non-linear responses to closing one strait but not another. More recent studies have indicated the need to represent details of the passages within the CAA (Wekerle et al. 2013) as well as the fact that variations in CAA and Fram Strait export are out of phase (Jahn et al. 2010; Zhang et al. 2016). Moreover, the impact of BS and/or CAA closure on the sea ice properties and freshwater fluxes have not been investigated. Otto-Bliesner et al. (2017) used a coupled climate model to look at the role of the Arctic gateways in the context of the late Pliocene climate. Their results suggested that closing Arctic gateways inhibited freshwater export to the Atlantic, strengthening the AMOC. They also pointed out the

importance of considering the non-linear climate response to closing straits. Unlike Otto-Bliesner et al. (2017), we do not use a coupled climate model, but instead an ocean sea ice model with a relatively high spatial resolution. Although this limits us in the documentation of ocean–atmosphere feedbacks, regional ocean modelling is necessary to run simulations at sufficiently high resolution to include the multiple complicated passageways associated with the CAA.

2 Methods

We use a regional configuration of the ice-ocean model, NEMO (Nucleus for European Modelling of the Ocean model; Madec, G. and the NEMO Team 2008) version 3.4 extracted from NEMO ORCA05 (Barnier et al. 2007). The domain covers the Arctic Ocean, the peripheral seas of the Arctic and the Northern-Hemisphere Atlantic down to 20° S. It has a tripolar grid with a resolution of 54 km (0.5°) at the equator which increases to 24 km in the Arctic Ocean near the model pole (Fig. 1b). In the vertical direction, *z*-coordinates are used, with 50 uneven layers with their thicknesses starting from ~1 m at the surface (8 levels in the first 10 m) and increasing smoothly to around 458 m at the bottom. The bathymetry data is constructed from the 2-min gridded global relief data ETOPO2 of the US National Geophysical Data Center (NGDC). The interpolation onto the model grid was produced by taking the median depth of all the original grid points falling into an ORCA05 grid box area.

The ocean component of NEMO is a primitive-equation ocean model. The model uses a linear free surface and a partial step representation of bottom topography. The lateral subgrid-scale mixing parameters include bilaplacian eddy viscosity of $-12 \times 10^{11} \text{ m}^4 \text{ s}^{-1}$ for the momentum diffusion and Laplacian eddy diffusivity of $600 \text{ m}^2 \text{ s}^{-1}$ for the tracer diffusion. The turbulent kinetic energy (TKE) closure scheme was used to compute the vertical eddy viscosity and diffusivity coefficients (Madec and the NEMO Team 2008). The background (minimum) values of the vertical mixing parameters were set to $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ and $1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ for the vertical eddy viscosity and diffusivity, respectively. Additionally, the background vertical diffusivity is reduced from the ice-free to the ice-covered regions as a function of ice concentration and becomes $1 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ in the fully ice-covered regions (Barnier et al. 2012). The sea ice component of the model, which is interactively coupled with the ocean part, is the thermodynamic-dynamic sea ice model LIM2 (version 2 of Louvain-La-Neuve ice model; Fichefet and Morales Maqueda 1997; Bouillon et al. 2009). The thermodynamic part of LIM2 has one layer of snow on top of two layers of ice to determine sensible heat storage and vertical heat conduction. Ice dynamics are found from a momentum balance driven by the atmosphere and ocean

by assuming sea ice as a two-dimensional elastic-viscous-plastic continuum (Hunke 2001).

All experiments have the same initial conditions and surface forcing. The initial conditions for temperature and salinity come from the Polar Science Center Hydrographic Climatology (PHC3.0; Steele et al. 2001). For the atmospheric forcing fields, a 1-year repeating cycle of forcing from the Coordinated Ocean-ice Reference Experiments Phase 2 (CORE-2; Large and Yeager 2004), known as “normal year forcing”, was used. CORE data are a combination of atmospheric reanalysis and remote sensing products, and the normal year forcing is representative of climatology-like conditions and consists of a single annual cycle of all the data needed to force an ocean-ice model (Griffies et al. 2009). The atmospheric forcing includes 6-h 10 m wind, air temperature and humidity, daily downward longwave and shortwave radiation, monthly precipitation and snowfall. We use present-day forcing to allow us to focus our experiments solely on the impact of closing the various gateways (rather than trying to drive the model with uncertain paleo-atmospheric forcing). The river runoff is from Dai and Trenberth (Dai and Trenberth 2002; Dai et al. 2009).

Our model domain includes two open boundaries; one at the BS and the other at the 20° S in the Atlantic (Fig. 1b). Along these boundaries, salinity, temperature, and zonal and meridional velocities are restored to monthly climatology values obtained from a global ocean simulation (averaged over the period of 1958–2009; Barnier et al. 2007). At the southern boundary, velocity adjustment is applied to keep the volume of the whole domain conserved. At BS, the velocity adjustment is not applied and the volume flux is determined by the upstream boundary conditions during the run. For the experiments where BS is closed, a buffer zone next to the southern boundary is also implemented, comprising 20 grid rows where temperature and salinity are restored to the values prescribed by the initial conditions. This is to damp the numerical noise caused by the mismatch between the prescribed open boundary condition and the interior model solution and is commonly used in NEMO regional simulations. The restoring timescale is 1 day in the inner row of the buffer and decreases to 2 h at the edge of the buffer.

We present four sensitivity experiments with various gateway configurations for the BS and CAA (Table 1): both BS and CAA closed, only CAA closed, only BS closed, and the present-day run with both BS and CAA open. All simulations were spun up for 60 years from the initial conditions to reach cyclo-stationarity assessed by the volume transport through the Arctic gateways (not shown). This is sufficient for the sea ice, freshwater and heat transport through the gateways, and the surface ocean to be at the cyclo-stationarity. To confirm this, we continued the run for the present-day simulation until year 150 and found that gateway transports

Table 1 The experiments in this study and their corresponding gateway configurations

Experiment	Bering Strait (BS)	Canadian Arctic Archipelago (CAA)
Closed-BS-CAA	Closed (–)	Closed (–)
Open-CAA	Closed (–)	Open (+)
Open-Bering	Open (+)	Closed (–)
Present-day	Open (+)	Open (+)

are basically stable from year 35. Results shown in the next section are the mean over years 51–60 of integration for each run.

3 Results

3.1 Evaluation of the model

We begin with a brief evaluation of model fields, focusing on those aspects relevant for our analysis of the impacts of opening and closing the Arctic gateways. In winter, the Arctic and its marginal seas are covered by sea ice (Fig. 2a), as expected. The maximum sea ice extent is $13 \times 10^6 \text{ km}^2$, close to the 1979–2010 mean of $13.6 \times 10^6 \text{ km}^2$ found by Comiso (2012). Note that these values are excluding the sea ice extent in the Pacific Ocean since our model domain does not include those basins in the south of Bering Strait (estimates of sea ice extent for the Seas of Okhotsk and Japan is $1.0\text{--}1.5 \times 10^6 \text{ km}^2$ and $0.5\text{--}1.0 \times 10^6 \text{ km}^2$ for the Bering Sea; Parkinson and Cavalieri 2008). Hence, despite the slight discrepancy, our maximum sea ice extent is in agreement with the observations. Our modelled ice edge positions in the Atlantic sector are consistent with satellite observations (OSI-450 from EUMETSAT Ocean Sea Ice Satellite Application Facility) and other model studies (e.g. Wang et al. 2016a). Over most of the domain, modelled sea ice thickness (Fig. 3a) is also consistent with CryoSat-2 and ICESat estimates of thickness (Laxon et al. 2013; Tilling et al. 2018; Kwok and Cunningham, 2008) and the results from other studies (Vancoppenolle et al. 2009; Hu and Myers 2013; Wang et al. 2016a). The modelled sea ice export from the Arctic through Fram Strait is 75 mSv ($2375 \text{ km}^3 \text{ year}^{-1}$), which is close to the observed value ($2300 \text{ km}^3 \text{ year}^{-1}$) in Serreze et al. (2006).

The model produces reasonable results for the general features of the surface circulation and the main currents. In the Arctic, the main features of the surface circulation, including the anticyclonic circulating Beaufort Gyre and the Transpolar Drift, are well captured (Fig. 4a). As can be seen in the sea surface height (SSH) field, the doming of the Beaufort Gyre (region of highest SSH in the Beaufort

Sea) is located around 75° N and 150° W , in agreement with observations (Giles et al. 2012). The ocean velocity (top 50 m) in the Beaufort Gyre is about $2\text{--}3 \text{ cm s}^{-1}$, and the barotropic transport of the gyre reaches up to 6 Sv. Pacific Water enters through the Bering Strait and continues into the basin interior by multiple routes. We also see the cyclonic flow of Atlantic Water entering the Eurasian basin through Fram Strait and the Barents Sea, although penetration is less than observed (Aksenov et al. 2010). Cold surface temperatures near the freezing point are seen throughout the Arctic (Fig. 5a), while surface salinity ranges from less than 30 on the shelves to around 33.75 in the Beaufort Gyre (Fig. 6a). Due to the resolution used in the current study, the transport of freshwater by eddies is underestimated which leads to higher salinity in our model Beaufort Gyre (Hu et al. 2019). Export of low salinity water from the Russian rivers off the Siberian shelves is simulated (Dmitrenko et al. 2010). In the Baffin Bay, the circulation is cyclonic (Fig. 4a) with thicker ice on the Baffin Island side (Fig. 3a) and the concentration gradient running to the northeast through Davis Strait (Fig. 2a), as observed (Tang et al. 2004). Strong southward ice advection is seen in the East Greenland Current (Fig. 2a), although extending slightly too far offshore in the Greenland Sea. The penetration of the warm and salty Atlantic Water can be seen in the Norwegian Sea, flowing north to the Fram Strait and the BSO, with cold and low salinity water leaving the Arctic in the East Greenland Current (Figs. 5a, 6a).

