

PARAMETERIZING THE RELEASE OF ICEBERG MELTWATER AT DEPTH: IMPACTS ON THE SUBPOLAR NORTH ATLANTIC

by

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Abstract

Around half of the mass loss from the Greenland Ice Sheet (GrIS) is in the form of solid ice discharge, partly represented by the calving of icebergs, which has been increasing in the last few decades. Icebergs release freshwater away from their sources, impacting ocean physical properties and primary production. Despite their significance to the marine environment, their representation in ocean models is still limited and simplified. In ocean models that represent icebergs, the release of meltwater is usually prescribed at the ocean surface. However, observations and laboratory experiments reveal that this is not always the case in the real world. Therefore, in this study we investigate the impacts of representing the release of iceberg meltwater through depth (along their keel) instead of at the ocean surface. We used simulations carried out with the Nucleus for European Modelling of the Ocean, with different numerical representations and depth of intrusion of the iceberg meltwater. Although the project initially employed four simulations, only two are considered here, since the differences between the simulations with different numerical representation of iceberg plumes showed very small differences compared to the ones obtained by varying the depth of meltwater intrusion. Here, we present the main findings in the form of a manuscript intended for submission, followed by a detailed methodology and a section presenting the differences between the initial simulations used in this project. This study demonstrated that parameterizing the intrusion of iceberg meltwater at depth, instead of at the surface, leads to significant changes in the subpolar North Atlantic and in the ocean surrounding Greenland. The intrusion of meltwater at depth decreases the water column stability, enhancing iceberg melt rates and allowing meltwater to spread into deeper layers and basin interiors (Chapter 1, section 3.1). This parameterization leads to colder and saltier upper layers in Baffin Bay, enhanced inflow of warm and salty Atlantic waters in the subpolar gyre, and saltier waters in the Labrador Sea (Chapter 1, section 3.2). The changes in physical properties in the interior of the Labrador Sea results in decreased water column stability and deeper mixed layers, during winter (Chapter 1, section 3.3). Finally, the intrusion of meltwater at depth enhances primary production of the study region (Chapter 1, section 3.4). These significant changes in physical properties in large-scale emphasizes the need for a more accurate parameterization of how the meltwater from icebergs is released in the ocean. The factors to be considered in an improved parameterization are discussed in Chapter 1, section 4.

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Firstly, I would like to express my deepest gratitude to Dr. Juliana Marson, for her invaluable guidance, support, encouragement and patience during the last two years. Dr. Marson was fundamental to the development of the present work and to my growth as an early career researcher. I appreciate not only the great opportunity to learn from her, but also, beyond her academic mentorship, I appreciate the great source of inspiration that she has been for me. I feel incredibly fortunate to have had the opportunity to work with her (I also thank MMM for guiding me toward this opportunity). I would also like to thank Dr. Paul Myers, for his time (despite his super demanding schedule) resources, and great scientific contributions to develop this project. Additionally, I would like to thank Dr. Karen Alley, for her insightful comments and valuable suggestions throughout this process. Additionally, I want to highlight that her support since the beginning of my studies at the University of Manitoba was fundamental to my progress and perseverance. Secondly, I want to thank the PORTAL team which provided amazing insights and comments to improve and push this project forward. Thank you to Paulo, for being the person with whom I shared the daily academic challenges, and my deepest feelings about life and the oceans (as you wrote in a poem, an intense baroclinic friendship). Thank you for Luiz, for all the amazing insights and, especially, for becoming a close friend along the way (since a long long way from Cassino Beach). I would also like to thank all the amazing friends and coworkers who I spent the last two years with. Finally, I would like to thank the Amaral Wasielesky family for the unwavering love, encouragement, and support. I am especially grateful to my dad who encouraged me to study throughout my life and who has given me the strength and courage to reach this point. I am deeply grateful to my sister, Ana, who has been an invaluable source of inspiration and intellectual support throughout my studies. Especially, for being my person and my best friend, and always being there for me through every challenge. Lastly, I want to express my deepest gratitude to my mom, whose love and encouragement were a constant source of inspiration and strength throughout my life.

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1. Paper to be Submitted

This chapter presents the central part of this thesis, which is a manuscript currently being prepared for submission to the Journal of Geophysical Research: Oceans.

Parameterizing the Release of Iceberg Meltwater at Depth: Impacts on the subpolar North Atlantic

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Key Points:

- We use a numerical model to assess the differences of representing the intrusion of iceberg meltwater at depth or at the surface.
- The release of meltwater at depth enhances vertical mixing, leading to an increase in temperature and salinity of the upper layers of the subpolar North Atlantic.
- Deep meltwater plumes also lead to the contraction and weakening of the eastern portion of the subpolar gyre and deepening of the Labrador Sea winter mixed layer.

Abstract

The calving of icebergs from the Greenland Ice Sheet has been increasing in the last few decades. Icebergs release freshwater gradually as they drift and melt, which can impact physical properties of the ocean and primary production far from coastal regions. Despite their importance, iceberg representation in ocean models is still limited and simplified. For ocean models that do represent icebergs, their meltwater is usually inserted at the modelled ocean top level – which is not always the case in the real world. Therefore, in this study we investigate the impacts of representing the release of iceberg meltwater along iceberg keels, instead of at the ocean surface, using two simulations carried out with the Nucleus for European Modelling of the Ocean. The release of iceberg meltwater at depth enhances vertical mixing, especially in regions densely populated by icebergs. This increased mixing leads to enhanced melt rates, dispersion of meltwater into deeper layers, nutrient upwelling, and entrainment of warmer and saltier waters into the upper layers of the subpolar North Atlantic. In Baffin Bay, the enhanced mixing leads to colder upper layers, which are linked to an acceleration of its cyclonic circulation and enhanced Arctic outflow. The subpolar gyre is strengthened in its western portion, while weakened and contracted in its eastern portion, leading to an enhanced inflow of subtropical Atlantic waters into the gyre. Finally, in the Labrador Sea the water column stability is decreased, resulting in deeper mixed layers during winter.

Plain Language Summary

The number of icebergs that break off from Greenland glaciers has been increasing in the last decades. When icebergs melt, they may release freshwater at the ocean surface or deeper down along their keels, which might reach hundreds of metres below the surface. Since freshwater is lighter than the ambient salty water, it tends to stay at the surface. If released at depth, the freshwater rises, mixing with its surroundings, altering the ocean's physical properties and ecosystem. Ocean models usually represent the release of iceberg meltwater only at the ocean surface; here we use an ocean model to understand the impacts of releasing the meltwater from icebergs along their keel, instead. Results show that releasing meltwater at depth can impact the ocean circulation in Baffin Bay, linked to its increased salinity and cooling, while in the subpolar North Atlantic the changes in circulation result in higher salinity and warming. Additionally, the release of meltwater in depth enhanced primary production on the Labrador Sea during spring.

1 Introduction

The Greenland Ice Sheet (GrIS) is losing mass at increasing rates (Mankoff et al., 2021; Mouginit et al., 2019; Simonsen et al., 2021; Otosaka et al., 2022). Around half of the GrIS mass loss is in the form of solid ice discharge, which is represented partly by the calving of icebergs (Bamber et al., 2018; Van den Broeke et al., 2016). This solid ice discharge has already risen from 450 km³/year to 500 km³/year, from the 1990s to recent years (Mankoff et al., 2020). While a significant portion (10% to 50%) of the icebergs' mass melts before reaching

open ocean regions (Davison et al., 2022; Enderlin et al., 2016), the remaining icebergs release freshwater in the open ocean, away from their sources (Martin and Adcroft, 2010). The released freshwater alters the water column properties (i.e. salinity and temperature) and, consequently, can impact marine ecosystems and ocean dynamics (e.g., Cenedese and Straneo, 2023).

Iceberg melt plumes affect primary production through nutrient enrichment of the upper ocean layers, either by releasing nutrients contained within the ice (Duprat et al., 2016; Wu and Hou, 2017) or by promoting vertical mixing and nutrient upwelling (Smith et al., 2013; Vernet et al., 2012). In terms of ocean dynamics, the freshwater fluxes from icebergs impact ocean stratification and mixed layer depth (e.g., Merino et al., 2016). Observations show that the meltwater input from icebergs can either increase the water column stability or, conversely, promote mixing of the waters surrounding icebergs (Abib et al., 2024; Kajanto et al., 2023). As described by Gade et al. (1993), when the iceberg melt plume is released at the ocean surface, the water column stability is increased, since the undiluted plume has lower density than seawater. In contrast, when the plume is injected at depth, the water column stability is decreased as the buoyant plume rises, promoting mixing of the water surrounding icebergs. Using measurements from underwater gliders, Lucas et al. (2025) observed effects from both processes (i.e., increased and decreased water column stability) around a single tabular iceberg in the Southern Ocean. The authors reported that the surface release of meltwater forms a shallow layer of cold and fresh water, while the side and bottom melting of icebergs increase vertical mixing under this layer, promoting the upwelling of warm, nutrient-rich waters in the surroundings of the iceberg.

The behavior of an iceberg melt plume – whether it enters the ocean through the surface or along its keel – was investigated by FitzMaurice et al. (2017). Through laboratory experiments, the authors showed that in non-stratified, homogeneous flows, this behavior is controlled by the background velocity. Iceberg melt plumes can remain attached to the iceberg when their velocity is higher than the relative velocity of the background flow. In this case, the plumes attached to the sidewalls of the icebergs ascend without being diluted, spreading on the ocean surface. In contrast, iceberg melt plumes can be detached from the iceberg when their velocity is smaller than the relative velocity of the background flow. In this case, the meltwater readily mixes with the ambient water and enters the ocean along the iceberg keel.

Despite their importance to ocean dynamics and marine ecosystems, the representation of icebergs and their related processes (e.g. calving, drifting, and melting) is not ubiquitous among ocean and climate models. The modelling of icebergs became more popular with a study on iceberg drift in the Arctic and North Atlantic by Bigg et al. (1997). In their model, the icebergs deteriorate due to basal melt, side melt, wave erosion, solar heat, sensible heat transfer from the air, and sublimation. Afterwards, Gladstone et al. (2001) applied this model to the Southern Ocean and added the influence of ocean temperature in the wave erosion

term, as well as the effect of sea ice on surface waves. Martin and Adcroft (2010) coupled a modified version of the Gladstone et al. (2001) iceberg model to a global circulation model, where wave erosion generates bergy bits, which travel with the parent iceberg and melt due to side and basal melt. Subsequently, Marsh et al. (2015) introduced the Martin and Adcroft (2010) model to the Nucleus for European Modelling of the Ocean (NEMO), using only ocean surface fields to force icebergs. Merino et al. (2016) and Marson et al. (2018) improved the NEMO iceberg module by including iceberg interaction with the vertically integrated profiles of ocean velocity and temperature. Finally, Marson et al. (2024) implemented a parameterization to represent the locking of icebergs in sea ice. Nonetheless, in all the above-mentioned studies, the iceberg meltwater was solely introduced at the ocean surface for simplicity. The implementation of “detached” plumes in ocean models has yet to be explored.

In this context, here we use NEMO to carry out two simulations that represent the freshwater input from icebergs at the ocean surface (i.e. first ocean level) or through depth, distributed along the iceberg’s keel. We analysed the differences between the simulations to assess the key changes in ocean dynamics and primary production in the subpolar North Atlantic, since this region receives the majority of Greenland’s solid ice discharge (Marson et al., 2018). We hypothesize that the release of iceberg meltwater at depth, instead of at the surface, could enhance vertical mixing due to the rising buoyant plumes that entrain the surrounding waters. Therefore, the simulations’ outputs will be used to assess the impact of the depth of iceberg melt plume intrusion on (1) the water column stability and meltwater distribution in large-scale; (2) physical properties and dynamics of the ocean; (3) the stability of the water column and mixed layer depth of the Labrador Sea during winter; and (4) nutrient distribution and primary production.

2 Model and Experiments

2.1 Experimental Configurations

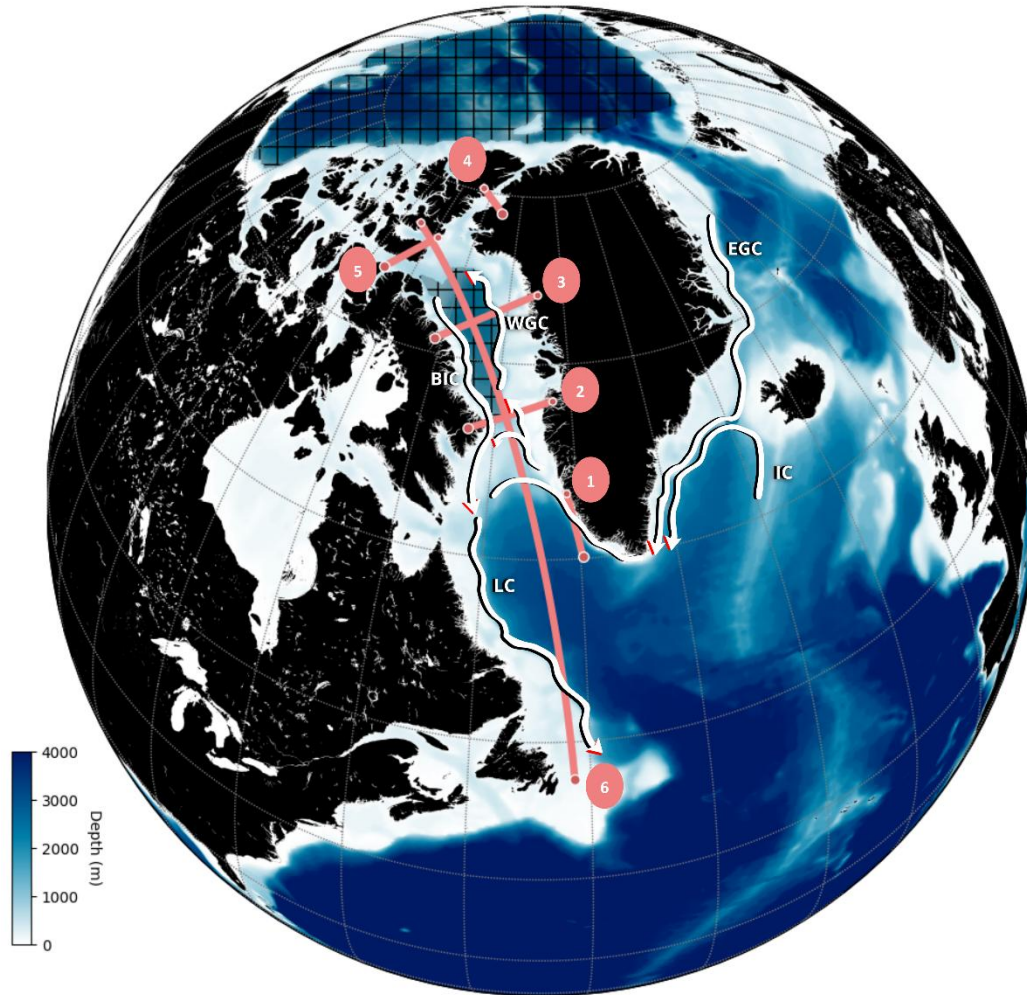


Figure 1: Map showing the main currents of Baffin Bay and Subpolar North Atlantic: EGC (East Greenland Current), IC (Irminger Current), WGC (West Greenland Current), BIC (Baffin Island Current), LC (Labrador Current). Salmon pink lines show the transects across: Southwestern Greenland Shelf (1), Davis Strait (2), Middle Baffin Bay (3), Nares Strait (4), Jones/Lancaster Sound (5), and between northern Baffin Bay and Grand Banks (6). Hatched areas were used for defining the interior of the Arctic Ocean and Baffin Bay.