In the North Atlantic Ocean, the cyclonic circulation in the Subpolar Gyre is reproduced (Fig. 4a) and has a barotropic transport reaching up to 52 Sv in agreement with other model studies (e.g. Marzocchi et al. 2015). However, it does not penetrate as far eastward as observed, and extends too far south. The North Atlantic Current being too zonal in the western basin, and not entering the Labrador Sea east of Cape Farewell. Cold and fresh waters are transported south in the Labrador Current (Figs. 5a, 6a), although too much low salinity water is transported offshore north of Flemish Cap and the tip of the Grand Banks (Myers 2005; Fratantoni and McCartney 2010). Although the East and West Greenland Currents have cold and low salinity waters, they are not as cold and fresh as observed (Sutherland and Pickart 2008; Myers et al. 2009). Deep mixed layer depths are found in the central Labrador Sea and the Nordic Seas southwest of Svalbard (Fig. 7a). These are typical places for convection as found in models and observations, although the mixed layer in the Labrador Sea is too deep and occurs over a broad area (Labrador Sea Group 1998; Chanut et al. 2008; Courtois et al. 2017). This is a typical feature in this region for forced numerical models (Rattan et al. 2010).

We summarize flux exchanges through the Arctic gateways in Table 2 (averaged over year 51–60 of the model). Our inflow through Bering Strait (1.28 Sv) is higher than observed (0.8–1.1 Sv; Woodgate et al. 2005, 2012) but

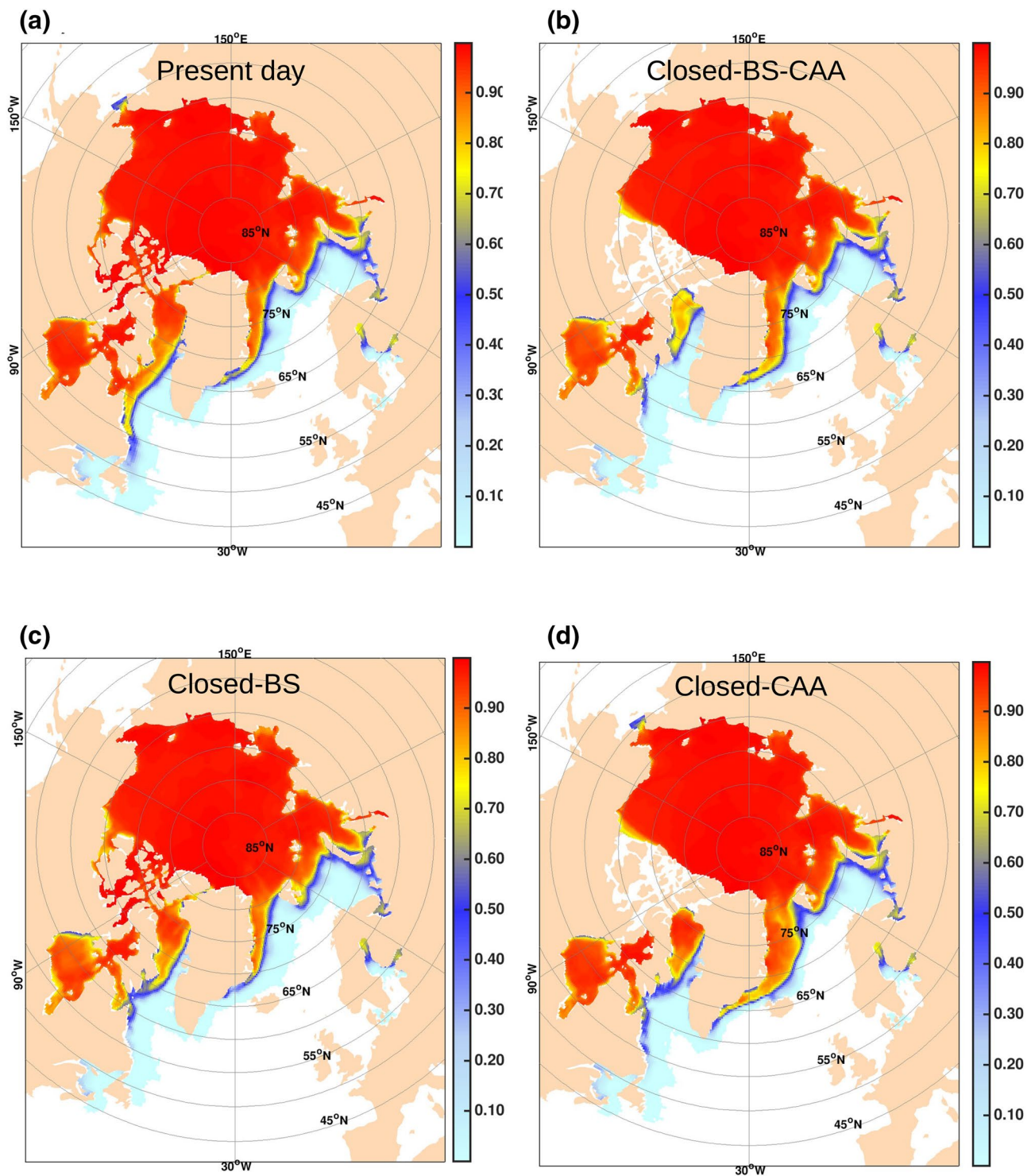


Fig. 2 Model sea ice concentrations in winter, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA

within the range reported in ocean reanalysis experiments of C-GLORSv7 (Pietschnig et al. 2017). The overestimation of Bering Strait inflow in our model, which is also commonly seen in global simulations (e.g., Ilicak et al. 2016),

comes from using the open boundary forcing retrieved from a global NEMO simulation (Sect. 2). In an assessment of the Arctic Ocean in the multi-model study by Wang et al. (2016b), the multi-model mean of Bering Strait transport is

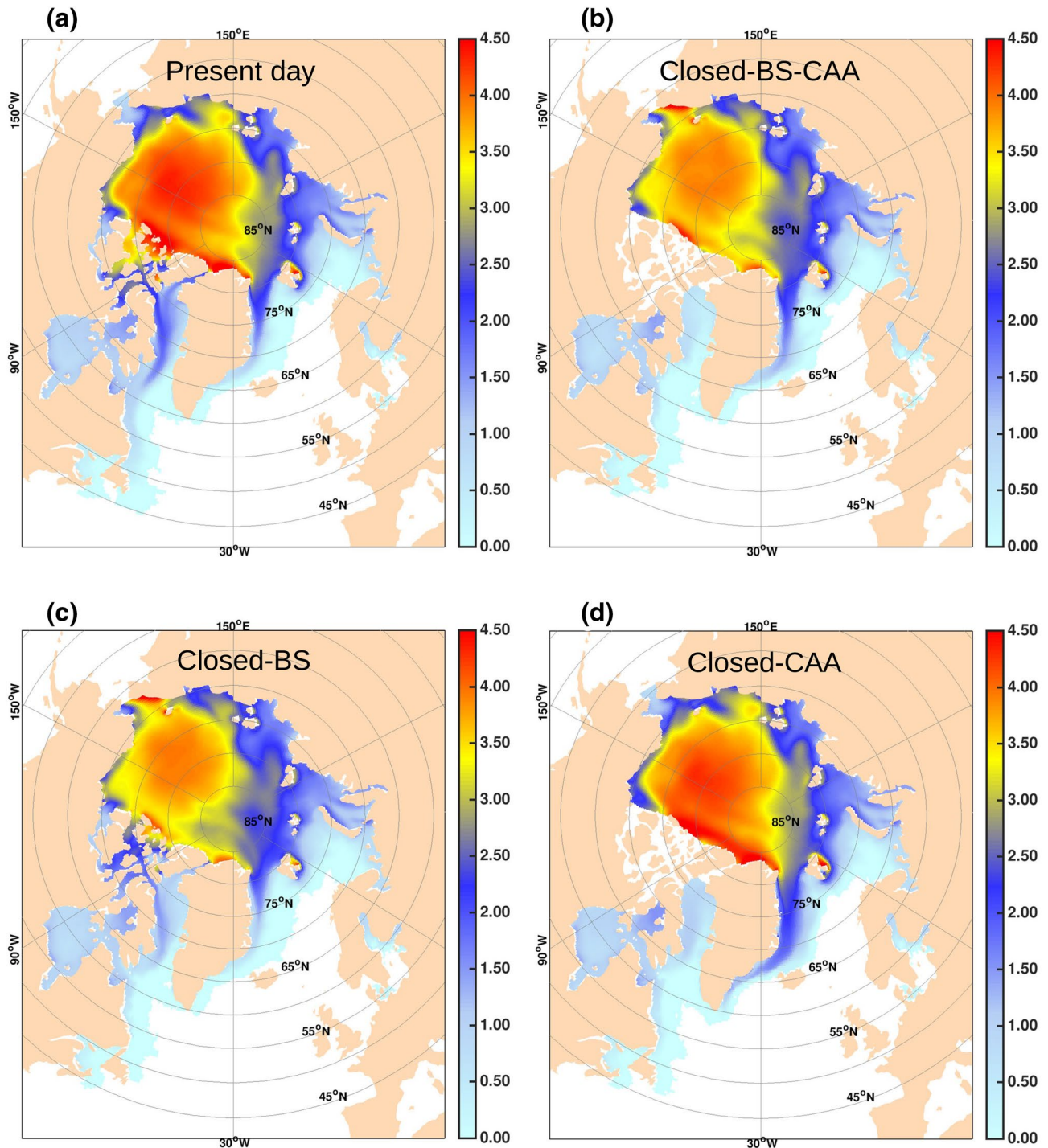


Fig. 3 Model sea ice thickness (m) in winter, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA

about 1 Sv, although the study includes 4 models with values similar to or larger than ours. Our model inflow through the BSO of 2.7 Sv is larger than the observations of 2.0–2.3 Sv (Skagseth et al. 2008; Smedsrud et al. 2013) but consistent with the multi-model mean of 2.7 Sv in Wang et al.