The simulations used in this study were carried out with NEMO v3.6 (Madec and the NEMO team, 2008), with its ocean component coupled with the Louvain-la-Neuve Sea Ice Model (LIM2; Fichefet and Maqueda, 1997). The ocean is also coupled with the simplified model of biogeochemical cycling called Biogeochemistry with Light Iron Nutrients and Gasses (BLING; Galbraith et al., 2010). The domain covers the Arctic and Northern Hemisphere Atlantic (ANHA¹; Courtois et al., 2017; Gillard et al., 2022; Marson et al., 2018; Marson et al., 2021; Marson et al., 2024; Myers et al., 2024; Pennelly et al., 2019; Pennelly and Myers, 2021), with open boundaries at the Bering Strait and at 20°S in the Atlantic Ocean. The simulations have a horizontal resolution

¹ <https://canadian-nemo-ocean-modelling-forum-community-of-practice.readthedocs.io/en/latest/Institutions/UofA/Configurations/ANHA4/index.html>

of 1/4° (ANHA4), and 50 vertical levels. The GLObal Ocean ReanalYsis and Simulations (GLORYS2v3; Masina et al., 2017) was used for boundary and initial conditions. The Canadian Meteorological Center's global deterministic prediction system reforecasts (CGRF; Smith et al., 2014) was used for atmospheric forcing, at a spatial resolution of 33 km, with hourly fields from 2002 to 2017 (the period in which the simulations were run). The dataset from Bamber et al. (2018) is used for Greenland's liquid and solid discharge rates, which vary monthly and interannually. The dataset used for river runoff forcing is Dai and Trenberth's Global River Flow and Continental Discharge Dataset (Dai et al., 2009; Dai and Trenberth, 2002).

2.2 NEMO-ICB and the iceberg melt plume parameterizations

In the iceberg module NEMO-ICB, described by Marsh et al. (2015), iceberg particles are released from fixed source points along Greenland's coast, following the calving rates proposed by Bamber et al. (2018). Those Lagrangian particles contain a group of icebergs with a particular size from the ten classes defined in Martin and Adcroft (2010), which interact with the 3D temperature and velocity fields occupied by the iceberg keel, as in Merino et al. (2016) and Marson et al. (2018). The particles' momentum is governed by the pressure gradient force, the Coriolis parameter, the wave radiation force, and ocean, atmosphere and sea ice drag forces. Iceberg deterioration occurs due to basal melt, side melt, and wave erosion. The parameterization of basal melt is described by Equation (1):

$$M_b = 0.58 |\vec{v} - \vec{v}_b|^{0.8} \frac{\tilde{T}_b - \tilde{T}}{L^{0.2}} \quad (1)$$

where $\vec{v}$ is the iceberg velocity (m/s), $\vec{v}_b$ is the ocean velocity at the base of the iceberg (m/s), $\tilde{T}$ is the iceberg temperature (°C, fixed at -4°C), $\tilde{T}_b$ is the ocean temperature at the base of the iceberg (°C), and L is the horizontal length of the iceberg, for icebergs with a draft smaller than the bathymetry depth. For icebergs with a draft larger than the bathymetry depth, $\vec{v}_b$ and $\tilde{T}_b$ are represented by ocean velocity and temperature nearest to the sea floor. The side melt parameterization is described by Equation (2):

$$M_v = \int_{-H_{icb}}^0 (7.62 \times 10^{-3} \tilde{T}_o(h) + 1.29 \times 10^{-3} \tilde{T}_o^2(h)) dh \quad (2)$$

where H_{icb} is the depth reached by the iceberg's keel (m) and $\tilde{T}_o$ is the ocean temperature averaged along the iceberg's keel. For icebergs with a draft larger than the local depth, $\tilde{T}_o$ is represented by the average temperature of the ocean between the surface and the sea floor. The wave erosion parameterization is described by Equation (3):

$$M_e = \frac{1}{12} S_s (1 + \cos[\pi A_i^3]) (\tilde{T}_o + 2) \quad (3)$$

where A_i is the sea ice area (m²) and S_s is the sea state, described by Equation (4):

$$S_s = \frac{3}{2} |\vec{v}_a - \vec{v}_o|^{\frac{1}{2}} + \frac{1}{10} |\vec{v}_a - \vec{v}_o| \quad (4)$$

where $\vec{v}_a$ and $\vec{v}_o$ are the velocities of the atmosphere and at the ocean surface, respectively.

The deterioration mechanisms result in changes in the iceberg's dimensions, as well as freshwater and latent heat fluxes between the iceberg and the ocean. The resulting melting term due to the above-mentioned mechanisms is inserted into the modelled ocean in two different forms as follows. The total melt ($M_v + M_b + M_e$) can be inserted at the surface (simulation **VS**), in the first ocean level, or distributed evenly through depth (simulation **VD**), from surface to the iceberg's draft, following the parameterization from Mathiot et al. (2017), to distribute vertically meltwater from ice shelves. Both simulations represent the iceberg melt plume as a volume flux of freshwater at the freezing point by adjusting the velocity divergence field.

2.3 Output analyses

The first two years of the simulations (2002 and 2003) were excluded from the analysis, since in this initial period the model is still adjusting to the atmospheric forcing. For fields that are dependent on the number of icebergs present in the ocean (such as temperature, salinity, velocity, chlorophyll, and phosphate), the analyses also excluded the years 2004 to 2006, since during this period the iceberg field is not stabilized. This happens because (1) there are no icebergs in the domain at the start of the simulations; and (2) the model needs to accumulate enough ice mass in each one of the ten size "bins" to generate a particle, and those bins are empty at the beginning of the model integration.

To investigate how the depth of intrusion of iceberg meltwater impacts vertical stratification, we calculated convective resistance (CR), a metric described by Equation (5):

$$CR(h) = \frac{g}{A \rho_0} \int \int_A \left[h \sigma_\theta(h) - \int_0^h \sigma_\theta(z) dz \right] dA \quad (5)$$

where g is the acceleration due to gravity (9.81 m.s^{-2}), A and z are the area and depth, respectively, of the grid cell being analyzed, ρ_0 is the reference density (1026 kg.m^{-3} ; Madec, 2008), σ_θ is the potential density, and h is the reference depth. Physically speaking, in Equation (5), the integrated buoyancy of the modelled water column, with potential density $\sigma_\theta(z)$ at each depth level, is subtracted from the integrated buoyancy of an idealized well-mixed water column, with potential density $\sigma_\theta(h)$. Therefore, CR is the amount of energy that must be removed in order to make the water column neutrally stratified (mixed) down to the reference depth (Pennelly and Myers, 2021). In other words, higher values of CR indicate that the water column has a tendency to resist convection, implying that the water column is stratified (Bailey et al., 2005; Gillard et al., 2022). Convective resistance was calculated to a reference depth of 250m, since this is the maximum iceberg draft in the model. In addition, to better understand the distribution of icebergs throughout the domain, and its link with the changes in water column stability, iceberg occurrence was estimated by counting

the number of unique iceberg particles that passed through each grid point throughout the model integration.

The melting of icebergs can induce turbulent mixing that affects depths below the iceberg's draft (Lucas et al., 2025). Therefore, to capture the effects of the iceberg meltwater injection on key dynamical processes, the study focused on the upper 541 m of the water column. The final tracer fields were integrated over the upper (0 - 55 m), subsurface (55 – 541 m) and deeper (below 541 m) layers. Salinity and temperature were averaged in the upper and subsurface layers. These layers were chosen since the differences between the simulations of those variables show different patterns between the surface and 55 m of depth, and below 55 m of depth (see Supplementary Information). Barotropic streamfunction was integrated from the surface to 541 m.

Additionally, given its importance as a key region of deep-water formation during winter, we want to assess the impacts of the depth of intrusion of iceberg meltwater on the Labrador Sea winter stratification. For that, we calculated the winter average CR to a reference depth of 2000 m, since mixed layer depth (MLD) in this area can reach such values during strong convection events (Piunno et al., 2025; Yashayaev and Loder, 2017). For this study, the MLD was estimated by the method developed by Courtois et al. (2017), which applied the method from Holte and Talley (2009) for Argo floats to numerical models. This method estimates the MLD based on density, temperature and salinity vertical profiles. For each variable, candidate MLDs are estimated, and the final MLD is selected as the depth that shows the greatest consistency among temperature-, density-, and salinity-based estimates.

Finally, to better understand the impacts of the iceberg melt parameterization in nutrient distribution and primary production we used chlorophyll and phosphate from BLING. The outputs from BLING cover the period from 2002 to 2016, thus, the period from 2007 to 2016 was used. Chlorophyll was integrated from the surface to the bottom. Phosphate was integrated down to 55 m of depth, since this is the approximate Labrador Sea euphotic zone (Cota et al., 2003; Harrison and Li, 2007). These variables were averaged over spring, since in the region of study seasonal spring blooms occur as a result of increased levels of solar radiation and enhanced nutrient supply due to the deep winter mixing (Pinet, 2016).

3 Results and Discussion

3.1 Impacts on the water column stability and meltwater distribution

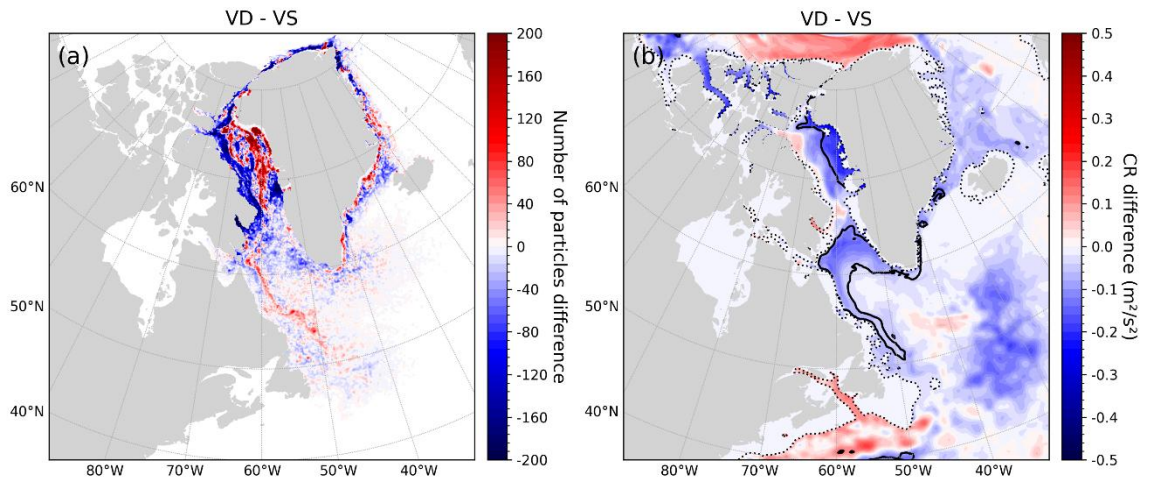


Figure 2: (a) Differences (VD-VS) in the total number of iceberg particles that passed through each grid point from 2002 to 2017. (b) Differences (VD-VS) of mean (2007-2017) convective resistance to a reference depth of 250 m. Black dotted contours show the 250 m isobath, while black solid contours indicate regions where the differences are statistically significant (p -value < 0.01).

Figure 2(a) shows the differences between the simulations in the number of particles that crossed each grid point throughout the period of integration. The differences in iceberg distribution show that VD has a higher occurrence of icebergs in eastern and central Baffin Bay, as well as along the Labrador Current. In those regions, the upper water column stability significantly changes between the simulations. This is displayed in the differences of CR between the simulations (Figure 2(b)), which measures the amount of energy necessary to mix the water column down to a reference depth. Figure 2(b) shows that VD has lower CR in most of the domain, with differences reaching 20-25 $\text{cm}^2.\text{s}^{-2}$ in the eastern portion of Baffin Bay, where typical CR values to a reference depth of 250 m range between 50 and 75 $\text{cm}^2.\text{s}^{-2}$. In other regions where iceberg melt is prominent, such as the Greenland southeast coast, northern Labrador Sea, and central Baffin Bay, the differences between the simulations exceed 8 $\text{cm}^2.\text{s}^{-2}$. Since typical CR values in the Labrador Sea are roughly between 10 and 30 $\text{cm}^2.\text{s}^{-2}$, this difference is significant in this region. The lower convective resistance for VD in the regions where a considerable amount of meltwater is released (see accumulated melt mass (kg) in Figure S1(a)) implies that parameterizing the freshwater input from icebergs at depth results in less stratification of the water column.

As a result of the enhanced mixing produced by the intrusion of meltwater at depth in VD the melt rates are altered. Figure 3 displays a time series of the spatially averaged iceberg melt rates in the simulations, as well as the differences between them. During summer, average melt rates range between 6.10^{-5} to $9.10^{-5} \text{ kg.m}^{-2}\text{s}^{-1}$ for VD, while they range from 5.10^{-5} to $7.10^{-5} \text{ kg.m}^{-2}\text{s}^{-1}$ for VS. The differences between the simulations

(VD - VS) have mostly positive values, especially during summer, when the melt rates reach their maximum. During this season, the differences between the simulations range from $0.8 \cdot 10^{-5}$ to $2.2 \cdot 10^{-5} \text{ kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}$. This indicates that prescribing the meltwater intrusion at depth leads to an enhancement of the iceberg melt rates by 16%, on average. The increased melt rates for VD occur since the release of freshwater from icebergs at depth results in a buoyant plume, leading to the entrainment of warmer ambient waters that surround the icebergs, consequently increasing the melt rates (McCutchan and Johnson, 2022; McCutchan et al., 2024).

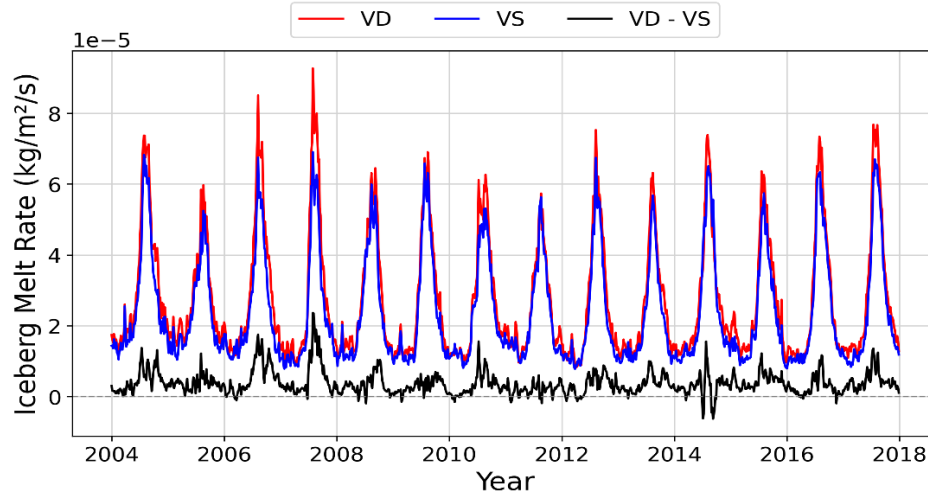


Figure 3: Time series of spatially averaged iceberg melt rates ($\text{kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}$) between 2004 and 2017 for VS (blue) and VD (red). Black line shows the differences between the simulations (VD - VS).

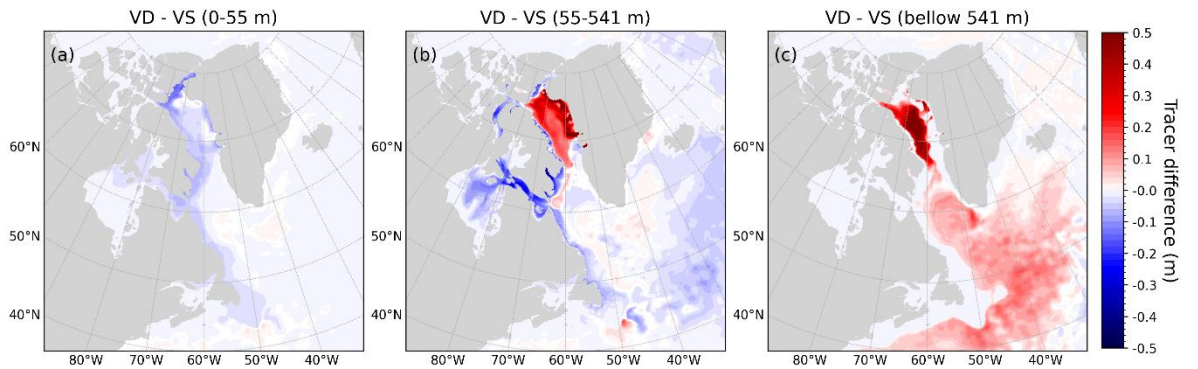


Figure 4: Differences (VD - VS) of the final distribution (December 31st, 2017) of integrated iceberg melt tracer, in the (a) upper (0-55m), (b) subsurface (55-541m) and (c) deeper (below 541m) layers. Regions with p -value > 0.01 were masked out.