(2016b). We have a strong two-way exchange through Fram Strait in our simulation. Even the net volume transport is weakly north into the Arctic, there is a southward component of 3.8 Sv, which is nearly consistent with observations from Schauer et al. (2008) and other model studies (e.g.

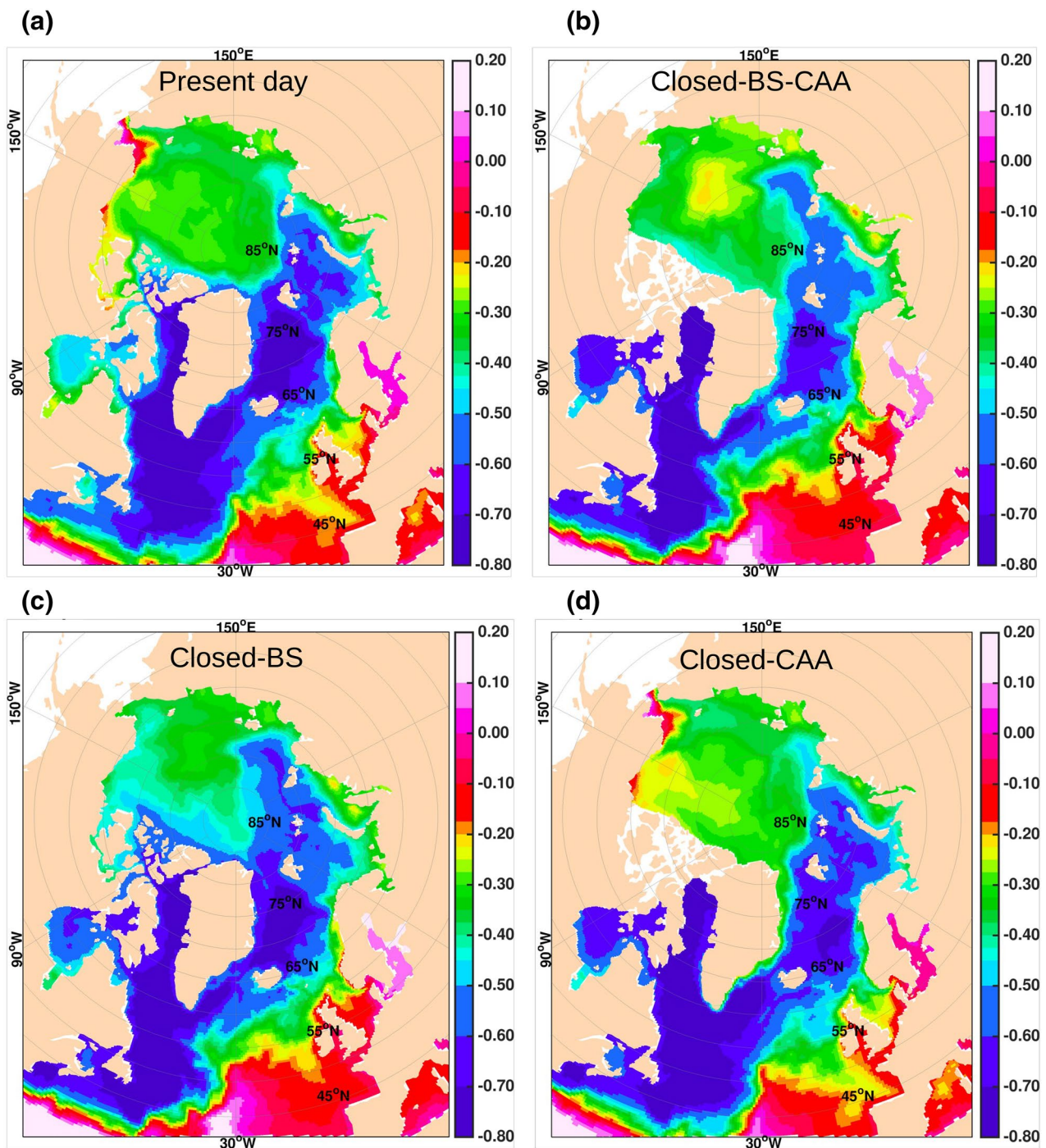


Fig. 4 Annually averaged model sea surface height in m, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Note that the color-bar is non-linearly discrete and there are more contours between -0.45 and 0.25 with 0.025 contour interval

Wekerle et al. 2013; Aksenov et al. 2016). The partitioning of the transport between the two CAA routes is consistent with Wekerle et al. (2013), although the values mentioned in their study are about 40 percent of our estimated transports (2.45 Sv and 2.22 Sv for Lancaster Sound and Nares, respectively).

At Bering Strait freshwater inflow is 88 mSv (relative to 34.8) consistent with the observations (Woodgate et al. 2005, 2012) and the multi-model mean in Wang et al. (2016b). Our heat flux at Bering Strait (2 TW) is lower than the estimated value of 10–20 TW in Woodgate et al. (2012), partly due to using of different reference temperatures in our calculations

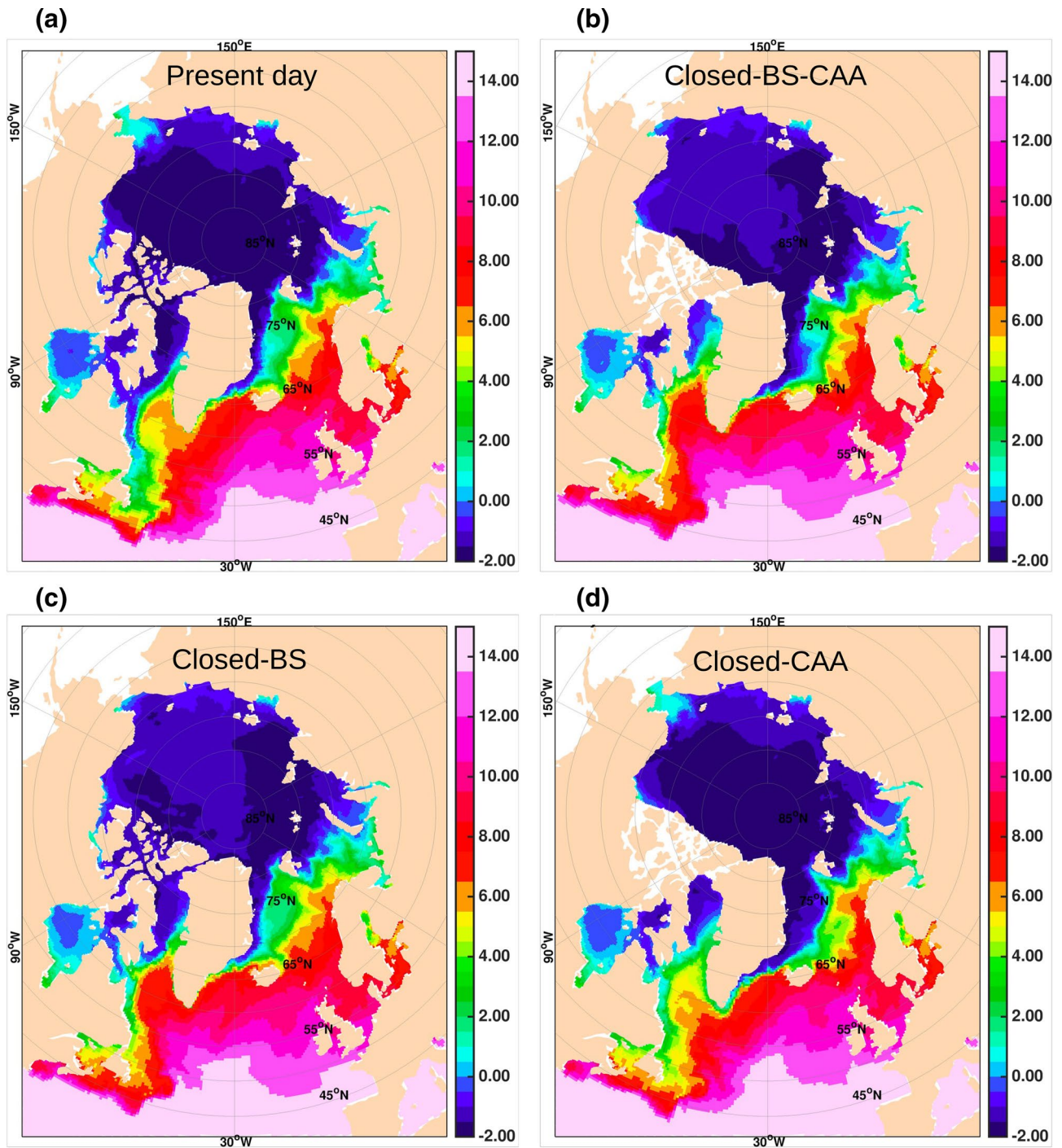


Fig. 5 Annually averaged model top 50 m temperature in degrees C, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Note that the color-bar is non-line-

arly discrete and there are more contours between -2 and 6 with 0.25 contour interval

(0°C in ours versus -1.9°C in Woodgate et al. 2012). As the BSO brings warm salty Atlantic Water into the Arctic, it has a negative freshwater flux of 59 mSv , consistent with the $55\text{--}60\text{ mSv}$ from Smedsrud et al. (2010). The BSO heat inflow of 63 TW also falls in the model-estimated range by