Although VD has higher iceberg melt rates, the amount of meltwater released in VD and VS throughout the simulations have the same magnitude. By the end of the period of integration, both simulations released approximately $6.5 \cdot 10^{15} \text{ kg}$ of meltwater (see time series of spatially accumulated melt mass (kg) in Figure S2). However, as a result of the iceberg melt being released at different depths, the large-scale distribution of meltwater is different in the simulations. Figure 4 shows the differences in the final distribution of the melt tracer between the simulations, integrated for the upper (0-55m), subsurface (55-541m) and deeper (below 541m) layers. In the upper layers

(Figure 4(a)), the tracer concentration is larger for VS, which is reasonable since the release of meltwater at the surface leads to its accumulation in the upper layers, while the release of meltwater at depth causes its dilution throughout the layers. The accumulation of tracers in the upper layers for VS is especially pronounced along the west boundary, from Nares Strait to the Labrador shelf, since more icebergs occupy these regions in VS (Figure 2(a)).

For the subsurface layers (Figure 4(b)), VD has a higher concentration of tracers in Baffin Bay. The spread of tracers into greater depths for this simulation suggests that the freshwater introduced at depth readily mixes with its surroundings as opposed to rising undiluted to the surface, as observed by FitzMaurice et al. (2017). VS tends to have higher tracer concentrations at subsurface only in shallower regions (south of Baffin Island, the Hudson Bay complex, and the Labrador shelf) likely as a result of advection by the upper ocean circulation. VD also shows more potential for transporting iceberg meltwater to the interior of the subpolar gyre at all depths. Below 541m, the iceberg meltwater tracer concentration is larger virtually everywhere in VD (Figure 4(c)), indicating that iceberg meltwater released at the surface tends to remain there.

3.2 Impacts to physical properties and dynamics of the ocean

Figure 5(a) shows the difference in average salinity on the upper layers (0-55m) between the simulations. It reveals that the meltwater injection at different depths leads to different patterns of salinity in the upper ocean layers on a large scale. Notably, VD has a higher salinity in most parts of the domain. The larger differences occur on the western coast of Greenland, reaching 0.4 psu. Great salinity differences also occur along other parts of Greenland's coast, in the Canadian Arctic Archipelago (CAA), and in the northern rim of the Labrador Sea, exceeding 0.2 psu. South of the subpolar gyre (SPG), the differences reach 0.1 psu. In contrast, the salinity is lower for VD along the Baffin Island Current (BIC), north of Greenland, and in the vicinity of the Gulf of Saint Lawrence. However, the differences are smaller, ranging between 0.05 and 0.1 psu. The differences in average salinity of the lower layers (55-541m) between the simulations (Figure 5(b)) show a similar pattern to the upper layers, albeit in smaller magnitude. The larger differences occur on the western coast of Greenland, in the northern Labrador Sea and south of the SPG, reaching 0.15 psu.

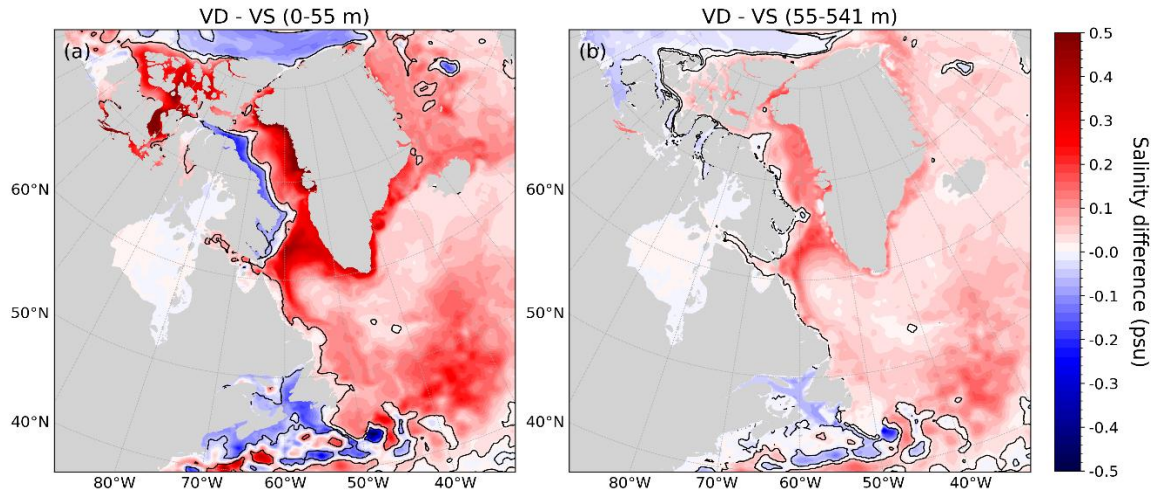


Figure 5: a) Differences (VD - VS) of depth-averaged salinity for the (a) upper (0-55m) and (b) lower (55-541m) layers. Black solid contours indicate regions where the differences are statistically significant (p -value < 0.01).

Similarly to salinity, the depth-averaged temperature in the upper and lower layers also shows large differences between the simulations. In the upper layers (0-55m), the most pronounced temperature differences occur on the southern Greenland coast, in the northern rim of the Labrador Sea and in the SPG (Figure 6(a)), where VD is 0.5°C to 1°C warmer than VS. In the lower layers (Figure 6(b)), the temperatures are also higher for VD on the southern Greenland coast, in the northern Labrador Sea, and especially in the eastern SPG, where differences can exceed 1°C. However, in the western SPG and in Baffin Bay the temperatures are lower for VD (by up to 0.5°C). Those differences in salinity and temperature are potentially caused by different processes happening in different regions, as we discuss next.

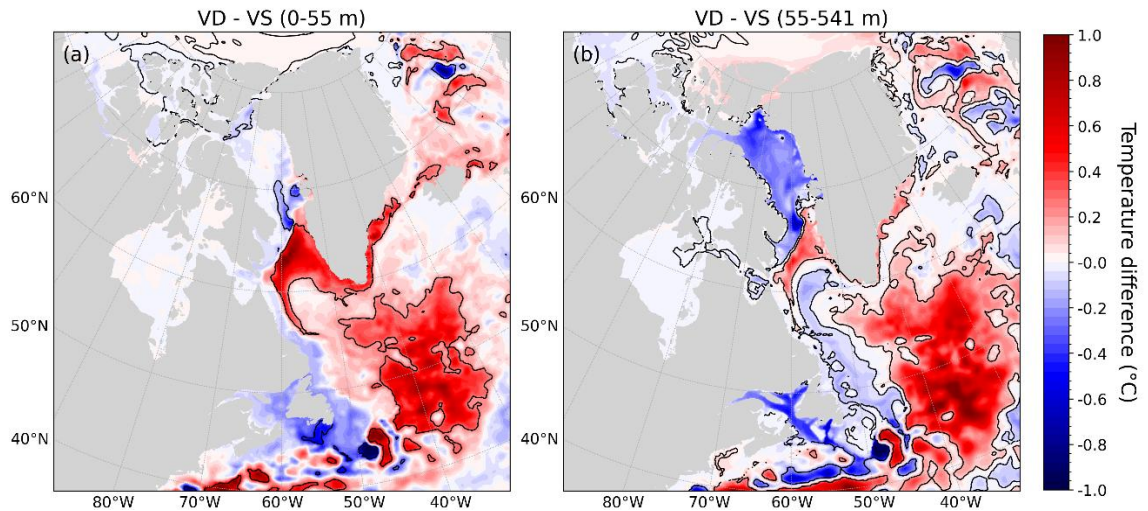


Figure 6: a) Differences (VD - VS) of depth-averaged temperature for the (a) upper (0-55m) and (b) lower (55-541m) layers. Black solid contours indicate regions where the differences are statistically significant (p -value < 0.01).

3.2.1 Baffin Bay

To understand what is happening in Baffin Bay, we start by investigating what is coming in and going out of it through its main gates (see locations in Figure 1) in the

two simulations. At Davis Strait (Figure 7), the water column is characterized by the presence of four main water masses: Arctic Water (AW), West Greenland Irminger Water (WGIW), West Greenland Shelf Water (WGSW), and Baffin Bay Deep Water (BBDW), following the definition formulated by Tang et al. (2004), and further adapted by Curry et al. (2014) and Azetsu-Scott et al. (2012). In the model, we can identify those four water masses. Although their TS values are slightly different from observations, they still occupy the expected endpoints in the TS diagram (Figure 7(c)). The AW is the coldest and freshest water mass that flows southward (above 300 m) along the Baffin Island Coast. The WGIW is the warmest and saltiest water mass flowing northward along the West Greenland shelf break. The waters that flow northward at shallower depths along the West Greenland shelf, namely the WGSW, are slightly colder and fresher than the WGIW, since they are influenced by glacial discharge. And finally, the densest water mass found in this region is the BBDW.

While the temperature difference is not significant for the averaged upper 55 m in Baffin Bay (Figure 6(a)), a warmer and saltier WGSW in VD is consistent with the enhanced mixing produced by the intrusion of meltwater at depth, which would entrain WGIW into upper layers. The “other side of the coin” is that, due to this enhanced mixing with the colder upper layers, the WGIW is visibly colder in VD (Figure 7(a) and 7(c)). This mechanism, generated by the direct impact of meltwater intrusion at depth, likely explains the higher temperature and salinity values along the southeastern Greenland shelf as well.

The AW is slightly fresher and colder in VD. The lower temperatures and salinities along the Baffin Island shelf, as well as the changes in the AW, can be attributed to a larger inflow of waters of Arctic origin into the interior of Baffin Bay in VD, as shown by the time series of southward volume transport across Jones Sound/Lancaster Sound and Nares Strait (Figure 8). The larger inflow of waters from Arctic origin into Baffin Bay in VD can be explained by the larger sea surface height (SSH) gradient between the Arctic Ocean and Baffin Bay in this simulation. The average SSH for the central Arctic Ocean is approximately -39cm in VS and -49cm in VD, while for central Baffin Bay, the average is -48cm for VS and -60cm for VD. Therefore, the SSH difference between the Arctic Ocean and Baffin Bay is ≈ 9 cm in VS, while it increases to ≈ 11 cm in VD. The larger SSH gradient in VD produces a stronger pressure gradient force between those basins, enhancing the inflow of Arctic waters into Baffin Bay.

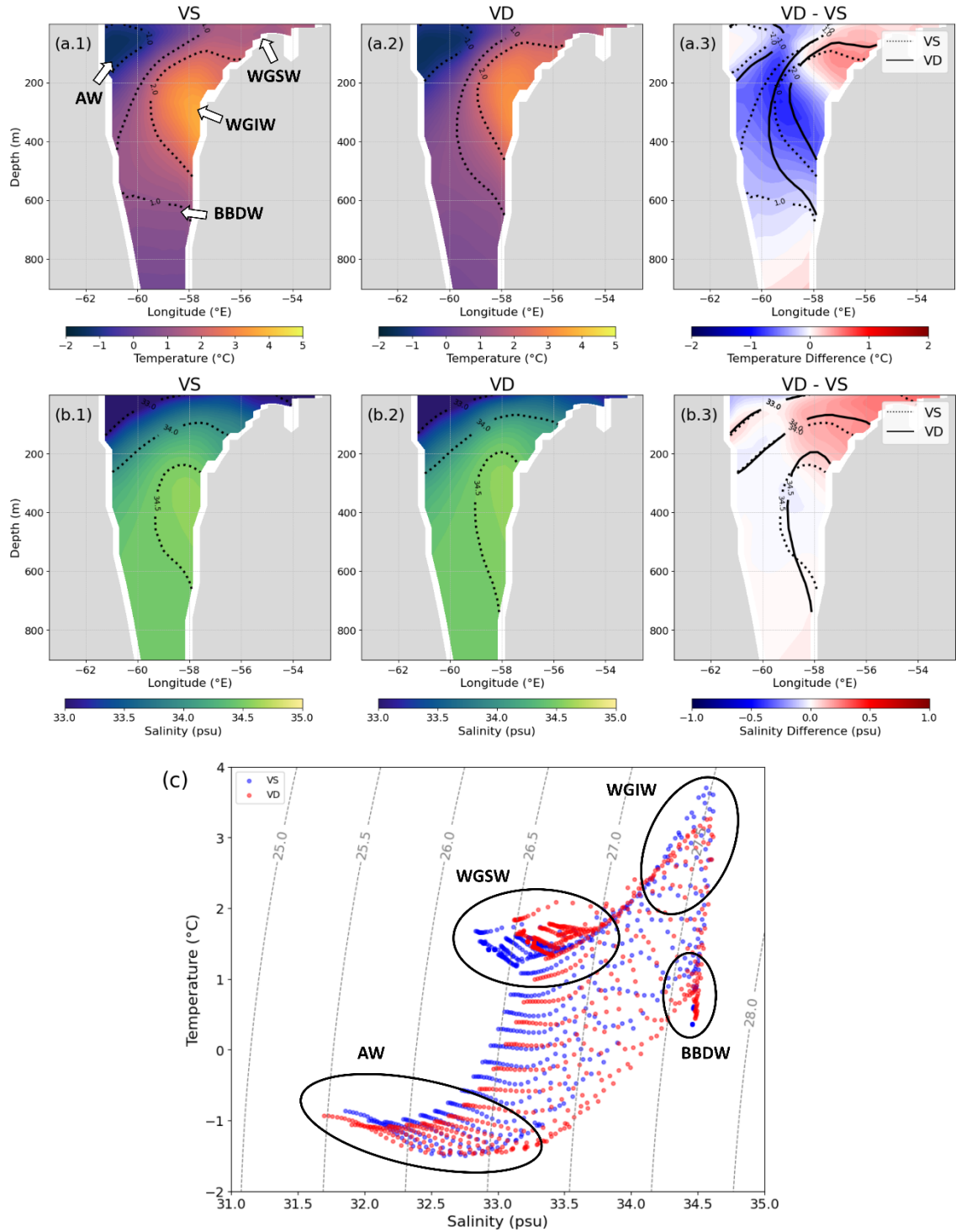


Figure 7: a) Average temperature for VS (a.1) and VD (a.2) on the Davis Strait transect. Black dotted contours show the -1°C, 1°C and 2°C isotherms. Differences of average temperature (VD - VS) are shown in (a.3). Black dotted (solid) contours show the -1°C, 1°C and 2°C isotherms for VS (VD). b) Average salinity for VS (b.1) and VD (b.2) on the Davis Strait transect. Black dotted contours show the 33 psu, 34 psu and 34.5 psu isohalines. Differences of average salinity VS (VD - VS) are shown in (b.3). Black dotted (solid) contours show the 33 psu, 34 psu and 34.5 psu isohalines for VS (VD). (c) T-S diagram showing the main water masses that occur along Davis Strait: AW, WGSW, WGIW, BBDW.