Smedsrud et al. (2010). The model exports fresh water at Fram Strait with an estimate of about 39 mSv which is smaller than the observations (de Steur et al. 2009; Rabe et al. 2009) and the multi-model mean (Wang et al. 2016a). However, 3 models from the multi-model study of Wang et al. (2016a)

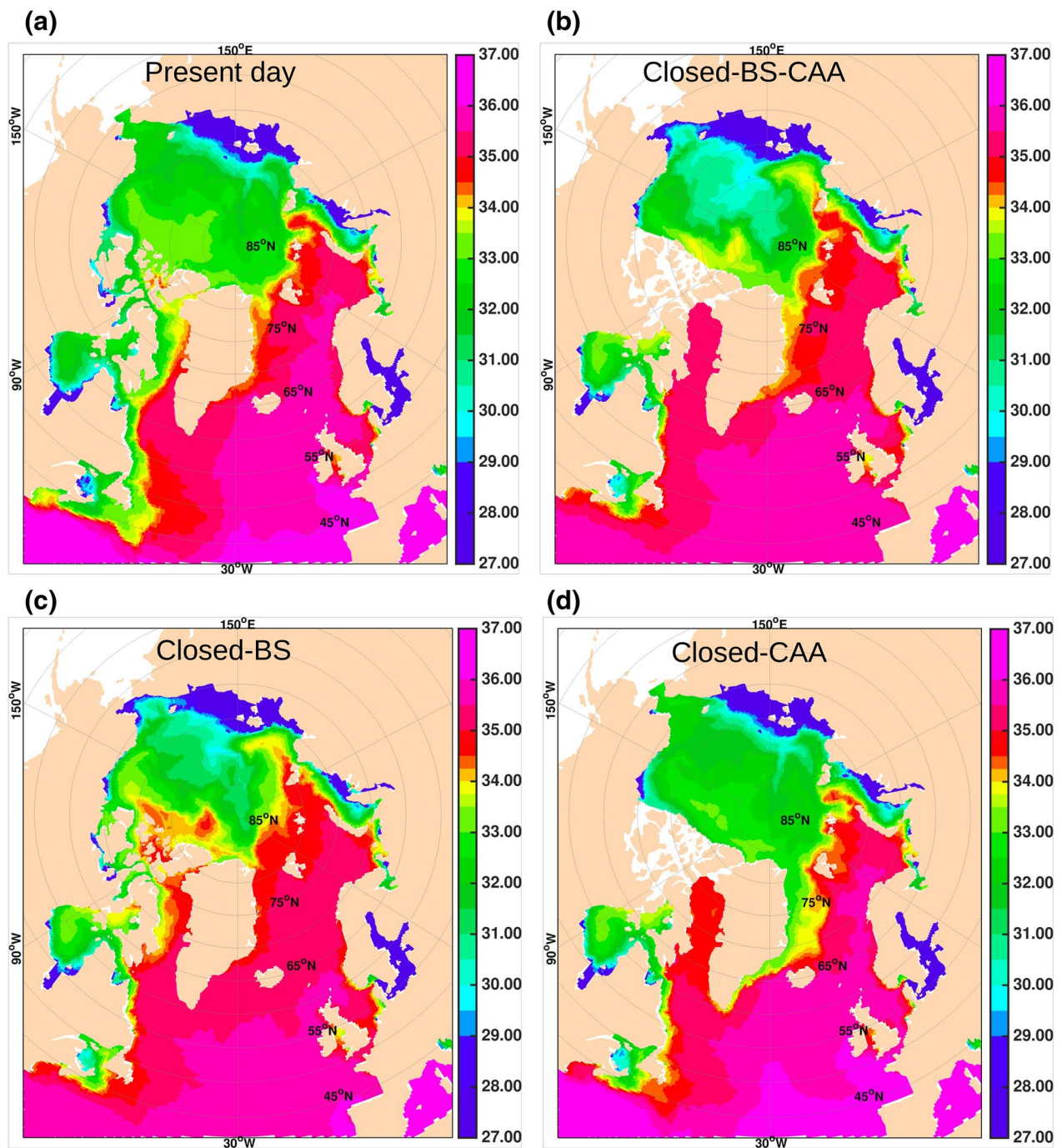


Fig. 6 Annually averaged model top 50 m salinity, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Note that the color-bar is non-linearly discrete and the

contour interval is 0.5 between 29 and 33.5, and is 0.25 between 33.5 and 34.5

as well as the ECMWF Ocean ReAnalysis system (ORAS4, Balmaseda, et al. 2013) have similar, or smaller, freshwater exports than our estimate. The 23 TW of heat input to the Arctic through Fram Strait is about half of the observed (Schauer et al. 2008; Rudels et al. 2008). Our estimate of 96 mSv of southward freshwater export at Davis Strait is well within the

observational range (Curry et al. 2014) and nearly equal to the multi-model mean found in Wang et al. (2016b). The model heat flux at Davis Strait (20 TW) is also within the range of observations (Cuny et al. 2005, 2011). We therefore argue that our coupled ice-ocean model satisfactorily represents Arctic

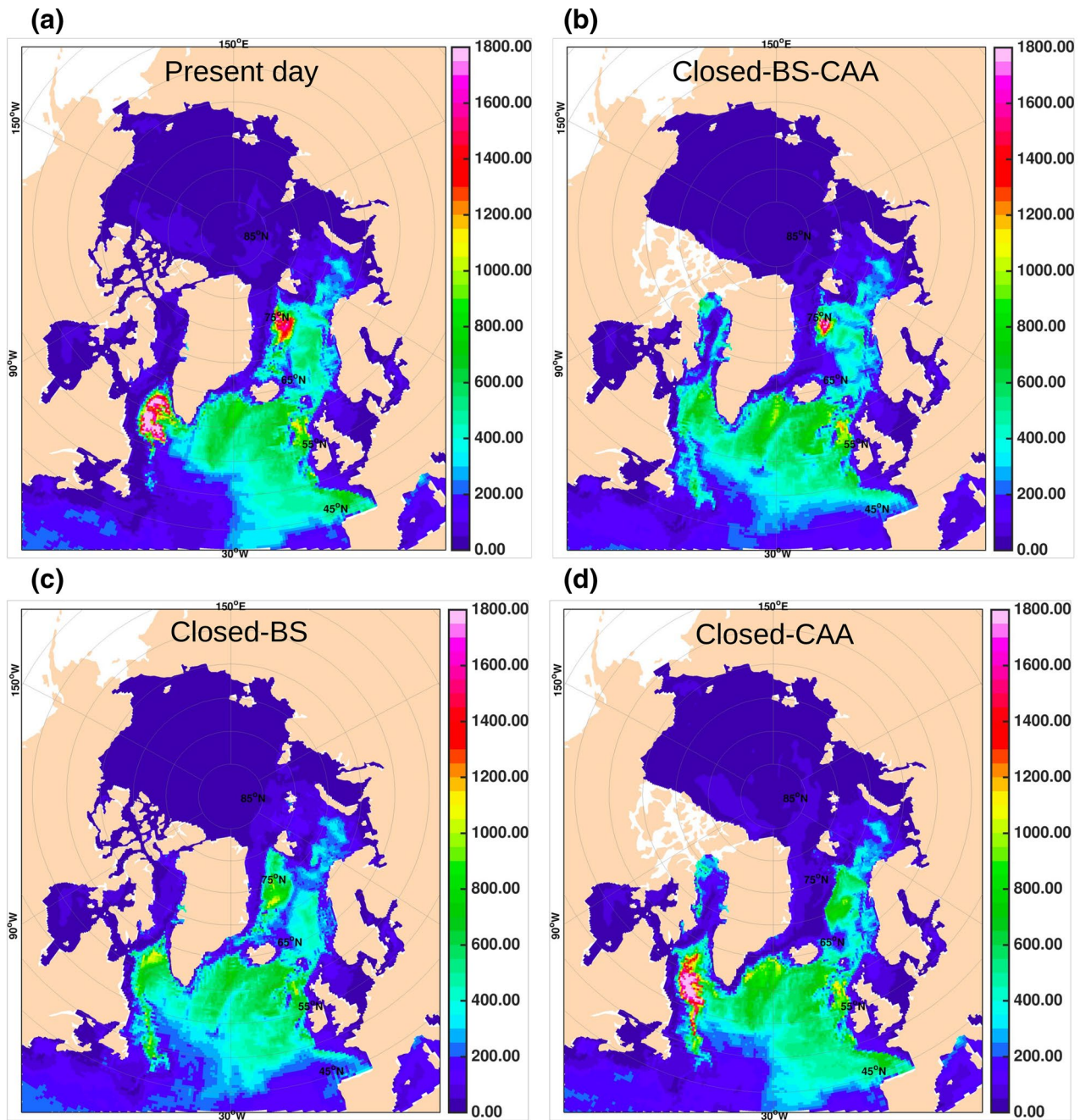


Fig. 7 Winter model mixed layer depth in m, over years 51–60 of integration, for **(a)** present day; **(b)** closed-BS-CAA; **(c)** closed-BS; and **(d)** closed-CAA. Contour interval is 50 m

ocean properties and gateway transports, and can be used to examine the impact of opening and closing those gateways.

3.2 Impact of BS and CAA on the Arctic Ocean

There are many changes within the model variables resulting from closing gateways. In the following, we mainly focus on large-scale and important changes.

Closing BS, regardless of the configuration of the CAA, shifts the center of Beaufort Gyre more to the east (the region of highest sea surface height in the central Arctic in Fig. 4b, c) and weakens the gyre by about thirty percent (based on velocity fields; not shown). This is consistent with Zhong et al. (2015) who suggested that the decreased (increased) freshwater through BS results in the weakening (spin-up) of the Beaufort Gyre. The barotropic transport

Table 2 The modelled yearly-mean net volume, freshwater and heat fluxes (Sv, mSv, TW) through the main gateways for different gateway configurations averaged over years 51–60

	Present-day				BS Closed				CAA Closed				BS and CAA Closed			
	Vol (Sv)	FW (mSv)	Heat (TW)		Vol (Sv)	FW (mSv)	Heat (TW)		Vol (Sv)	FW (mSv)	Heat (TW)		Vol (Sv)	FW (mSv)	Heat (TW)	
Bering Strait	1.28	88	2		0.00	0	0		1.28	87	1.2		0.00	0	0	
Barents Sea Opening	2.70	– 59	63		3.21	– 47	68		2.40	– 57	57		2.83	– 46	60	
Fram Strait	0.53	– 39	23		0.17	– 16	19		– 3.84	– 111	31		– 2.99	– 20	18	
Davis Strait	– 4.67	– 96	20		– 3.52	– 23	30		0.00	– 6	26		0.00	– 6	65	

Positive values mean that the net fluxes are going into the Arctic. The sum of volume fluxes fairly balances with the River Runoff and P-E, although small differences from year to year within the experiments and among experiments are expected. If the freshwater flux has the opposite sign compared to its corresponding volume flux, that means the transported salinity is larger than the reference salinity 34.8. Negative freshwater means that either water less saline than 34.8 is exported or salty water is imported

of the gyre also reduces about 1–2 Sv in agreement with Spall (2020). The Beaufort Gyre is generally seen as a wind-driven gyre with Ekman pumping (e.g. Proshutinsky et al. 2002, 2009). However, since the winds are the same in all our experiments, we suggest that closing BS and the associated change in the density field are responsible for the weakening of the Beaufort Gyre (see “Appendix”). As a result of a weaker and displaced Beaufort Gyre, and in the absence of strong Pacific Water inflow, the Eurasian river runoff changes its pathway and penetrates farther offshore to the west rather than being directly advected away by the Transpolar Drift. Similar changes in the circulation pattern that shifts the path of runoff has been also suggested but under interannual variability of atmospheric forcing (e.g., Jahn et al. 2010). Such changes enhance the freshwater input from the Siberian rivers into the central Arctic and East Siberian and Chukchi Seas. Thus, in the central Arctic and East Siberian and Chukchi Seas (150° E–150° W), salinity (freshwater content) in the top 50 m decreases (increases) by at least 1 salinity unit (2 units; Figs. 6, 9, 11). Moreover, the increased freshwater in the central Arctic is also caused by the retention of low salinity water at the center of the Beaufort Gyre. Due to the less saline water in the upper layers, balancing the freshwater budget requires an increased salinity in the deeper layers. This implies stronger stratification between the upper and subsurface waters. To show the change in ocean stratification, we computed the density difference—which has a direct relation with stratification—across 50 m (Fig. 12). It can be seen that the density difference, and thus the stratification, increases by more than 0.2 kg m^{–3} in most regions when BS is closed.

As proposed by Carmack et al. (2015), there is a narrow contiguous pan-Arctic riverine coastal domain where water flows counterclockwise around Northern Eurasia and northern North America and is normally interrupted in the western Arctic by the BS inflow. With BS closed, there is a continuous coastal pathway for the Siberian water to flow westward until it encounters the Mackenzie River discharge, which results in significant reduction of salinity (more than – 1.5 salinity unit) in the upper water layer of the western Arctic Ocean, notably in the Chukchi Sea and along the Alaskan slope (Fig. 6b, c). Without Pacific Water flowing into the Canada Basin by the various pathways through the Chukchi Sea (e.g. Gong and Pickart 2015), salinity increases in the southern Canada Basin along the Canadian and CAA coasts due to circulation changes. The increase in the surface inflow from the Atlantic Ocean which reaches the CAA by mixing and circulation also contributes to the mentioned salinity increase. Thus, there appears an important gradient of surface salinity from the Chukchi Sea to the Canada basin (Fig. 6b, c). Salinity also increases in the Laptev and Kara Seas with a closed BS due to more Atlantic Water entering into the Arctic.

Another consequence of closing the BS concerns the temperature and heat content of the Arctic Ocean. We find the highest temperature (0.5°) and heat content ($0.5\text{--}1.5 \times 10^9 \text{ J}$) of the top 50 m in the experiments with BS closed (Fig. 8). This may seem counterintuitive given that the Pacific Water flowing into the Arctic Ocean through BS is a source of heat as it is warmer than the Arctic surface water above the

halocline. Hence this means that closing the BS results in circulation changes with a reorganization and an increase of heat storage in the upper layers of the Arctic. The increase in the heat content can also be associated with the upward shift of the relatively warm waters lying between 50 and 120 m (not shown), since the Pacific water beneath the mixed layer is no longer present. This in turn allows more heat to

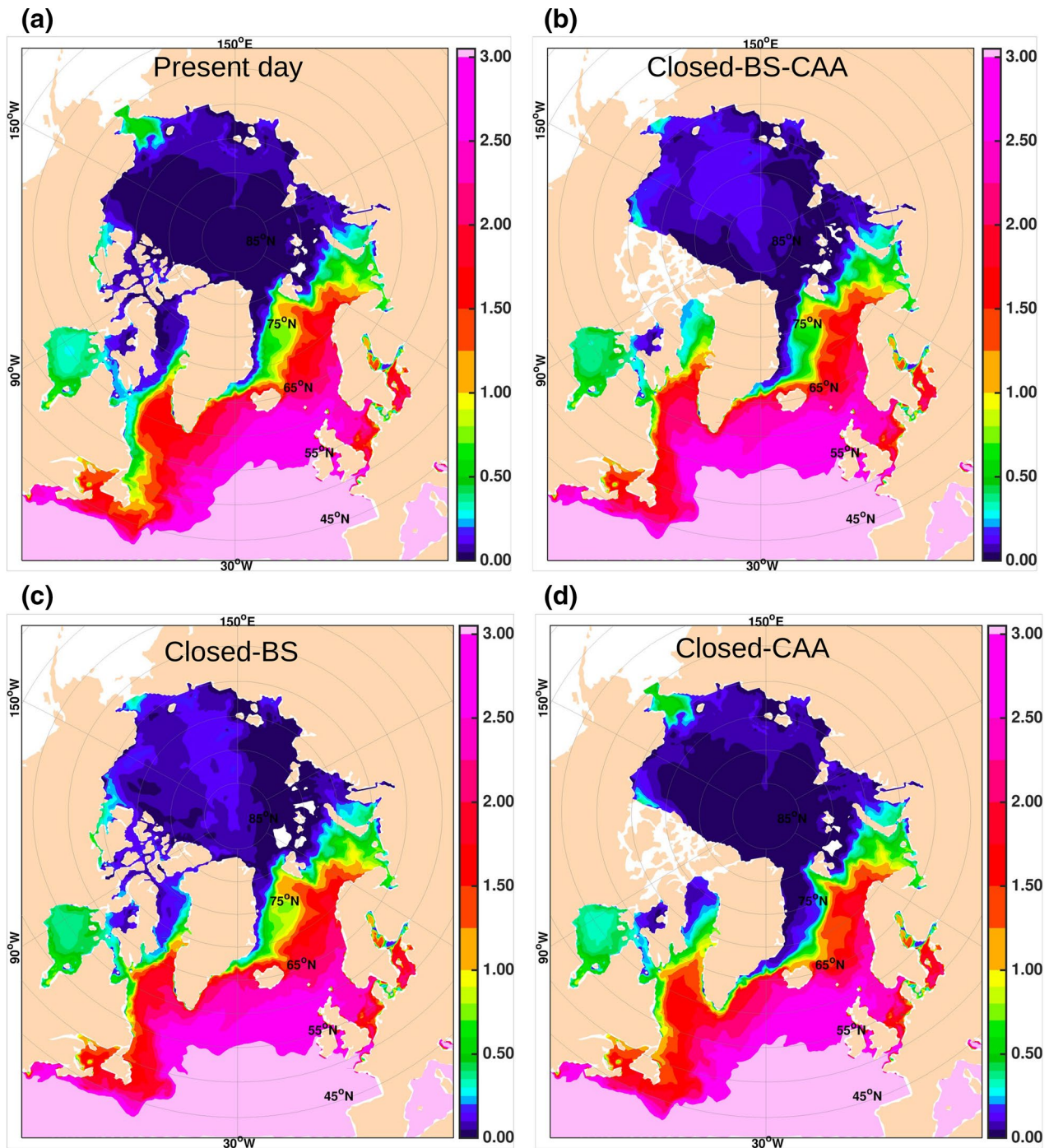


Fig. 8 Annually averaged model top 50 m heat content in 10^9 J , over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Note that the color-bar is non-linearly discrete and there are more contours between 0 and 0.5 with 0.05 contour interval

be transported to the surface layers, but further experiments and analysis are needed to investigate this. Consistent with this, sea ice thickness decreases throughout the Arctic, being on average 0.5 m thinner than in those experiments with the BS open (Fig. 3a, d). An exception is the Chukchi Sea where very thick sea ice (Fig. 3b, c) is formed due to the absence of heat flux from the Pacific Ocean.

Closing the CAA overall has a minor impact on the Arctic circulation, temperature and sea ice, and the Beaufort Gyre and the pathways of Eurasian runoff do not change unlike the BS-closed case. We see, however, that salinity in the top 50 m reduces in most places in the Arctic and particularly along the CAA coasts (reduction reaching 1 salinity unit) after closing the CAA. Overall, we conclude that the impact of BS is more dominant than CAA over the Arctic. In our sensitivity experiments, sea ice, temperature and salinity in the Barents Sea show only minor changes even with closed BS, which suggests other factors such as atmospheric forcing controls water properties and sea ice in this basin. It should be mentioned that in our results above, we did not explore the possible impact of sea ice change on the momentum transfer and the circulation, but that could be particularly important for the Beaufort Gyre dynamics (e.g. Dewey et al. 2018; Meneghello et al. 2018a, b; Zhong et al. 2018).