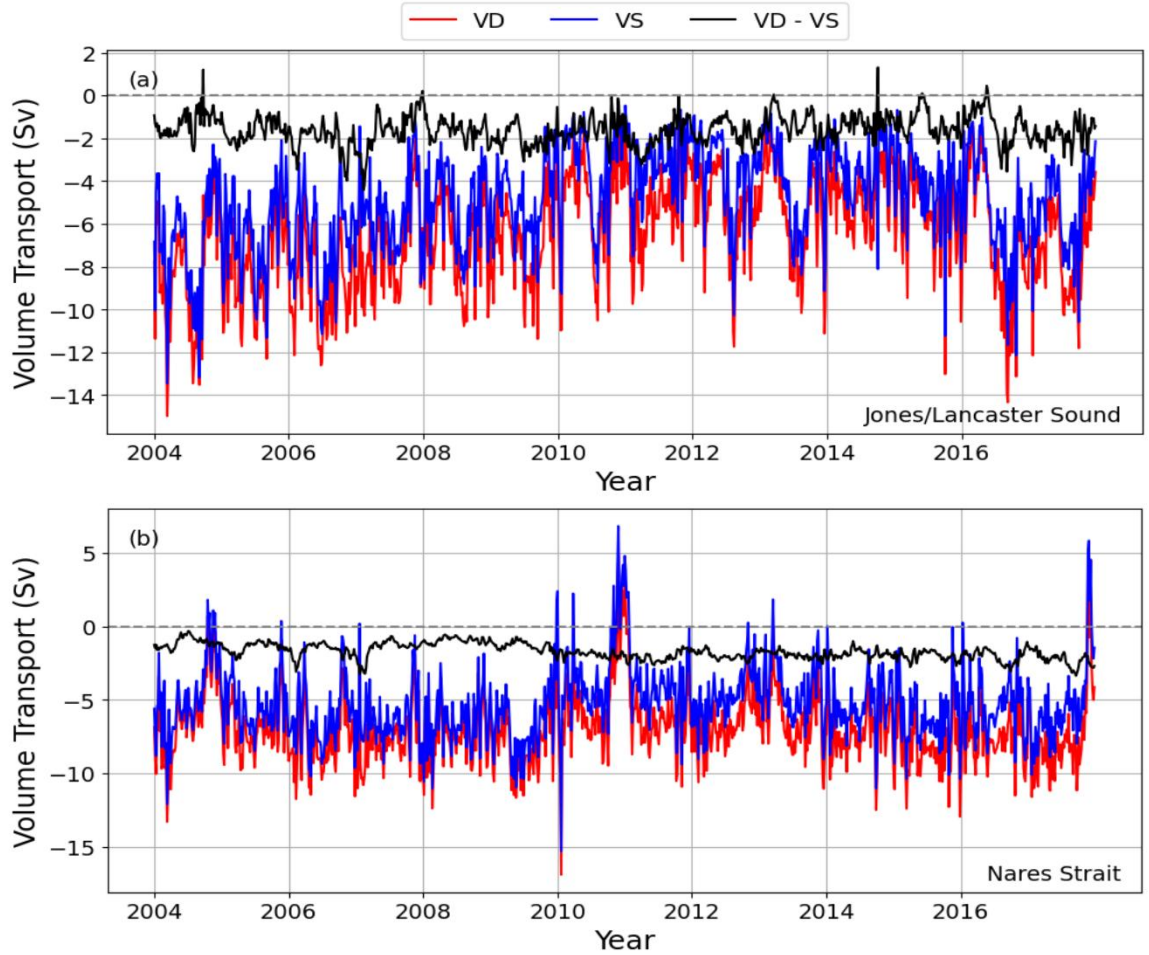


Figure 8: (a) Time series of volume transport (Sv) across Jones Sound and Lancaster Sound, between 2004 and 2017, for VS (blue) and VD (red). Black line shows the differences between the simulations (VD - VS). (b) Time series of volume transport across Nares Strait, between 2004 and 2017, for VS (blue) and VD (red). Black line shows the differences between the simulations (VD - VS).

Differences of SSH also occur in the interior of Baffin Bay due to direct impact of meltwater release. In VS, the release of meltwater at the ocean surface leads to lower density on the upper layers of Baffin Bay, resulting in a higher SSH. In contrast, the release of meltwater in depth in VD, and the associated enhanced mixing, leads to a higher density on the upper ocean layers of Baffin Bay, resulting in a lower SSH. The lower SSH on central-western Baffin Bay in VD produces a slightly steeper zonal SSH gradient in the western portion of the bay (Figure S3). As a consequence of this, as well as a larger inflow of Arctic Water into Baffin Bay, the Baffin Island Current is strengthened (Figure S4). The acceleration of the Baffin Island Current channels icebergs along the Baffin Island shelf break, and prevents their advection to the Baffin Island Coast, explaining the negative differences along the Baffin Island Current in Figure 2(a). Figure 9 shows the barotropic streamfunction for VS and VD in Baffin Bay. A light intensification of the bay's cyclonic circulation in VD is also apparent there. As a result of this intensification, the associated Ekman upwelling in the interior of Baffin Bay is likely increased, which causes uplifting of the isotherms and isohalines (Figure

10(a) and 10(b); Castro de la Guardia et al., 2015), leading to a general decrease in temperature and increase in salinity between 55-541m (Figure 5(b) and 6(b)).

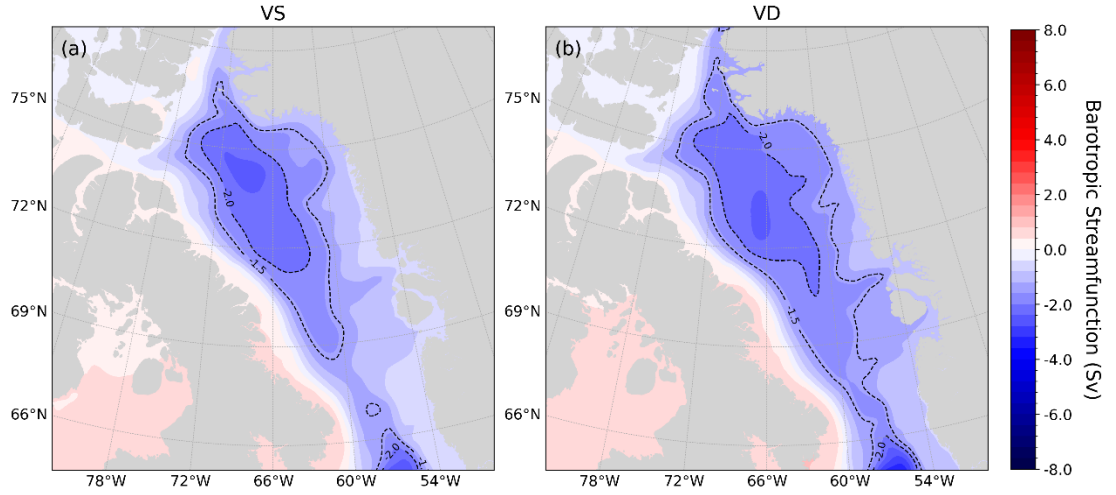


Figure 9: Depth integrated barotropic streamfunction (surface - 541m) for (a) VS and (b) VD, in Baffin Bay. Black dashed contours show the -1.5 Sv and -2 Sv isolines.

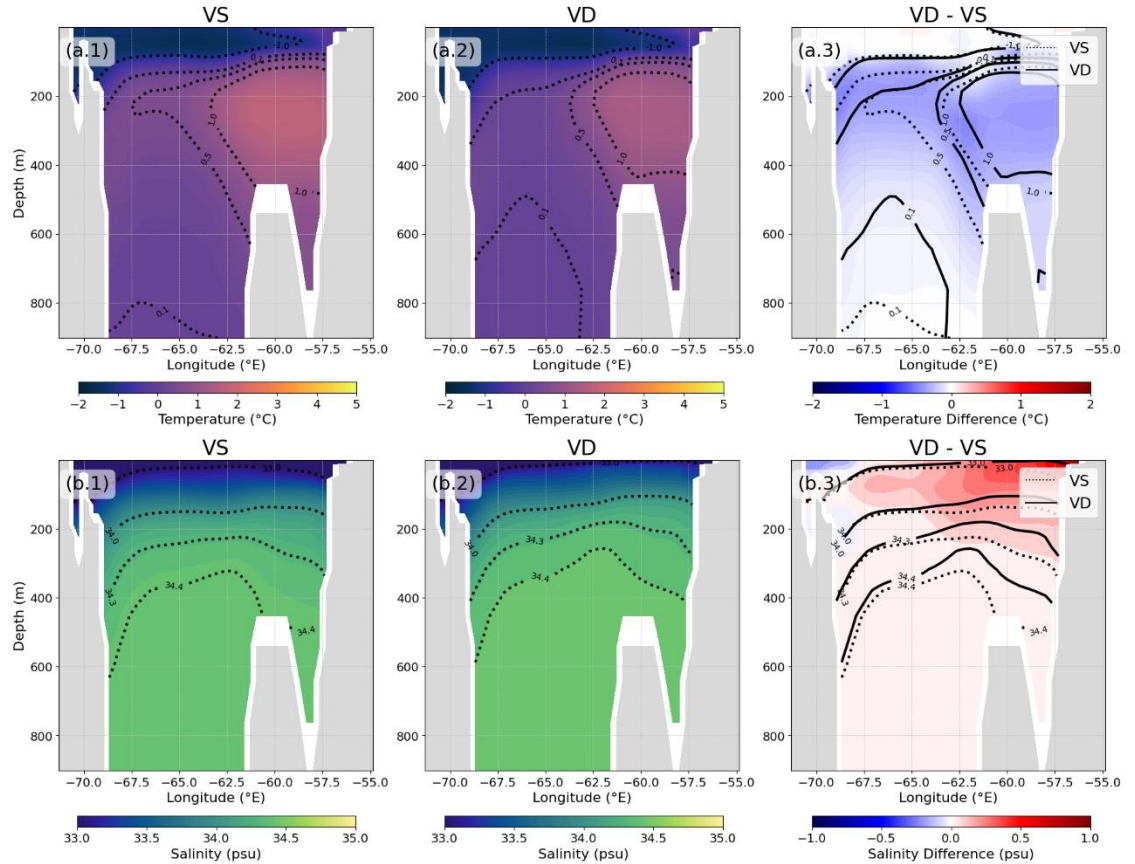


Figure 10: a) Average temperature for VS (a.1) and VD (a.2) on the middle Baffin Bay transect. Black dotted contours show the -1°C , 0.1°C , 0.5°C and 1°C isotherms. Differences of average temperature (VD - VS) are shown in (a.3). Black dotted (solid) contours show the -1°C , 0.1°C , 0.5°C and 1°C isotherms for VS (VD). (b) Average salinity for VS (b.1) and VD (b.2) on the Davis Strait transect. Black dotted contours show the 33 psu, 34 psu, 34.3 psu and 34.4 isohalines. Differences of average salinity (VD - VS) are shown in (b.3). Black dotted (solid) contours show the 33 psu, 34 psu, 34.3 psu and 34.4 isohalines for VS (VD).

3.2.2 Subpolar Gyre

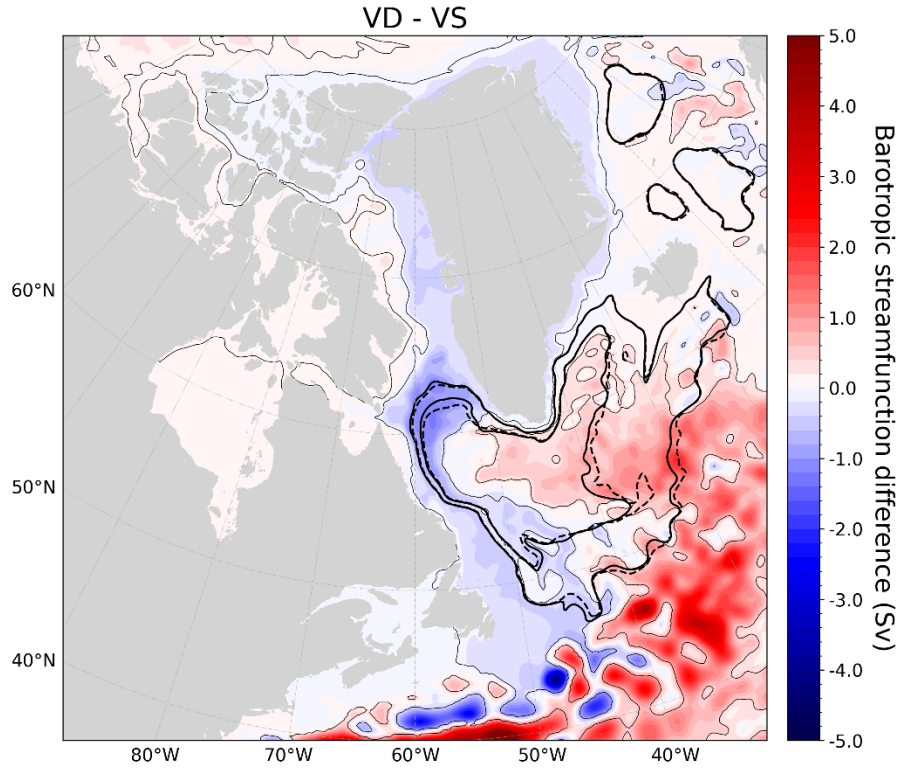


Figure 11: Differences of the depth-integrated barotropic streamfunction (surface - 541m) between VD and VS (VD - VS). Black solid contours show the -5 Sv and -10 Sv isolines in VD, while black dashed contours show the -5 Sv and 10 Sv isolines in VS. Thin black solid contours indicate regions where the differences are statistically significant (p -value < 0.01).

Modifications in physical properties on large-scale in the subpolar North Atlantic are usually associated with changes in the strength and position of the subpolar gyre (Asbjørnsen et al., 2021; Desbruyères et al., 2021; Holliday et al., 2020; Jutras et al., 2023; Robson et al 2016; Tesdal et al., 2018). Therefore, we calculated the differences between the simulations' barotropic streamfunction (integrated from surface to 541m). Given that the subpolar gyre is counterclockwise, the barotropic streamfunction values in its interior are negative. Thus, when looking at the differences between the simulations (Figure 11), negative values indicate strengthening of the SPG, while positive values indicate its weakening for VD. Figure 11 shows that the western SPG is intensified in VD, while its eastern portion is weakened in this simulation, with differences between the simulations exceeding 1 Sv. The intensification of the western SPG is consistent with the enhanced Arctic outflow in VD, as noted previously. On the eastern SPG, the weakening of the gyre is also seen through the displacement between the solid and dashed contours of -5 and -10 Sv, indicating that the subpolar gyre is slightly contracted westward in VD. Weakening and contraction of the SPG has been associated with warming and salinification of the subpolar North Atlantic, due to an enhanced inflow of subtropical Atlantic waters (Desbruyères et al., 2021; Kenigson and Timmermans, 2021; Koul et al., 2020). The shift in the eastern SPG physical properties (i.e., higher temperatures

and salinities in VD), therefore, likely arises from the changing proportion of subpolar and subtropical waters that enter this region. The mechanism linking the introduction of iceberg meltwater at depth and the eastern SPG weakening, however, remains to be determined.

3.3 Impacts in the Labrador Sea during winter

As highlighted in the previous section, the depth of injection of iceberg meltwater impacts the physical properties in the interior of the Labrador Sea. The Labrador Sea is a center of convective activity and the region of formation of the Labrador Sea Water (Clarke and Gascard 1983; Mizoguchi et al., 2003). Thus, the changes in physical properties could impact the winter MLD and potentially impact the volume of LSW produced. Here, we analyze the winter average CR and MLD during winter in the interior of the Labrador Sea. Figure 12(a) shows that, in both simulations, the lower values of CR occur in the western portion of the Labrador Sea, a region where deep convection typically takes place (e.g., Holte et al., 2017). However, in VD, the lower CR values cover a wider region. Indeed, the differences between the simulations (Figure 12(c)) show that VD has overall lower CR values in the Labrador Sea. The greater differences occur in the northern portion and close to the shelf break, where they range between $20\text{cm}^2.\text{s}^{-2}$ and $40\text{cm}^2.\text{s}^{-2}$. The lower convective resistance for VD implies that parameterizing the freshwater input from icebergs at depth results in a more well-mixed water column in the interior of the Labrador Sea during winter.