3.3 Impact of BS and CAA on the GIN Seas and North Atlantic

Closing the BS with an open CAA slightly increases (decreases) temperature and salinity in the Greenland Sea (Norwegian Sea) probably due to the re-distribution of the Atlantic water in these basins (Figs. 5, 6c, 10, 11). It also decreases the sea ice in the Greenland Sea because of the reduced sea ice export amounting to $1760 \text{ km}^3 \text{ year}^{-1}$ (Fig. 3c). Closed BS also leads to a shallower mixed layer depth in the GIN Seas (Fig. 7c). By closing the CAA, however, temperature and salinity decrease significantly (being larger than -1.5 unit in some regions) in most part of the GIN Seas, as there is an enhanced export of cold and fresh water with the East Greenland Current through Fram Strait (Figs. 5, 6d). For both CAA and BS being closed, the salinity and temperature anomalies are similar to the closed-CAA case but with smaller anomalies suggesting that the CAA has a stronger impact than BS in the GIN Seas (Figs. 5, 6b). In our experiments, there is more sea ice in the GIN Seas if the CAA is closed, especially if BS is still open, as Fram Strait drives all the export for the Arctic Ocean (Fig. 3d). In the case of closed CAA but open BS, the sea ice export through Fram Strait is larger by 50% ($3570 \text{ km}^3 \text{ year}^{-1}$) which contributes to the thicker sea ice in the GIN Seas. The extra freshwater and sea ice entering the GIN Seas when the CAA is closed leads to shallower mixed-layer depths in these basins (Fig. 7d).

Baffin Bay is warmer and more saline and has thinner sea ice when either of the BS or the CAA and particularly both are closed (Figs. 5, 6). It can be seen that Baffin Bay has the lowest heat content and highest freshwater content in the present-day configuration, with all gateways open (Figs. 8, 9). In the Labrador Current, less sea ice is carried south if either of the CAA or BS is closed (Fig. 2), but there is a little change in sea ice thickness (given it is mainly first-year ice formed in Baffin Bay or locally along the Labrador shelf). The upper ocean temperature and salinity in the Labrador Sea increase when BS and/or CAA is closed (Figs. 5, 6). This is partly related to the consequent warming and salinization of Baffin Bay water which feeds the Labrador current and propagates further south along the Labrador shelf to the Grand Banks. The increased temperature and salinity in the Labrador Sea is stronger when BS is closed, likely related to more saline and warm Atlantic water entering the Labrador Sea. Nevertheless, the deepest mixed layers (Fig. 7) occur when BS is open. Closing BS leads to reduced mixed layer depth especially when the CAA is also closed, but it shifts the region with the deepest mixed layer northward. In the Subpolar Gyre, the barotropic transport increases by forty percent when BS is closed.

3.4 Impact of BS and CAA on transport into and out of the Arctic

The volume transports through Arctic gateways are controlled by the conservation of volume and dynamical factors (e.g., circulation) in the Arctic. Closing BS while the CAA remains open only causes small changes in the volume transports at the other Arctic Gateways (Table 2) as the circulation re-organizes to respond to the loss input through BS. The inflow to the Arctic Ocean slightly increases at the BSO, while there is more outflow through the Fram Strait and less outflow via Lancaster Sound and Davis Strait. Given the fact that we use the same atmospheric forcing in all experiments, this is consistent with the island rule of Joyce and Proshutinsky (2007) whereby the climatological flow drives cyclonic circulation around Greenland. Closing the CAA, however, significantly enhances the southward flow through Fram Strait (3.9 Sv) and leads to zero net flow through Davis Strait. Closing the CAA outflow implies that the outflow from the Arctic Ocean through Fram Strait balances the inflow through Bering Strait and the BSO. Without the Pacific Water inflow at BS, the central Arctic freshwater content decreases as shown before (Fig. 9). Thus, the freshwater export is lower, particularly at the CAA. Closing the CAA while leaving BS open means that all the outflow has to occur through Fram Strait, switching that gateway to a strong net southward flux, with a significantly enhanced liquid freshwater flux (amounting to 111 mSv). Closing both BS and the CAA significantly changes the transports.

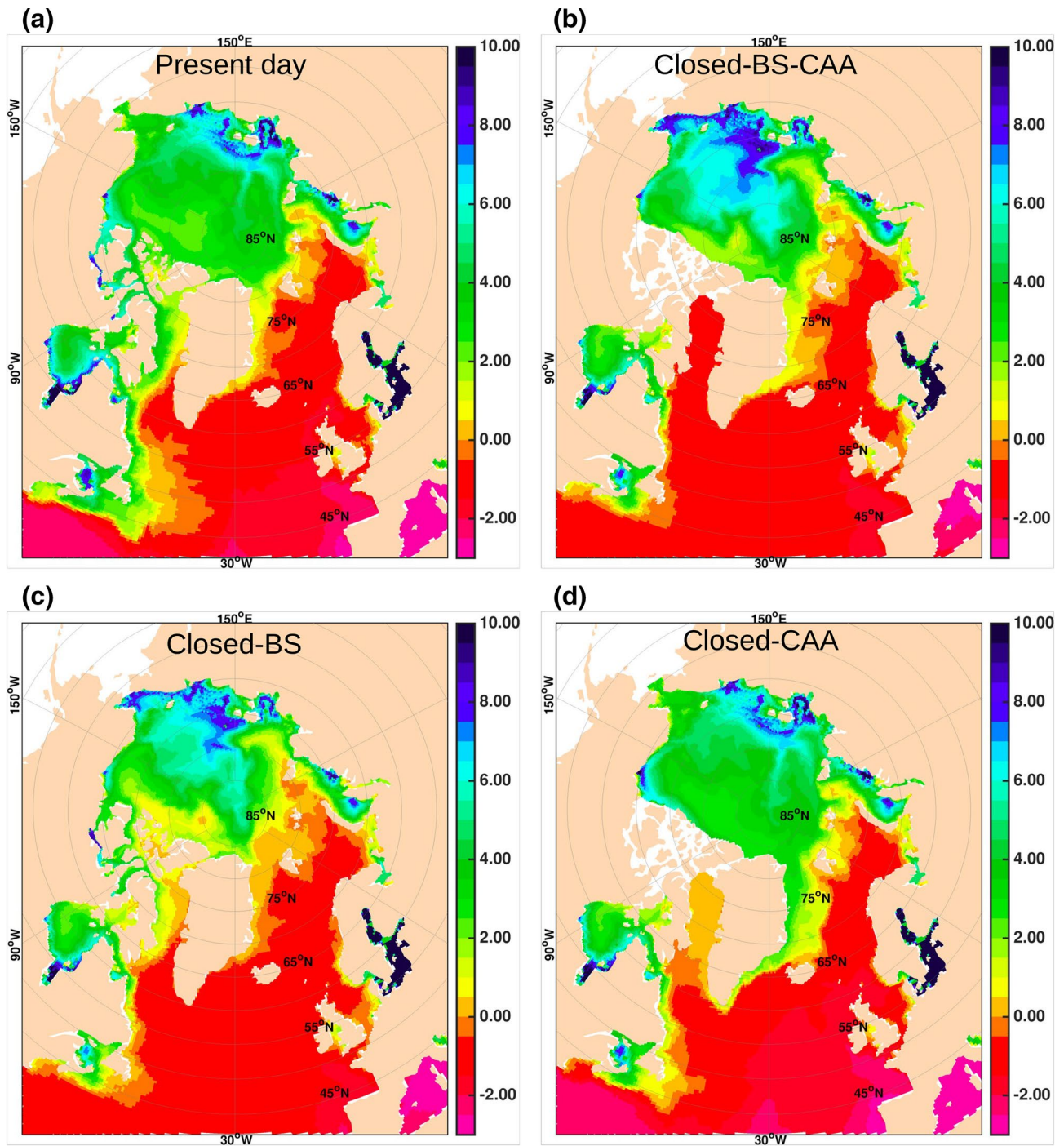


Fig. 9 Annually averaged model top 50 m freshwater content, relative to salinity of 34.8 in m, over years 51–60 of integration, for **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Contour interval is 0.25

Having the only open gateways east of Greenland, a simple loop type circulation develops, with inflow through the BSO and Fram Strait recirculating back out with the East Greenland Current through Fram Strait. Even with this gateway configuration, there is little change in the volume, freshwater and heat fluxes through the BSO suggesting that these are

set by other external factors, rather than the overall Arctic volume, freshwater and heat budget. At Davis Strait, the freshwater export decreases by more than seventy percent when BS and/or CAA are closed. This partially plays a role in increasing the salinity in the western North Atlantic, and in particular along the shelves, as shown before.

4 Discussion

The volume, freshwater and heat fluxes of the Arctic Ocean determine the thermodynamic and dynamic characteristics of the Arctic Ocean and play roles in sea-ice and water properties of the Arctic. While the Pacific waters entering the Arctic Ocean through the BS are a net source of heat and freshwater, the export of low salinity and cold waters from the Arctic to the subpolar North Atlantic through the CAA and the Fram Strait impacts the salt and heat content of the North Atlantic Ocean. By performing a series of sensitivity experiments with the NEMO ocean model, our study investigates the impact of the BS and/or CAA closure on the Arctic and North Atlantic Oceans, and the consequences of flux changes through BS and CAA which is of relevance for the paleo, present-day and future Arctic. The approach developed here could be tested by simulations covering the Holocene interval, which was marked by large amplitude regional climate variations, notably in the Arctic and subarctic areas. It should also be used for constraining the Arctic component in the projections of future climate that need to take into consideration the freshwater and heat budget of the Arctic. We discuss more implications of our study in the following sections.