Figure 12(b) illustrates the winter averaged MLD, showing that the deeper mixed layers occur in the western portion of the Labrador Sea, consistent with the area that have lower CR. In accordance with having a wider region with lower CR, in VD the 400m, 800m and 1200m mixed layer isolines also extend through a much larger area. In addition, the differences between the simulations (Figure 12(d)) demonstrate that VD has a significantly deeper MLD, with differences exceeding 300 m in the northwestern portion of the Labrador Sea. The occurrence of lower values of winter average CR and deeper MLD in VD can be attributed to the differences in water mass properties in the interior of the Labrador Sea. As shown by 5(a) and 5(b), VD has a higher salinity from surface to 541m in the Labrador Sea, particularly in its northern portion, which results in a higher density, with differences exceeding 0.2 kg.m^{-3} . The southwest Greenland shelf is known to export a significant volume of freshwater to the interior of the Labrador Sea (see green arrows in Fig 12(c)) (e.g., Pennelly et al, 2019). Given that VS has fresher Greenland shelf waters compared to VD, the exported shelf waters would decrease the salinity of the northern Labrador Sea in VS, causing its MLD to be shallower. The higher export of fresher waters in VS can be seen in Figure 13, that shows that the westward transport of freshwater from the southwestern Greenland shelf to the Labrador Sea.

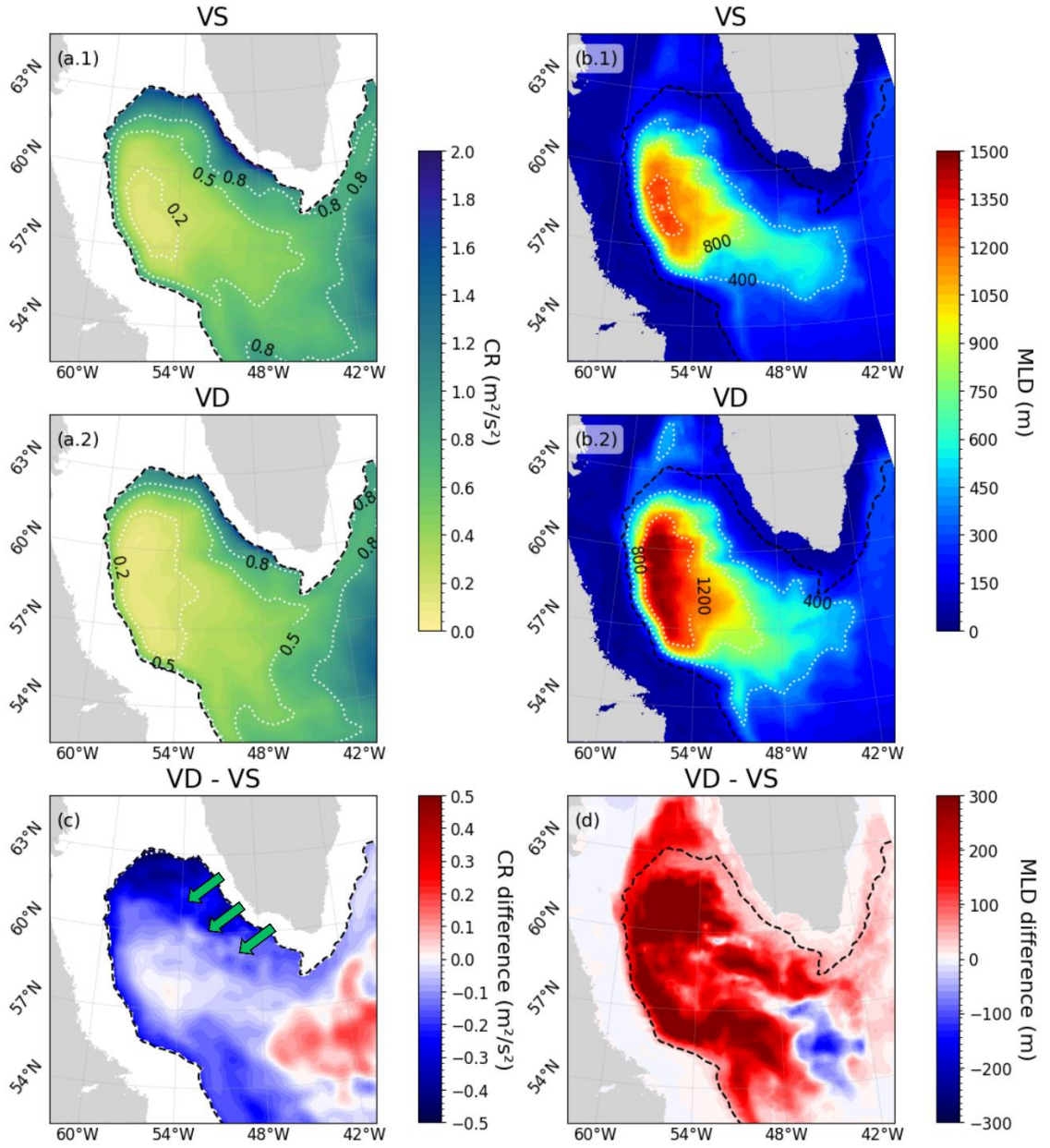


Figure 12: (a) Convective resistance (to a reference depth of 2000m) for VS (a.1), and VD (a.2). White dotted contours show the 0.2, 0.5 and 0.8 m^2/s^2 isolines. (b) Mixed layer depth for VS (b.1), and VD (b.2). White dotted contours show the 400, 800 and 1200m mixed layers. (c) Differences of convective resistance (to a reference depth of 2000m) between the simulations (VD - VS). Green arrows indicate the export of freshwater from the southwestern Greenland shelf to the interior of the Labrador Sea. (d) Differences of mixed layer depth between the simulations (VD - VS). Black dashed contour displays the 2000m isobath.

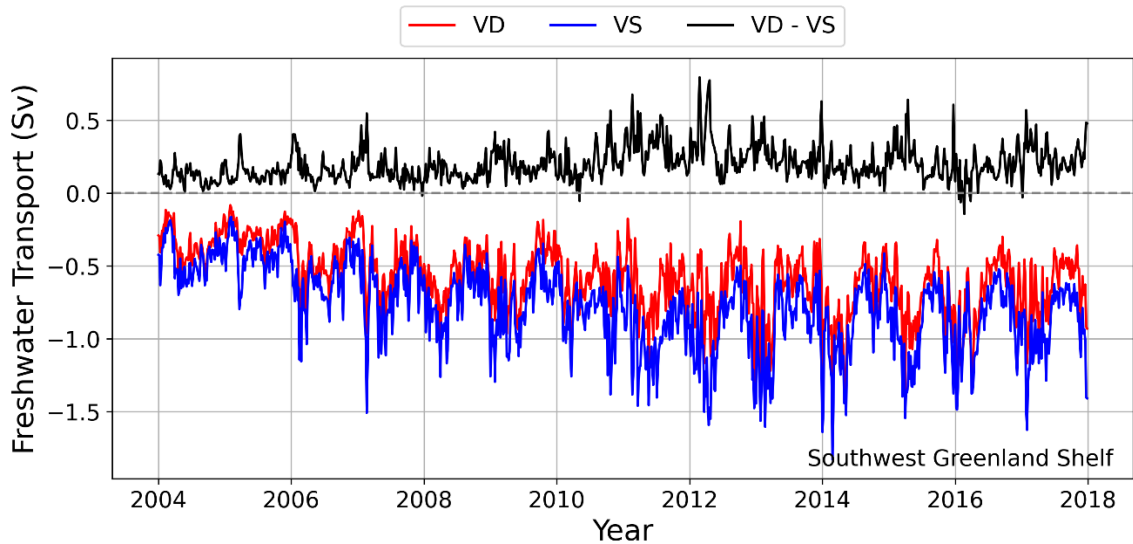


Figure 13: Time series of westward volume transport of freshwater (Sv) (referenced to 34.8) across southwest Greenland shelf, between 2004 and 2017, for VS (blue) and VD (red). Black line shows the differences between the simulations (VD - VS).

3.4 Impacts on Nutrient Distribution and Primary Production

The differences in spring chlorophyll concentration between the simulations in Figure 14(a) show that VD has overall higher values, especially in the SPG and northern Labrador Sea (where p -value < 0.01). This suggests that parameterizing the meltwater input at depth enhances primary production in the Labrador Sea, which can be explained by the elevated concentrations of phosphate (the nutrient “currency” in BLING; Figure 14(b); Galbraith et al., 2010) in VD. The combination of two mechanisms likely produces the difference of phosphate distribution between the simulations: vertical mixing and horizontal advection of nutrients by currents. In VD, vertical mixing is enhanced either directly from the rising of the buoyant plume (especially in eastern Baffin Bay) or indirectly, by generating density anomalies on the upper levels of the ocean that produce deeper mixed layers (such as in the Labrador Sea). Indeed, higher chlorophyll concentrations in the Labrador Sea have been linked to increased vertical nutrient supply due to deeper winter convective mixing (Tesdal et al., 2022). This enhanced mixing brings nutrient-rich waters from the subsurface to the nutrient-depleted photic zone, increasing primary production (e.g., Castro de la Guardia et al., 2019). Additionally, the Labrador Sea nutrient supply is strongly influenced by the advection of cold and nutrient-enriched waters from the Arctic (Glen Harrison et al., 2013). Given that VD has a larger outflow of Arctic waters, more of these phosphate-enriched waters are transported southward by the Baffin Island Current (Figure 14(c)), resulting in higher phosphate concentrations downstream.

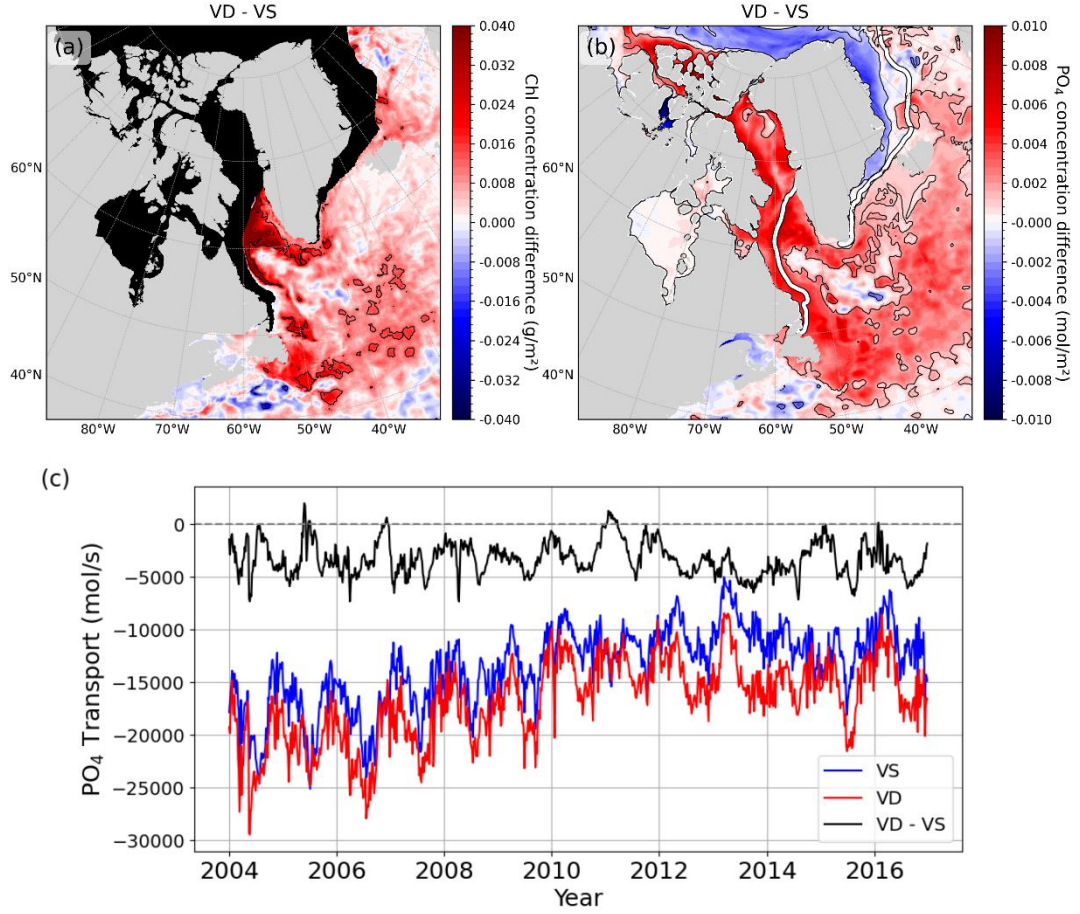


Figure 14: (a) Differences between VD and VS of depth integrated chlorophyll concentration ($\text{g}\cdot\text{m}^{-2}$). Black solid contours indicate regions where the differences are statistically significant ($p\text{-value} < 0.01$). Black shading covers the region where spring average sea ice cover $> 10\%$. (b) Differences between VD and VS of depth integrated phosphate concentration ($\text{mol}\cdot\text{m}^{-2}$) for the upper layers (0-55m). Thick white solid contour shows the limit of the region where sea ice cover = 10% . Black solid contours indicate regions where the differences are statistically significant ($p\text{-value} < 0.01$). (c) Time series of southward phosphate transport ($\text{mol}\cdot\text{s}^{-1}$) across Davis Strait, between 2004 and 2016, for VS (blue) and VD (red). The black line shows the differences between the simulations (VD - VS).

4 Conclusions

To the extent of our knowledge, all previous iceberg modelling studies (e.g. Bigg et al., 1997, Gladstone et al., 2001; Marsh et al. 2015; Marson et al., 2018, Marson et al., 2024; Martin and Adcroft, 2010; Merino et al., 2016) have prescribed the freshwater input from icebergs at the ocean surface. However, laboratory studies and in situ observations reveal that this is not always the case (e.g., Fitzmaurice et al., 2017; Lucas et al., 2025). Therefore, we used two simulations carried out with NEMO v3.6 to assess the impacts of representing the release of iceberg meltwater along the iceberg's keel (simulation **VD**), instead of at the surface (simulation **VS**). The core of this study is based on the hypothesis that parameterizing the injection of iceberg meltwater at depth could enhance vertical mixing due to the rising buoyant plumes that entrain surrounding waters. The enhanced mixing could impact the melting of icebergs itself, along with changes in physical properties, ocean dynamics, and primary production. Our findings, summarized below, not only confirmed this hypothesis, but also showed that the impacts are more significant in terms of magnitude and spatial scale than expected.

The direct impacts of the release of iceberg meltwater at depth in the model are a consequence of vertical mixing, which was enhanced as we had hypothesized. The mixing generated by the rising plume weakens the stratification of the water column and creates a positive feedback by increasing the melting rate of icebergs since the plume entrains subsurface ambient waters that are usually warmer (and saltier) than the surface (McCutchan and Johnson, 2022; McCutchan et al., 2024). This entrainment can be seen in our results as positive salinity and temperature anomalies in regions densely populated by icebergs, such as eastern Baffin Bay and southern Greenland. The enhanced vertical mixing promotes nutrient upwelling, as seen by previous studies (Smith et al., 2013; Vernet et al., 2012). Additionally, the intrusion of freshwater at depth impacts how the meltwater is distributed in large-scale, since it mixes with its surroundings instead of ascending undiluted to the surface, as observed by FitzMaurice et al. (2017). Besides the obvious enhanced dispersion into greater depths, iceberg meltwater is also able to travel more into the interior of basins (e.g. Baffin Bay and Subpolar North Atlantic).

Injecting iceberg melt at depth also leads to large-scale, indirect impacts. For example, we found that the steric height contribution of vertically distributed iceberg plumes is smaller than that of undiluted, superficial plumes. This translates into a relatively lower SSH in Baffin Bay, which causes an increased outflow of Arctic Water. The additional volume of Arctic Water flowing southwards not only cools and freshens the west side of Baffin Bay but also enhances the western subpolar gyre. Moreover, these cold waters bring more nutrients towards the Labrador Sea, promoting primary production in that region. Finally, the enhanced Arctic outflow contributes to the acceleration of the Baffin Island Current, which strengthens the Baffin Bay cyclonic gyre and the associated Ekman upwelling (e.g., Castro de la Guardia et al., 2015), bringing saltier and colder deep waters into the subsurface layers. Another example of indirect impact is the increase in the depth of the winter mixed layer in the Labrador Sea (and in the area where deep convection might occur). When the iceberg meltwater is injected at depth, the southwestern Greenland shelf waters become less fresh. Since a substantial volume of those waters is exported to the interior of the Labrador Sea (Pennelly et al., 2019), the advection of saltier waters into this region leads to a higher density of the upper ocean layers, decreasing the water column stability. Additionally, the higher concentration of nutrients in upper layers in Baffin Bay and in the Arctic, in combination with the strengthening of the southward volume transport, leads to an enhanced advection of nutrients to the interior of the Labrador Sea and, therefore, larger spring blooms. Perhaps the most puzzling finding is that the release of meltwater at depth leads to the weakening and contraction of the eastern subpolar gyre. As reported by previous literature (Desbruyères et al., 2021; Kenigson and Timmermans, 2020; Koul et al., 2020), the weakening and contraction of the SPG allows more subtropical waters to enter this region, resulting higher values of temperature and salinity in the subpolar North Atlantic.

Despite the strengths and consistency of this study, there are a number of limitations that must be considered, and that shape the future directions of iceberg melt plume parameterizations. For example, the major mechanism of deterioration of icebergs is wave erosion, which in the real world leads to the release of bergy bits (El-Tahan et al., 1987, Martin and Adcroft 2010). However, the release of bergy bits is not included in the simulations used in this study. Martin and Adcroft (2010) highlighted that the representation of bergy bits delays the release of meltwater to the ocean, reducing the freshwater input on coastal regions. Hence, the freshwater input in coastal regions might be overestimated in this study. On the other hand, the model's horizontal resolution is too coarse to represent fjords, where a considerable amount (10-50%) of the mass of icebergs is lost (Enderlin et al., 2016). Therefore, our study overestimates the mass of icebergs that reach open ocean regions (i.e., part of this mass should be released along the coast as liquid runoff). Another key limitation is that, in VD, the iceberg-melt terms resulting from the different deterioration mechanisms (M_e , M_v , and M_b) are summed up and then divided equally among the layers occupied by the keel, instead of being added at their appropriate level. For example, the melt term attributed to wave erosion should be injected at the surface only, while the basal melt should enter the ocean at the iceberg draft. Moreover, laboratory experiments conducted by FitzMaurice et al. (2017) and FitzMaurice et al. (2018) demonstrate that the distribution of meltwater through the water column depends on the velocity and shear of the background flow. Therefore, a more realistic parameterization of iceberg melt plumes would also take into account the characteristics of the background flow to determine where the meltwater should be inserted in the modelled ocean.

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Supplementary Information

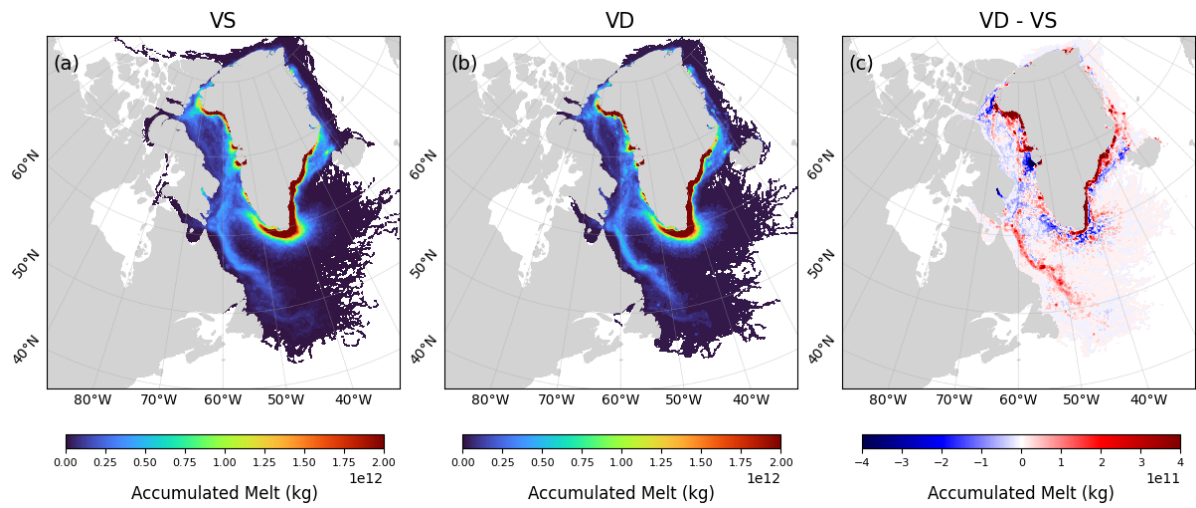


Figure 15: Accumulated melt mass (kg) between 2004 and 2017 for (a) VS and (b) VD. (c) Shows the differences between the simulations (VD - VS). (Figure S1)

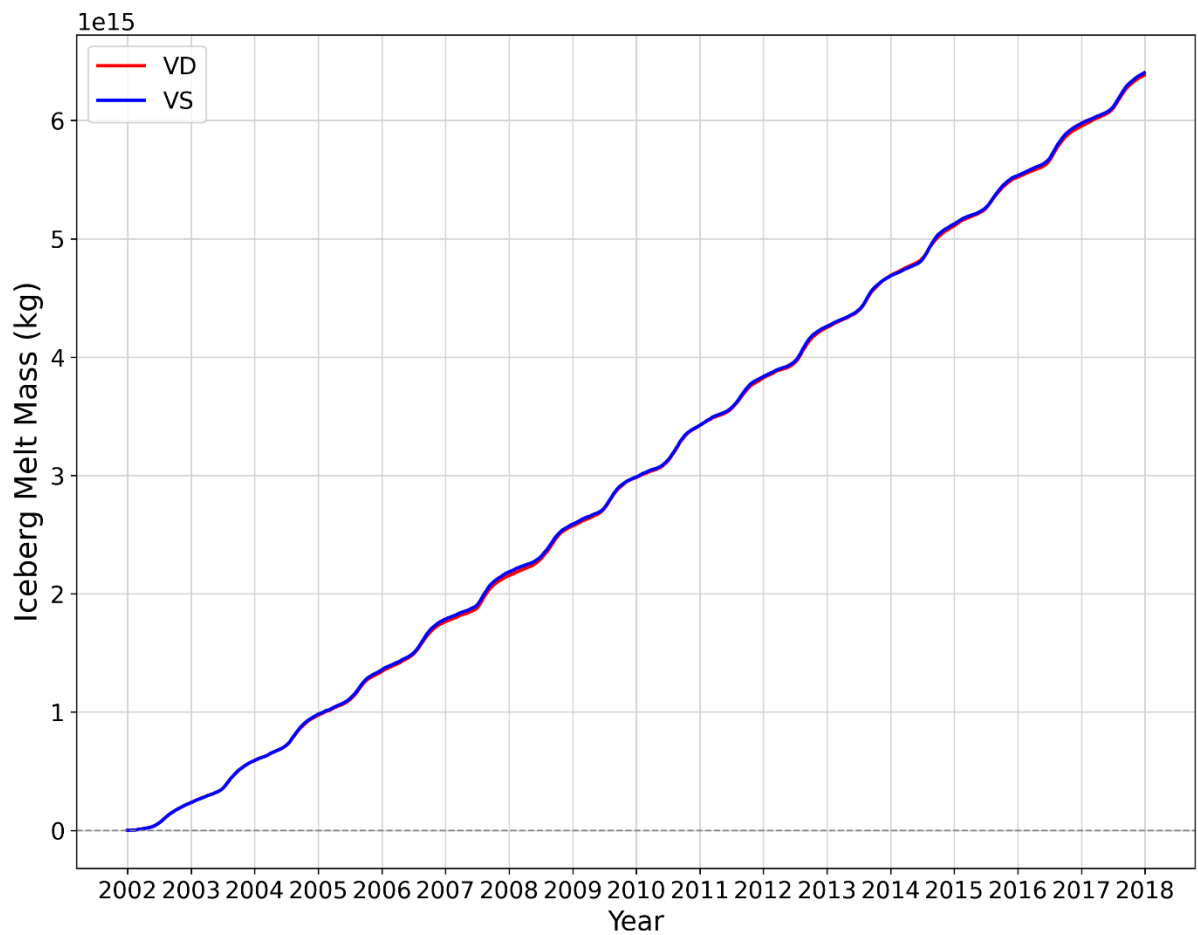


Figure 16: Time series of accumulated melt mass between 2002 and 2017 for VS (blue) and VD (red). (Figure S2)

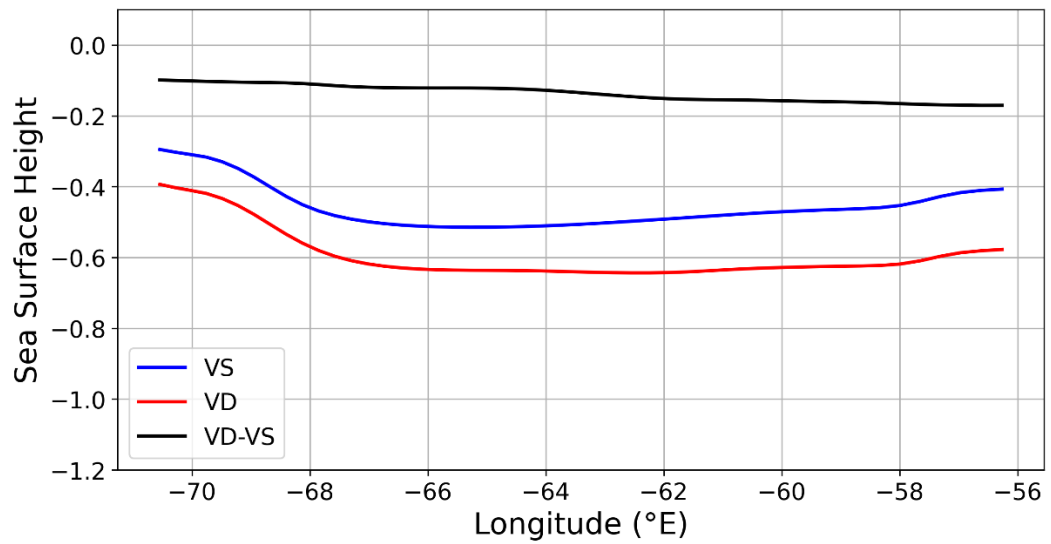


Figure 17: Zonal section of average sea surface height (m), between 2007 and 2017 for VS (blue) and VD (red). Black line shows the differences between the simulations (VD - VS). (Figure S3)

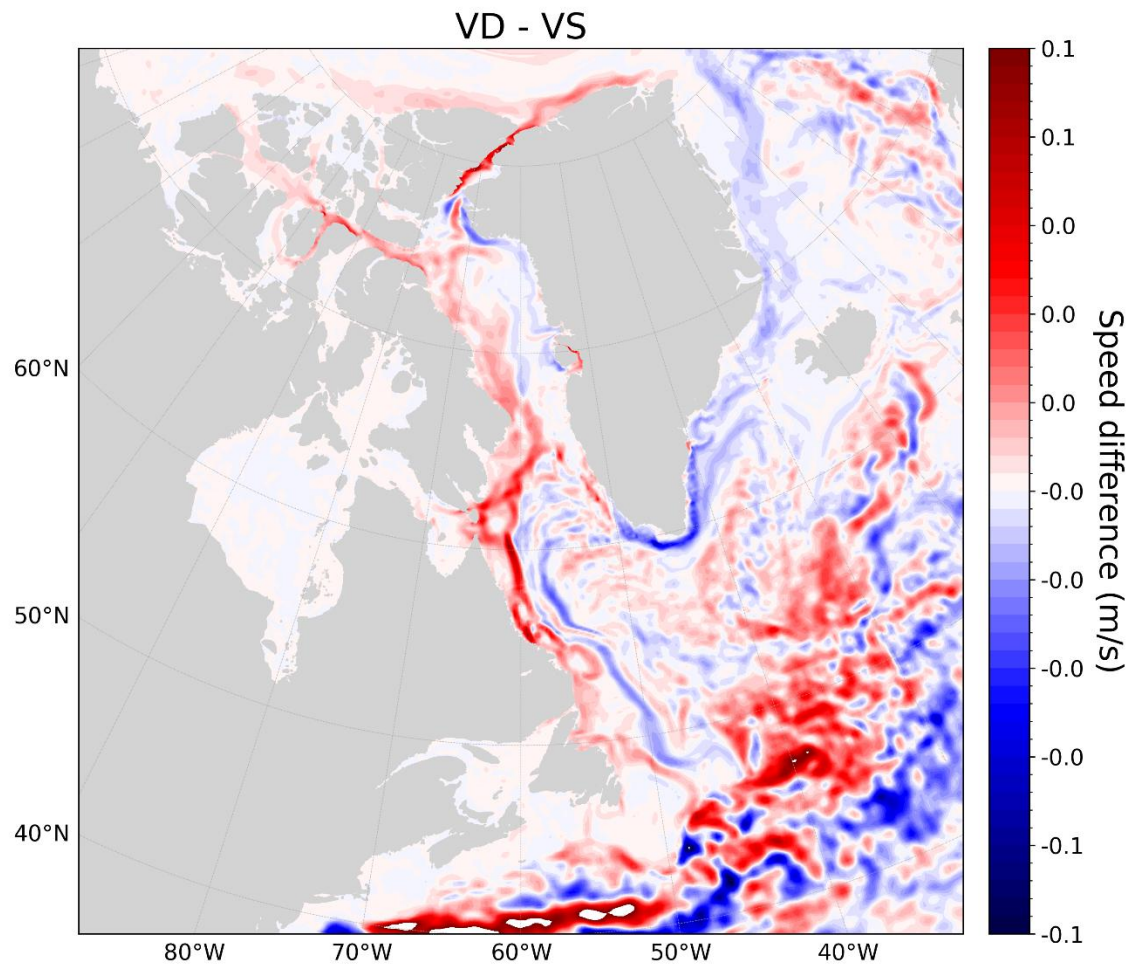


Figure 18: Differences of depth-averaged speed (surface-541m) between VD and VS, between 2007 and 2017. (Figure S4)

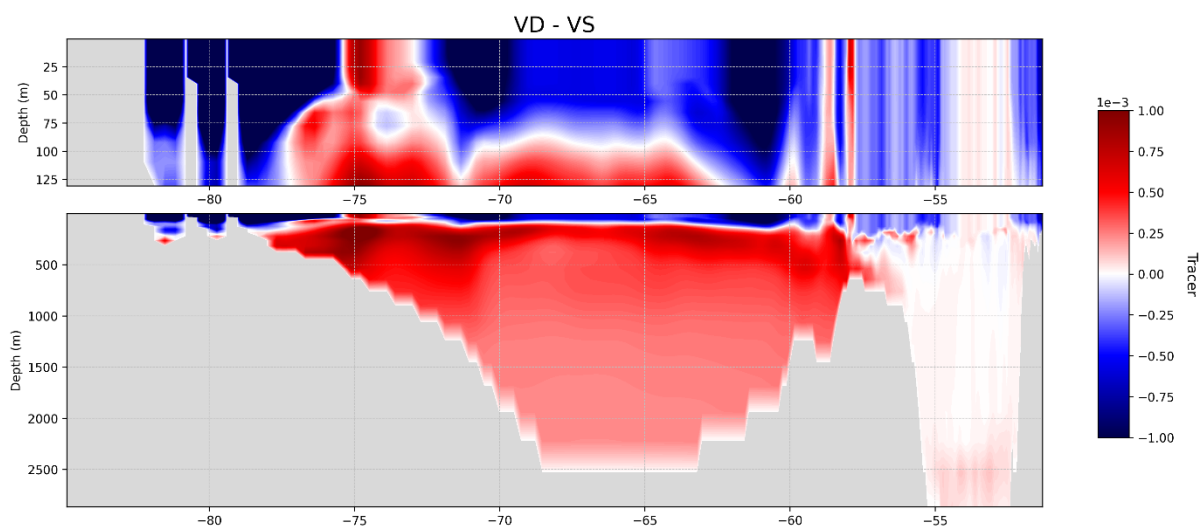


Figure 19: Difference of final tracer fields (31/12/2017) between VD and VS (VD - VS) along the transect between northern Baffin Bay and Grand Banks. (Figure S5)