4.1 Implications for the present-day and future Arctic

Yang (2005) shows that changes in volume fluxes through the gateways can change the circulation in the Arctic, and this relation is linear to a certain extent. In our results, we find that the strength of circulation is subject to changes after closing BS and/or CAA while the pattern of circulation remains unchanged. This also suggests a linear relation between the gateway fluxes and the circulation. Given the impact of circulation on the water and sea ice properties of the Arctic, we can assume there is a linear or at least a quasi-linear relation between the gateway fluxes and their impacts on the Arctic Ocean. This implies that we can extrapolate our results, regarding closing BS/CAA, to any decrease or increase of the fluxes through the BS and CAA gateways. For instance, we find that the Beaufort Gyre weakens after closing BS, which is consistent with Zhong et al. (2015) and Spall (2020) that also find the Beaufort Gyre weakens in response to the decreased freshwater flux through BS. Contrariwise, increasing BS inflow should result in spinning up the Beaufort Gyre. Giles et al. (2012) finds that the Beaufort Gyre has spun-up since 2001 but contrary to what they suggest wind stress cannot be the responsible forcing as it is not correlated with the gyre strength after 2001 (cf. Figure 2 in Giles et al. 2012).

We, therefore, suggest that strengthening of the Beaufort Gyre is likely to be related to the increased flow through BS. Observations indeed confirm that the BS inflow has increased by 0.4 Sv between 2001 and 2011 (Woodgate et al. 2012). Such an increase in BS inflow has a significant impact on the Arctic subsurface waters (Aksenov et al. 2016) consistent with our results.

It is important to understand how the Arctic outflow is partitioned between the CAA and Fram Strait because of their distinct role in controlling the AMOC. Comparing all our simulations, during both spin up and cyclo-stationarity states, we find that the volume fluxes through the CAA and Fram Strait are anticorrelated (cf. de Boer et al. 2018). This behavior is probably linked to the circulation changes in the Arctic, and should be considered for the studies of the future Arctic. It could be used to explain the differences between the models in simulating the Arctic circulation and the partitioning of its outflow between the Canadian straits and Fram Strait (cf. Yang 2005; Aksenov et al. 2016). Our results can also be useful in understanding the discrepancies between different model simulations and inter-model comparison projects such as EU H2020 PRIMAVERA for the Arctic (Fuentes-Franco and Koenigk 2019). For instance, understanding how much the differences in BS and CAA fluxes between the models, play a role in their discrepancies for the Arctic and Atlantic freshwater contents.

4.2 Implications for the paleo-Arctic Ocean

Although we did not aim to model the paleoceanography of the Arctic during LGM in this study, the results provide information that can be compared with the proxy data. The interval considered here mostly covers from the Last Glacial stage, including the early Holocene, which was a period of transition marked by large differences in sea-surface temperatures between the western and eastern North Atlantic (e.g. de Vernal and Hillaire-Marcel 2006) and the very late establishment of optimal postglacial conditions in Baffin Bay (e.g. Gibb et al. 2015). This period was marked by major changes in relative sea levels due to ocean volume variations (ice sheet growth and decay) and isostatic adjustments. At the time scale of the last deglaciation, however, the retreat of the continental ice sheets has been accompanied by ice streams and very large freshwater discharges to the Arctic and North Atlantic oceans (e.g. Strokes and Clark 2001; Dyke 2004; Hughes et al. 2016), which modified considerably the hydrography and ocean circulation (e.g. Clark et al. 2001; Tarasov and Peltier 2005). Hence, the transition from the last glacial stage to the present interglacial is complex with uncertainties concerning the meltwaters fluxes and their routing in addition to the timing of the opening of the main Arctic gateways. In the present study, our sensitivity experiments

provide scenarios of hydrographical changes that could be compared with paleoceanographic reconstructions from proxy records. We do not pretend that the scenarios are realistic, especially for the glacial stage and deglaciation as there is no continental ice meltwater parametrization in our model. However, the results can help to understand and weigh the role of the Arctic gateways in the ocean reorganization during the transition from glacial to interglacial and to identify strengths and weaknesses of the modelling results and paleoclimate data.

From 8 to 4 ka, data in the northern Baffin Bay suggest an increase in sea ice coverage while those from the eastern Greenland show the opposite (Levac et al. 2001; Solignac et al. 2006; de Vernal et al. 2013). These trends were associated with the general cooling of the mid-late Holocene in the northern Baffin Bay and to decrease meltwater discharges along the eastern Greenland margins. According to our results, the opening of the CAA with increasing Arctic water flow to Baffin Bay relative to Fram Strait could have had a similar impact and contribute to the opposite trends reconstructed from paleodata west and east of the Greenland margins.

Closing the BS led to the higher heat content in the top 50 m and thinner Arctic sea ice. This is supported to some extent by proxy evidence indicating warmer than present conditions in subsurface waters of the western Arctic Ocean during the early Holocene when the Bering Strait was still very shallow or closed (Hillaire-Marcel et al. 2004). Similarly, reduced summer sea ice cover in the southeastern Arctic Ocean during the middle Holocene before sea level reached its maximum high stand (de Vernal et al. 2020), is compatible with lower than modern flow through BS as could be proposed from our results.

The penetration of freshwater from the Siberian rivers into the western and central Arctic in the absence of strong Pacific Water inflow and significantly reduced salinity in the upper water layer of the western Arctic Ocean suggest that the likely pathway for meltwaters from Eurasian glaciers during deglaciation would have been to the west until BS opens. If correct, a way to trace back in time the opening of BS might be the use of a diminution of geochemical tracers from Siberian sources in sediment cores from the Alaskan/Beaufort shelf and slope. During the last glacial-interglacial transition, however, BS opened well after the retreat of most Eurasian ice caps (Hughes et al. 2016). Nevertheless, the main transition occurred from 11 ka with initial opening of the BS (Jakobsson et al. 2017) until the end of the middle Holocene (Manley 2002; Graham et al. 2016). This interval of sea level transition is also marked by changes in the mineralogical content of cores from the Chukchi Sea suggesting maximum fluxes through the BS were reached after 6 ka (e.g. Ortiz et al. 2009; Yamamoto et al. 2017).

Closing the BS affects the freshwater budget of the Arctic Ocean causing stronger stratification between the upper water layer and subsurface waters, which is consistent with proxy data of the last glacial stage (Norgaard-Pedersen et al. 2003). A strong stratification at the origin of the decoupling between epipelagic and mesopelagic proxies in the Chukchi Sea during the early Holocene (de Vernal et al. 2005a, b) could be explained by a weak BS inflow, while sea level was still rising (Manley 2002).

With BS closed, the upper ocean salinity tends to decrease in part of the Canada Basin (Fig. 6) except in the southern Canada Basin (Fig. 6) due to circulation changes. This results in a pronounced gradient of surface salinity from the Chukchi Sea to the Beaufort Sea, which might explain regional discrepancies in proxy data (e.g. Farmer et al. 2011; de Vernal et al. 2013; Stein et al. 2017).

In our experiments, there is more sea ice in the Nordic Seas if the CAA is closed, especially if BS is open, as Fram Strait drives all the export from the Arctic Ocean. Given higher freshwater and sea ice export through Fram Strait, the resulting sea ice in the Nordic Seas along the route of the East Greenland Current is also thicker. Because of significantly high freshwater export, the Nordic Seas also become fresher with the CAA closed. Such a scenario has implications on stratification, convection and deep-water formation in the Nordic Seas.

In Baffin Bay, there is less sea ice when either the CAA or BS is closed, with the lowest concentrations when both are closed. This is partly due to changes in the import of warm Atlantic water, leading to an increase of top 50 m heat content when either strait is closed. This is consistent with the basin being saltier, given the high salinity of Atlantic waters (Curry et al. 2011). Baffin Bay heat content being low when both straits are open, suggests that changing sea level (opening Bering Strait) and ice sheet retreat (opening the CAA) worked together to end early Holocene warmth in this area. Such a scenario, however, only partly applies to the last glacial stage, during which Baffin Bay experienced subsurface advection of Atlantic water (de Vernal et al. 1992) but very extensive sea ice cover (de Vernal et al. 2005a, b; Gibb et al. 2015) likely due to the large continental ice sheets and general colder climate (e.g. Waelbroeck et al. 2009).

5 Conclusion

We used a regional sea ice ocean general circulation model to explore the role of fluxes through the Arctic Ocean gateways on the high latitude oceans. By running simulations with combinations of BS and CAA open and closed, we examined changes in sea ice properties, salinity and temperature, freshwater and heat fluxes, and the mixed layer depth in the Arctic, GIN Seas and North Atlantic Ocean.