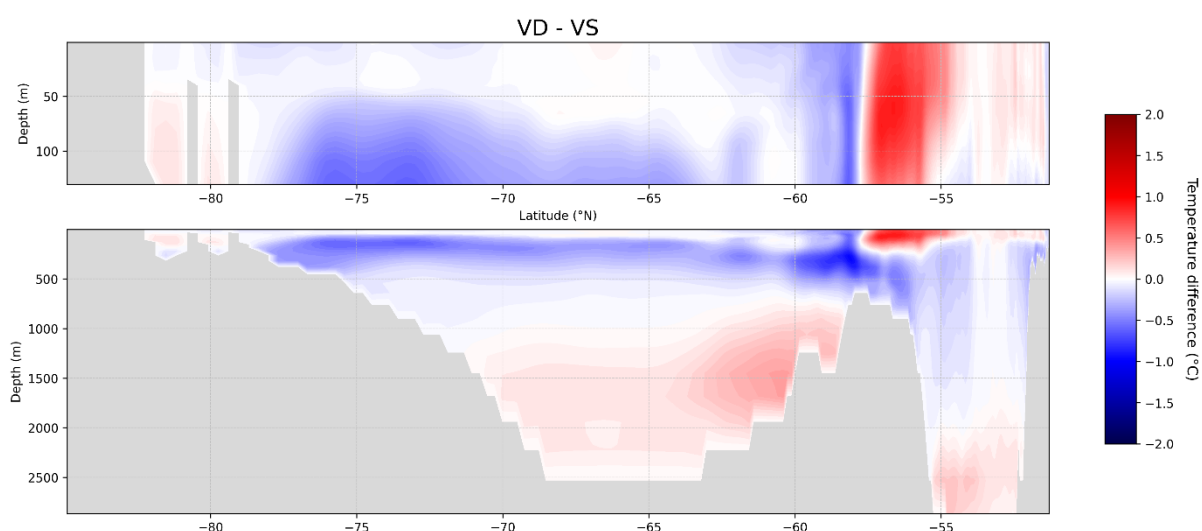


Figure 20: Difference of average temperature (2007-2017) between VD and VS (VD - VS) along the transect between northern Baffin Bay and Grand Banks. (Figure S6)

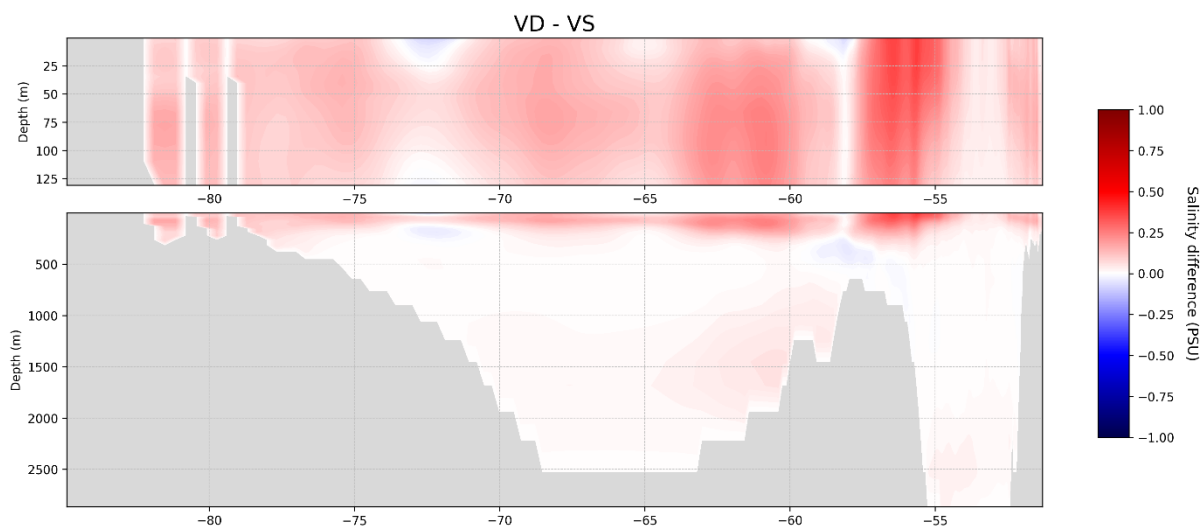


Figure 21: Difference of average salinity (2007-2017) between VD and VS (VD - VS) along the transect between northern Baffin Bay and Grand Banks. (Figure S7)

2. Additional methodology details

2.1 Model Description

The Nucleus for European Modelling of the Ocean (NEMO v3.6; Madec & the NEMO team, 2015) modelling framework was used to carry out the simulations used in the present work. NEMO is a primitive equation-based numerical model, which allows multiple components of the environment to interact through coupling. The ocean state in NEMO is governed by six main primitive equations: the momentum balance (Eq. (1)), the hydrostatic equilibrium (Eq. (2)), the incompressibility equation (Eq. (3)), the heat conservation equation (Eq. (4)), the salt conservation equation (Eq. (5)) and the equation of state (Eq. (6)):

$$\frac{\partial \mathbf{U}_h}{\partial t} = - \left[(\nabla \times \mathbf{U}) \times \mathbf{U} + \frac{1}{2} \nabla (\mathbf{U}^2) \right]_h - f \mathbf{k} \times \mathbf{U}_h - \frac{1}{\rho_o} \nabla_h p + \mathbf{D}^U + \mathbf{F}^U \quad (1)$$

$$\frac{\partial p}{\partial z} = -\rho g \quad (2)$$

$$\nabla \cdot \mathbf{U} = 0 \quad (3)$$

$$\frac{\partial T}{\partial t} = -\nabla \cdot (T \mathbf{U}) + D^T + F^T \quad (4)$$

$$\frac{\partial S}{\partial t} = -\nabla \cdot (S \mathbf{U}) + D^S + F^S \quad (5)$$

$$\rho = \rho(T, S, p) \quad (6)$$

Since the gravitational force is prevailing in the equations of large-scale motion, an orthogonal set of unit vectors (i,j,k) is employed, where (i,j) are tangent to the geopotential surfaces, and k represents the local upward direction, orthogonal to both i and j. In the equations, $\mathbf{U}$ is the velocity vector, t is time, ∇ is the derivative vector operator in (i, j, k) directions, $f=2\Omega \cdot \mathbf{k}$ is the Coriolis term, ρ_o is reference density, ρ is the in situ density, p is pressure, z is the vertical coordinate, g is gravitational acceleration, T is potential temperature and S is salinity. D^U , D^T and D^S are the parameterizations of small-scale physics, and F^U , F^T and F^S are the

surface forcing terms for momentum, temperature and salinity. Considering the typical ranges for large-scale motions, the 6 following approximations are made:

- **Spherical Earth approximation:** The earth is assumed to be a sphere, instead of a geoid. Thus, the gravity vector is considered to be parallel to the Earth's radius instead of slightly varying between points.
- **Thin-shell approximation:** The ocean depth ($\sim 10\text{km}$) is considerably smaller than the Earth's radius ($\sim 6,400\text{ km}$). Thus, the gravity force is not affected by the ocean depth, being constant in every grid point.
- **Turbulent closure hypothesis:** Small-scale processes, represented by turbulent fluxes, are expressed in terms of large-scale features.
- **Boussinesq hypothesis:** Since the density in the open ocean shows little variations between the surface and deeper ocean, density variations are only considered in their contribution to buoyancy force.
- **Hydrostatic hypothesis:** Convective processes are removed from the initial Navier-Stokes equations, and are parameterized. Therefore, the vertical momentum is ruled by the balance between the vertical pressure gradient and the buoyancy force.
- **Incompressibility hypothesis:** The three-dimensional divergence of the velocity vector is considered to be zero, since the ocean water is assumed to be incompressible.

The model's equations are discretized by the traditional centered second-order finite difference. The variables are distributed in an Arakawa C-grid, with vector points in the center of each face and scalar points in the center of the cell, as shown in Figure 22 (Arakawa, 1997).

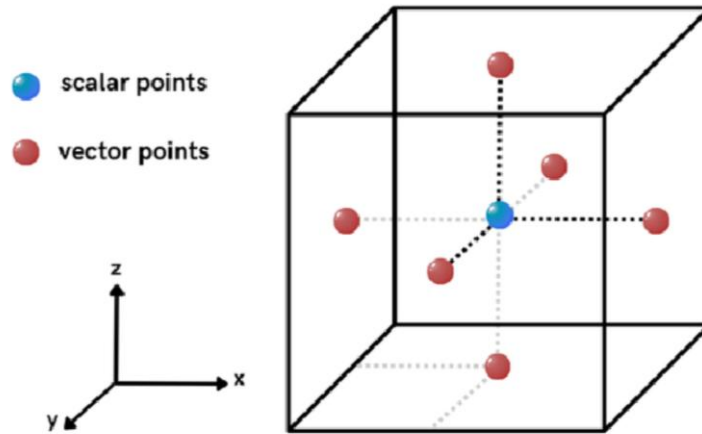


Figure 22: Arakawa C grid, with scalar points in the center and vector points in the faces of the cell.

The ocean is constrained by complex coastlines and bottom topography below and laterally, as well as by the interface between ocean-atmosphere and ocean-sea-ice above. These boundaries are defined by two surfaces: $z = -H(i, j)$ and $z = \eta(i, j, k, t)$, where H is the depth of the ocean bottom and η is the sea surface height, referenced to the mean sea surface height. The ocean can exchange fluxes of heat, salt, freshwater and momentum with the solid earth, continental margins, sea ice and atmosphere, through the two previously mentioned boundaries. On the land-ocean interface, the major exchange is mass exchange of freshwater through river runoff, which modifies the sea surface salinity in the vicinity of river mouths. On the solid earth-ocean interface, the heat and salt fluxes are set to zero, since the fluxes through the sea floor are small. Momentum is also set to zero, since there is no velocity normal to the ocean bottom and coastlines. However, the ocean and earth exchange momentum in small scales through frictional processes, which are parameterized in terms of turbulent fluxes. On the ocean-atmosphere interface there is mass flux, represented by precipitation and evaporation, in addition to heat and horizontal momentum (wind stress). Lastly, on the ocean-sea-ice boundary there is exchange of heat, salt, freshwater and momentum, due to sea ice freezing, melting and advection.

2.2 Simulations Setup

In this project, NEMO's ocean component (OPA; Madec, et al., 1998) was coupled to the sea ice thermodynamic and dynamic model Louvain-la-Neuve Ice Model (LIM2; Fichefet and Maqueda, 1997). LIM2 is a three-layered sea ice model, with a layer of snow on top of two

layers of ice with the same thickness. These layers function as sensible heat storage and allow vertical heat conduction within the snow and ice. The simulated sea ice is not embedded in the ocean but is instead represented as a levitating layer above the ocean surface. The elastic-viscous-plastic (EVP) sea ice rheology from Hunke and Dukowicz (1997) is used in the sea ice model, where sea ice acts elastically and as a viscous plastic in short and long time scales, respectively. In this model, the sea ice exchanges heat and momentum with both the ocean and atmosphere. The formation and melting of sea ice is ruled by heat transfer in the ice-ocean and atmosphere-ocean interfaces. Also, sea ice exchanges salt with the underlying ocean due to the melting of snow and ice, brine rejection, meteoric ice, and evaporation-precipitation. NEMO's iceberg module (NEMO-ICB), implemented by Marsh et al. (2015), was used in this study and is described in detail in the next subsection

The ocean is also coupled to the simplified model of biogeochemical cycling Biogeochemistry with Light Iron Nutrients and Gasses (BLING; Galbraith et al., 2010). BLING is a prognostic model embedded in the ocean, and the simulated outputs are driven by the ocean's physical environment. The core of BLING is "NPZD" (nutrients, phytoplankton, zooplankton and detritus); however, only nutrients are explicitly represented. Four explicit tracers are used for representing nutrients: dissolved inorganic phosphorus (PO_4), dissolved organic phosphorus (DOP), dissolved iron (Fe), and oxygen (O_2). In the model, phytoplankton growth is ruled by temperature, light, macronutrients (phosphate) and micronutrients (iron). The growth rates are used to determine biomass, uptake, dissolved organic matter production and export of sinking particles, through parameterizations of the ecosystem processes.

The atmospheric component of the environment was forced with the dataset from the Canadian Meteorological Centre's Global Deterministic Prediction System's Reforecast (CGRF; Smith et al., 2014). This re-forecast product has a spatial resolution of 0.45° of longitude and 0.3° of latitude (approximately 33 km at 49°N). CGRF provides hourly fields of temperature and specific humidity at 2 m, zonal and meridional wind speed at 10 m, downward longwave and shortwave radiation, and precipitation. The period of integration of the simulations spans from 2002 to 2017; hence, CGRF data from 2002 to 2015 was used, and the 2015 forcing was repeated for the last two years.

Greenland discharge was forced with the reconstructed surface mass balance model from Bamber et al. (2018). Discharge rates vary monthly (peaking during summer) and interannually, with a spatial resolution of 5 x 5 km. The liquid component of the Greenland discharge is composed by runoff from land ice and from tundra. The solid discharge represents ice flux across the grounding line of marine-terminating glaciers and is estimated using ice thickness and satellite-based observations of glacier flow speed. Bamber et al. (2018) dataset was used from 2002 to 2016, and the last year of forcing was repeated for 2017.

The monthly climatology based on Dai and Trenberth Global River Flow and Continental Discharge Dataset (Dai et al., 2009; Dai and Trenberth, 2002), remapped onto the ANHA4 grid, was used as river runoff forcing.

The initial and boundary conditions of the simulations were provided by the reanalysis GLORYS2V3 (Masina et al., 2017). This product has a global coverage, with a spatial resolution of $1/4^\circ$ and 75 vertical levels. Initial and boundary fields used were temperature, salinity, meridional and zonal velocities, and sea surface height.

The simulations used in this project have a z-coordinate system vertical grid, with 50 vertical levels that increase in thickness with depth. The domain of the simulations covers the Arctic, North Atlantic Ocean and a portion of the Southern Atlantic Ocean, with open boundaries on Bering Strait and at 20°S in the Atlantic Ocean. The horizontal grid employed has a $1/4^\circ$ of resolution, and is a subset of the global NEMO ORCA025, named Arctic and Northern Hemisphere Atlantic (ANHA4²).

2.3 NEMO-ICB

Bamber et al. (2018) estimates indicate that around half of the mass loss of the Greenland Ice Sheet occurs in the form of solid ice discharge. Therefore, in the present simulations, the mass of icebergs released to the ocean is generally around this proportion, although it varies over time and space. The solid ice discharge rates are taken as rates of accumulation of ice in fixed

² <https://canadian-nemo-ocean-modelling-forum-community-of-practice.readthedocs.io/en/latest/Institutions/UofA/Configurations/ANHA4/index.html>

grid cells closest to Greenland’s main glacial outlets. Once there is enough mass of ice in one of those points to create a cluster of icebergs, which can be from one of the ten defined classes from Martin and Adcroft (2010) (Table 1), an iceberg particle is released in the ocean model.

Table 1. Iceberg size categories with iceberg length, thickness, mass, mass scaling factor and calving distribution. The mass scaling factor represents the number of icebergs in one model particle. The calving distribution represents the proportion of the solid discharge provided by Bamber et al. (2018) assigned to each category.