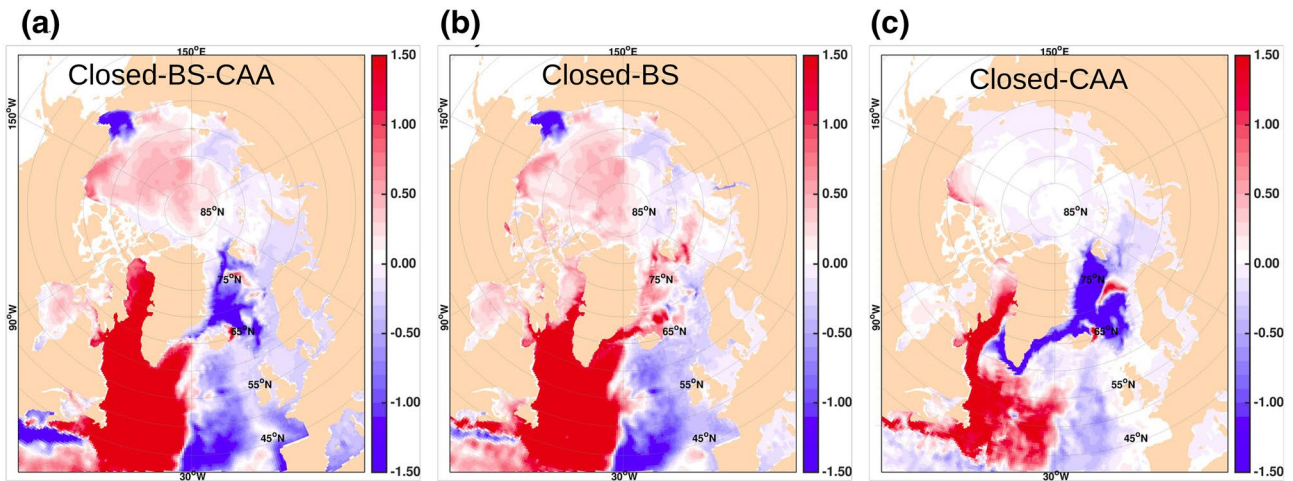


Fig. 10 Temperature difference (annually averaged model top 50 m) with the control (present-day) experiment **a** closed-BS-CAA; **b** closed-BS; and **c** closed-CAA

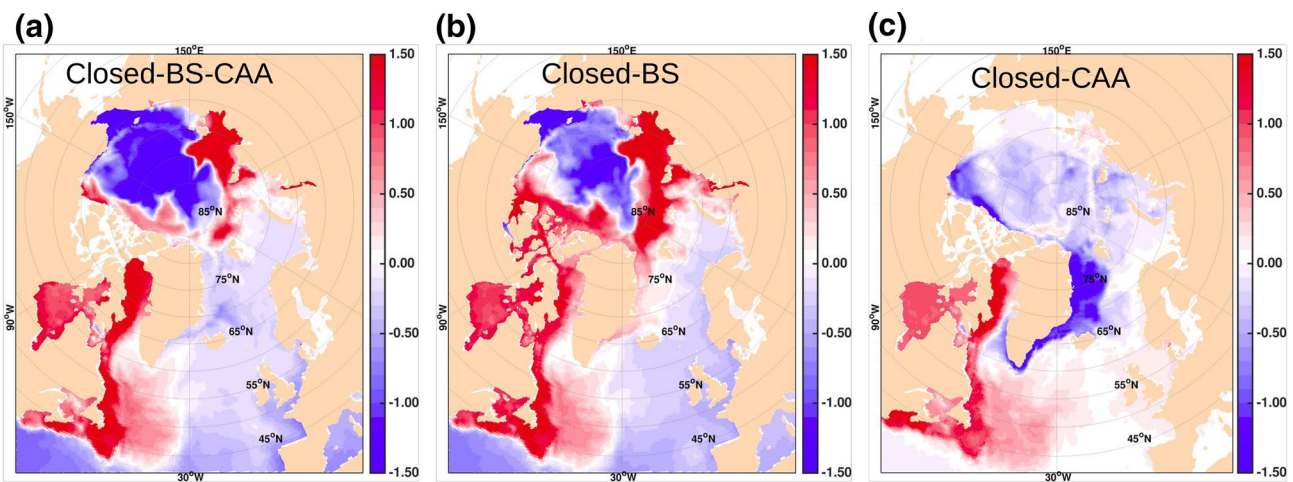


Fig. 11 Salinity difference (annually averaged model top 50 m) with the control (present-day) experiment **a** closed-BS-CAA; **b** closed-BS; and **c** closed-CAA

Closing BS changed the surface circulation in the Arctic and the North Atlantic Oceans and decreased the strength of Beaufort Gyre, which consequently modified the pathways of run-off water in the Arctic Ocean. This in turn resulted in a decrease of salinity in the upper 50 m and an increase of stratification over much of the Arctic. BS closure also counterintuitively increased the heat content in the upper layers of the Arctic leading to a thinner sea ice cover. Closing the CAA, however, had a minor impact on the Arctic Ocean as BS was found to be more dominant. Our results showed that both BS and the CAA influenced the sea ice, temperature and salinity of GIN Seas, with CAA being dominant in this region. The BS and the CAA were found to have opposite impacts in the GIN Seas, but similar impacts in Baffin Bay. With regards to the mixed layer depth, BS was found to have

a larger impact than the CAA on the mixed layer depth in the Labrador Sea, but both BS and CAA affected the mixed layer depth in the GIN Seas. We also presented the implications of our results for the scenarios where the Arctic gateway fluxes had changed in the paleo-Arctic and possibly will change in the future Arctic.

Appendix: Strength of Beaufort Gyre

From a theoretical point of view, the influx of PV through the BS has the largest share in the PV budget of the Arctic as it is a shallow gateway (Yang 2005). Since the wind field is unchanged, the change in the potential vorticity balance is the main factor changing the strength of Beaufort Gyre. The

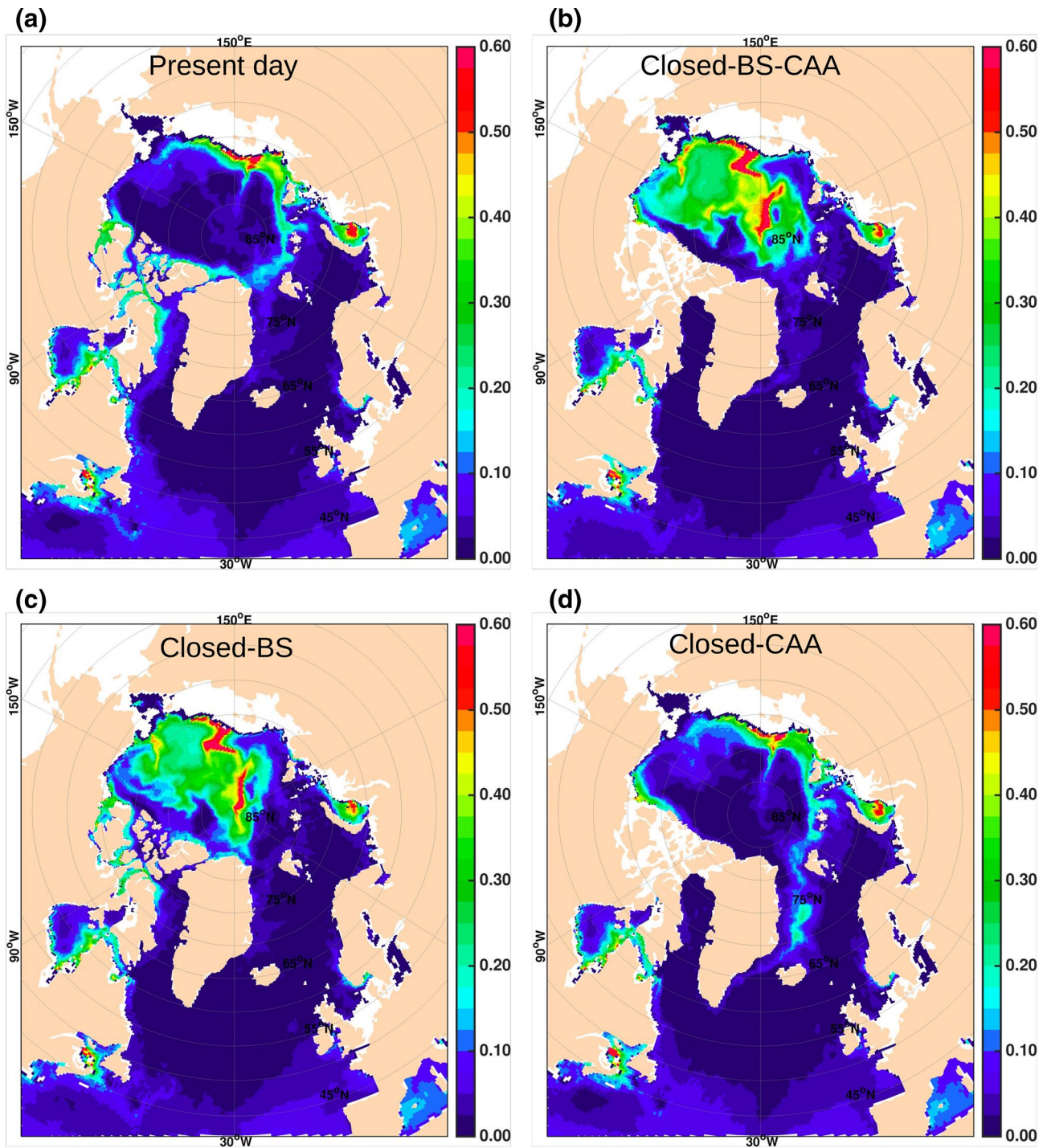


Fig. 12 The density difference across 50 m, in kg m^{-3} , over years 51–60 of integration, to show the upper ocean stratification: **a** present day; **b** closed-BS-CAA; **c** closed-BS; and **d** closed-CAA. Note that

the regions shallower than 50 m (in the Arctic Ocean and Baltic Sea) were masked in white

advection of potential vorticity is a function of torque due to the joint effect of density and topographic gradients (JEBAR effect) and the torques exerted by the wind and bottom stresses. By applying this theory to our model analysis, we find that since those terms for wind and bottom stresses are

kept unchanged in our sensitivity experiments, closing BS and the associated change in the density field is responsible for the change in the advection of PV, and thus the change in the Beaufort Gyre. It should be noted that the resolution of our model (Fig. 1) is slightly coarser than the Rossby radius

of deformation in the Arctic (ranging between 2 and 15 km; Timmermans and Marshall 2020), and therefore, our model is not eddy resolving and might not be fully compatible with the mentioned theory (Figs. 10, 11, 12).

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