Category	Length (m)	Thickness (m)	Mass (kg)	Mass Scaling	Calving Distribution
1	60	40	$8.8 \cdot 10^7$	2000	0.24
2	100	67	$4.1 \cdot 10^8$	200	0.12
3	200	133	$3.3 \cdot 10^9$	50	0.15
4	350	175	$1.8 \cdot 10^{10}$	20	0.18
5	500	250	$3.8 \cdot 10^{10}$	10	0.12
6	700	250	$7.5 \cdot 10^{10}$	5	0.07
7	900	250	$1.2 \cdot 10^{11}$	2	0.03
8	1200	250	$2.2 \cdot 10^{11}$	1	0.03
9	1600	250	$3.9 \cdot 10^{11}$	1	0.03
10	2200	250	$7.4 \cdot 10^{11}$	1	0.02

In the ocean, iceberg clusters are treated as virtual particles, i.e., the particles are not embedded in and do not exert pressure on the ocean. However, the model represents the interaction of the iceberg particles with the velocity and temperature fields along their keels, as if they were partially submerged in the ocean (Merino et al., 2016; Marson et al., 2018). In order to simplify the calculation of the different working surfaces in the momentum and mass balance equations, the icebergs are represented as cuboids, with length L , width W and total thickness T . The iceberg’s thickness is divided into the part above the water level and the part submerged into the water, namely freeboard (F) and draft (D), respectively. The iceberg’s

draft is estimated based on the relation between the density of the icebergs and ocean, described by Eq. (7):

$$D = \frac{\rho}{\rho_o} T \simeq 0.8T \quad (7)$$

Where ρ is the iceberg's density assumed to be 850 kg.m^{-3} , following Silva et al. (2006), and $\rho_o = 1025 \text{ kg.m}^{-3}$ is the average ocean water density. The momentum balance of an iceberg of mass M is governed by the Coriolis force, the air, water and sea ice drag forces, the wave radiation force, and the horizontal pressure gradient force, described by Eq. (8):

$$M \frac{d\vec{v}}{dt} = -Mf \times \vec{v} + \vec{\tau}_a + \vec{\tau}_o + \vec{\tau}_i + \vec{F}_r + \vec{F}_p \quad (8)$$

where $\frac{d}{dt} = \frac{\partial}{\partial t} + \vec{\nabla} \cdot \vec{v}$ is the absolute derivative in time and f is the Coriolis parameter. The drag forces from the atmosphere, ocean, and ice are described by Eq. (9), Eq. (10), Eq. (11):

$$\vec{\tau}_a = \rho_a(0.5c_{a,v}WF + c_{a,h}LW) |\vec{v}_a - \vec{v}| (\vec{v}_a - \vec{v}) \quad (9)$$

$$\vec{\tau}_o = \rho_o(0.5c_{o,v}W(D - T_i) + c_{o,h}LW) |\vec{v}_o - \vec{v}| (\vec{v}_o - \vec{v}) \quad (10)$$

$$\vec{\tau}_i = \rho_i(0.5c_{i,v}WT_i) |\vec{v}_i - \vec{v}| (\vec{v}_i - \vec{v}) \quad (11)$$

where a , o and i are the indexes for atmosphere, ocean and sea ice, respectively. ρ_a , ρ_o and ρ_i are the average air, ocean and sea ice densities. Following Gladstone et al. (2001), $c_{a,v} = 1.3$ is the atmosphere vertical drag coefficient, $c_{a,h} = 0.0055$ is the atmosphere horizontal drag coefficient, $c_{o,v} = 0.9$ is the ocean vertical drag coefficient and $c_{o,h} = 0.0012$ is the ocean horizontal drag coefficient. The vertical drag coefficient for sea ice $c_{i,v}$ is assumed to be the same as the ocean vertical drag coefficient. The wave radiation force is described by Eq. (11):

$$F_r = \frac{1}{2} \rho_o C_r g a \min(a, F) \frac{2LW}{L + W} \frac{\vec{v}_a}{|\vec{v}_a|} \quad (12)$$

where g is the acceleration due to gravity and a is the wave amplitude, related to the wind speed, and C_r is the wave radiation coefficient, described by Eq. (13):

$$C_r = 0.06 \min \left(\max \left[0, \frac{L - L_c}{L_t - L_c} \right], 1 \right) \quad (13)$$

where the cutoff length ($L_c = 0.125 L_w$) and the upper limit ($L_t = 0.25 L_w$) are chosen following Carrieres et al. (2001), with the wavelength described by Eq. (14):

$$L_w = 0.32 |\vec{v}_a - \vec{v}_o|^2 \quad (14)$$

The pressure gradient force includes the impact of the sea surface slope (η) in the momentum balance of icebergs, and is described by Eq. (15):

$$\tilde{F}_p = -Mg\vec{\nabla}\eta \quad (15)$$

The mass balance of the iceberg particles is governed by basal melt (M_b), side melt (M_v) and wave erosion (M_e), since these are the major mechanisms in iceberg deterioration (Gladstone et al., 2001). This balance is described by Eq. (16):

$$\rho \frac{d(LWT)}{dt} = \rho(-LWM_b - T(L+W)(M_e + M_v)) \quad (16)$$

Basal melt rate is described by Eq. (17):

$$M_b = 0.58|\vec{v} - \vec{v}_b|^{0.8} \frac{\tilde{T}_b - \tilde{T}}{L^{0.2}} \quad (17)$$

where $\vec{v}$ is the iceberg velocity (m.s^{-1}), $\vec{v}_b$ is the ocean velocity at the base of the iceberg (m.s^{-1}), $\tilde{T}_b$ is the ocean temperature at the base of the iceberg ($^{\circ}\text{C}$), and $\tilde{T}$ is the iceberg temperature ($^{\circ}\text{C}$), assumed to be constant at -4°C (Løset, 1993). When icebergs have a draft larger than the bathymetry depth, $\vec{v}_b$ and $\tilde{T}_b$ are assumed to be the ocean velocity and temperature at the sea floor depth. The side melt rate parameterization is described by Eq. (18):

$$M_v = \int_{-H_{icb}}^0 (7.62 \times 10^{-3}\tilde{T}_o(h) + 1.29 \times 10^{-3}\tilde{T}_o^2(h)) dh \quad (18)$$

where H_{icb} is the depth reached by the iceberg's keel (m). The wave erosion rate parameterization is described by Eq. (19):

$$M_e = \frac{1}{12}S_s(1 + \cos[\pi A_i^3])(\tilde{T}_o + 2) \quad (19)$$

where A_i is the sea ice area (m^2) and S_s is the sea state, which follows Eq. (20):

$$S_s = \frac{3}{2}|\vec{v}_a - \vec{v}_o|^{\frac{1}{2}} + \frac{1}{10}|\vec{v}_a - \vec{v}_o| \quad (20)$$

Basal melt, side melt, and wave erosion result in changes in the iceberg's dimensions L , W and T , in addition to modifications in ocean properties due to the freshwater and latent heat fluxes, in the ocean layers where the meltwater is inserted. The latent heat flux Q_h (W.m^{-2}), due to melting-freezing, follow Grosfeld et al. (1997), and is described by Eq. (21):

$$Q_h = \rho c_p \gamma (T_w - T_f) \quad (21)$$

where ρ (kg.m^{-3}) is the ocean density, c_p (J.kg/K) is the ocean specific heat capacity (assumed to be 3992 J.kg.K^{-1}), γ is the thermal exchange coefficient, T_w ($^{\circ}\text{C}$) is the ocean temperature averaged over the iceberg's keel, and T_f ($^{\circ}\text{C}$) is the freezing point.

2.4 Parameterizing the injection of iceberg meltwater into the ocean

During the present work four simulations were used, with different forms of parameterizing the input of meltwater from icebergs in the ocean. The simulations differ in terms of the numerical representation of the iceberg melt plumes, as well as depth of intrusion of the plumes. The iceberg melt plumes were parameterized as a:

- Salinity anomaly at the ocean surface (SS)
- Volume flux at the ocean surface (VS)
- Salinity anomaly through depth, along the iceberg's keel (SD)
- Volume flux through depth, along the iceberg's keel (VD)

In terms of numerical representation, the iceberg melt plumes were parameterized as salinity anomalies or volume fluxes. For the “salinity anomaly” numerical representation (SS and SD) the release of iceberg melt plumes is represented as a virtual salt flux and an effective heat flux, with no volume change, following Losch (2008). For the “volume flux” numerical representation (VS and VD) the release of iceberg meltwater results in an adjustment of the velocity divergence to represent a volume flux of freshwater at the freezing point. With respect to the depth of intrusion, the iceberg meltwater release was parameterized at the surface (first ocean vertical level) or through the iceberg keel. The method that considers the release of iceberg meltwater through depth was adapted from Mathiot et al. (2017), originally applied for ice shelves. In this parameterization, the freshwater and latent heat fluxes due to the melting of icebergs are uniformly distributed along their keels rather than restricted to the first ocean level, resulting in plumes along the side walls of the icebergs. The parameterizations are summarized in Table 2.

Table 2. Summary of the numerical representation and depth of intrusion parameterizations for the four simulations.

Simulation	Numerical Representation of the Iceberg Melt Plumes	Depth of Intrusion of the Iceberg Melt Plumes
Salinity anomaly at the surface (SS)	Effective heat flux and virtual salt flux following Losch (2008)	Input at the first ocean level
Volume flux at the surface (VS)	Divergence of velocity is adjusted to represent iceberg melt as a volume flux of freshwater at freezing point	Input at the first ocean level
Salinity anomaly through depth (SD)	Effective heat flux and virtual salt flux following Losch (2008)	Input along the iceberg keel, following Mathiot et al. (2017)
Volume flux through depth (VD)	Divergence of velocity is adjusted to represent iceberg melt as a volume flux of freshwater at freezing point	Input along the iceberg keel, following Mathiot et al. (2017)

2.5 Additional information on the analyses of outputs

The analysis of the outputs included the period from 2004 to 2017, since in the first two years the model is still adjusting to the atmospheric forcing. For all the following analyses, the differences between the simulations were performed. To assess the impacts of prescribing the iceberg melt plumes as volume fluxes or as salinity anomalies, the differences between the VD and SD, and VS and SS was done (VD - SD and VS - SS). Additionally, to better understand the impacts of inserting the iceberg meltwater at the ocean surface or through depth, the differences between VD and VS, and SD and SS was done (VD - VS and SD - SS). Since the differences between the simulations with different numerical representation have shown minor differences (see Appendix), we focused exclusively on the differences between VD and VS. We chose the simulations which numerically represent the iceberg melt plumes as volume fluxes, since adjusting the velocity divergence produces more turbulence, and is more realistic. More details about output analysis are described in Chapter 1, section 2.3.

To investigate potential impacts of the iceberg meltwater parameterization on the strength and position of the gyres (i.e. Subpolar Gyre and Baffin Bay Gyre), the barotropic

streamfunction (ψ ; BSTF) was calculated. BSTF is the ocean's vertically-integrated horizontal flow, which reflects the horizontal volume transport, in Sv ($1\text{Sv} = 10^6\text{m}^3.\text{s}^{-1}$). For the northern hemisphere, positive values indicate clockwise circulation (i.e. North Atlantic Subtropical Gyre), while negative values correspond to counterclockwise circulation (i.e. Subpolar Gyre). In order to capture the main near-surface currents, as well as focus on the same levels as the previous analysis, the BSTF was integrated from surface to 541m. This variable was calculated by solving a Poisson equation forced by the relative vorticity, shown in Eq. (22):

$$\nabla^2\psi = \zeta \quad (22)$$

where ∇^2 is the horizontal Laplacian operator and ζ is the relative vorticity, estimated by the Eq. (23):

$$\zeta = \frac{dV}{dx} - \frac{dU}{dy} \quad (23)$$

with U and V being the barotropic velocities, described by Eq. (24) and Eq. (25):

$$U(x, y) = \int_{-541}^0 u(x, y, z) dz \quad (24)$$

$$V(x, y) = \int_{-541}^0 v(x, y, z) dz \quad (25)$$

Additionally, to determine if the differences between the simulations were statistically significant a t-test was performed in parallel with the above-mentioned analysis. The t-test determines if there are significant differences between the averages (in time) of the samples (simulations outputs). The sample averages are considered independent and normally distributed, and the samples are assumed to have equal variances. Alongside with the t-test, p-values were calculated. Contours of p-value < 0.01 were added in the maps of all the variables, to show the regions where the differences were statistically significant.

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3. Appendix

This appendix presents the differences between the initial simulations used in this study. It provides supplementary information to support the decision of focusing on the differences between VD and VS. For all the following figures, it is possible to notice that the differences between the simulations which differ in terms of the depth of intrusion of iceberg melt plumes (VD – VS and SD – SS) exhibit highly similar patterns. Most importantly, for most variables these differences are substantially more pronounced than the differences between the simulations which differ in terms of the numerical representation of the iceberg melt plumes (VD – SD and VS – SS). Also, the differences between the simulations with differences in terms of the numerical representation of the iceberg melt plumes not always show consistent patterns.

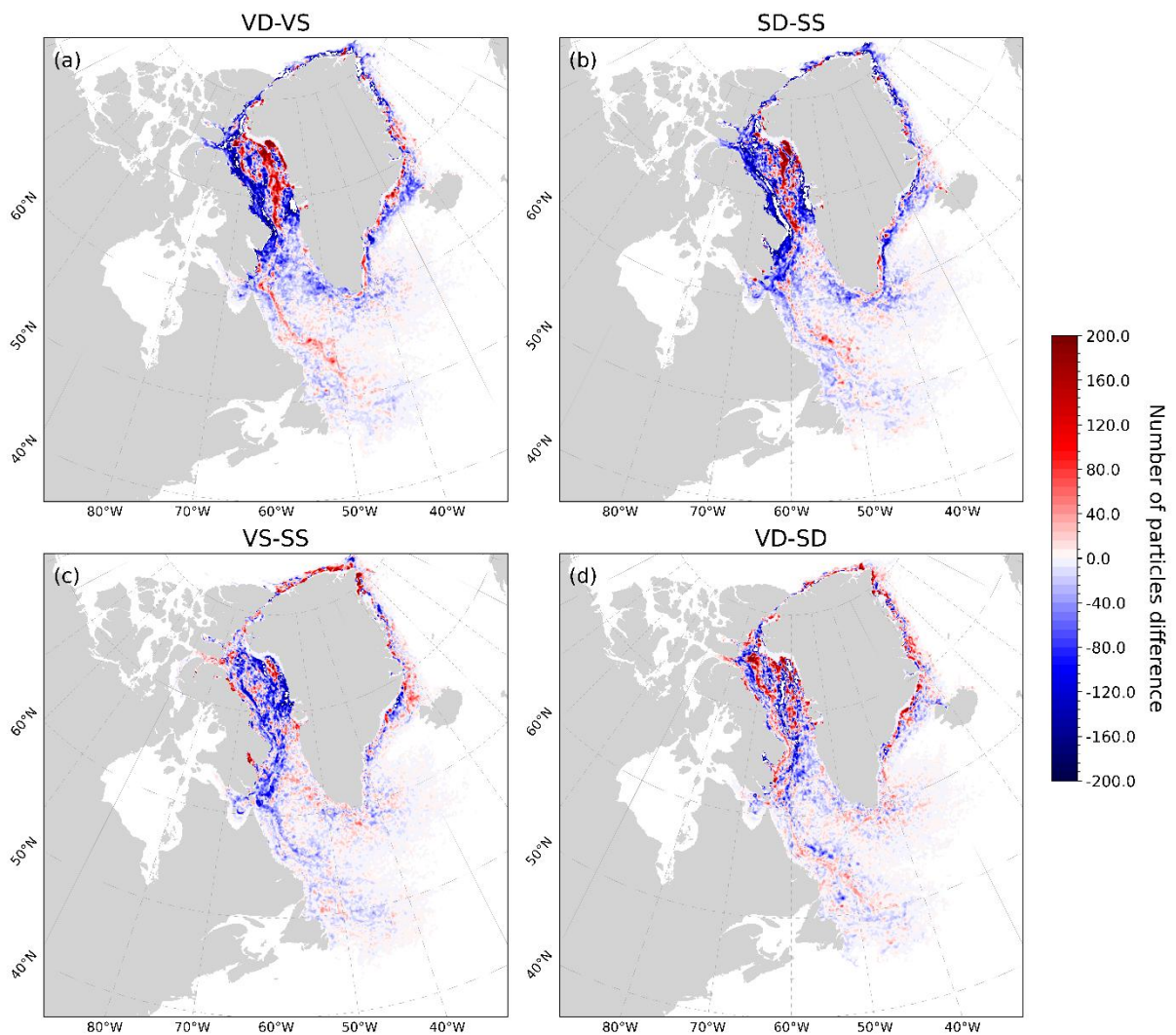


Figure 23: Differences of the total number of iceberg particles that passed through each grid point from 2002 to 2017 between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

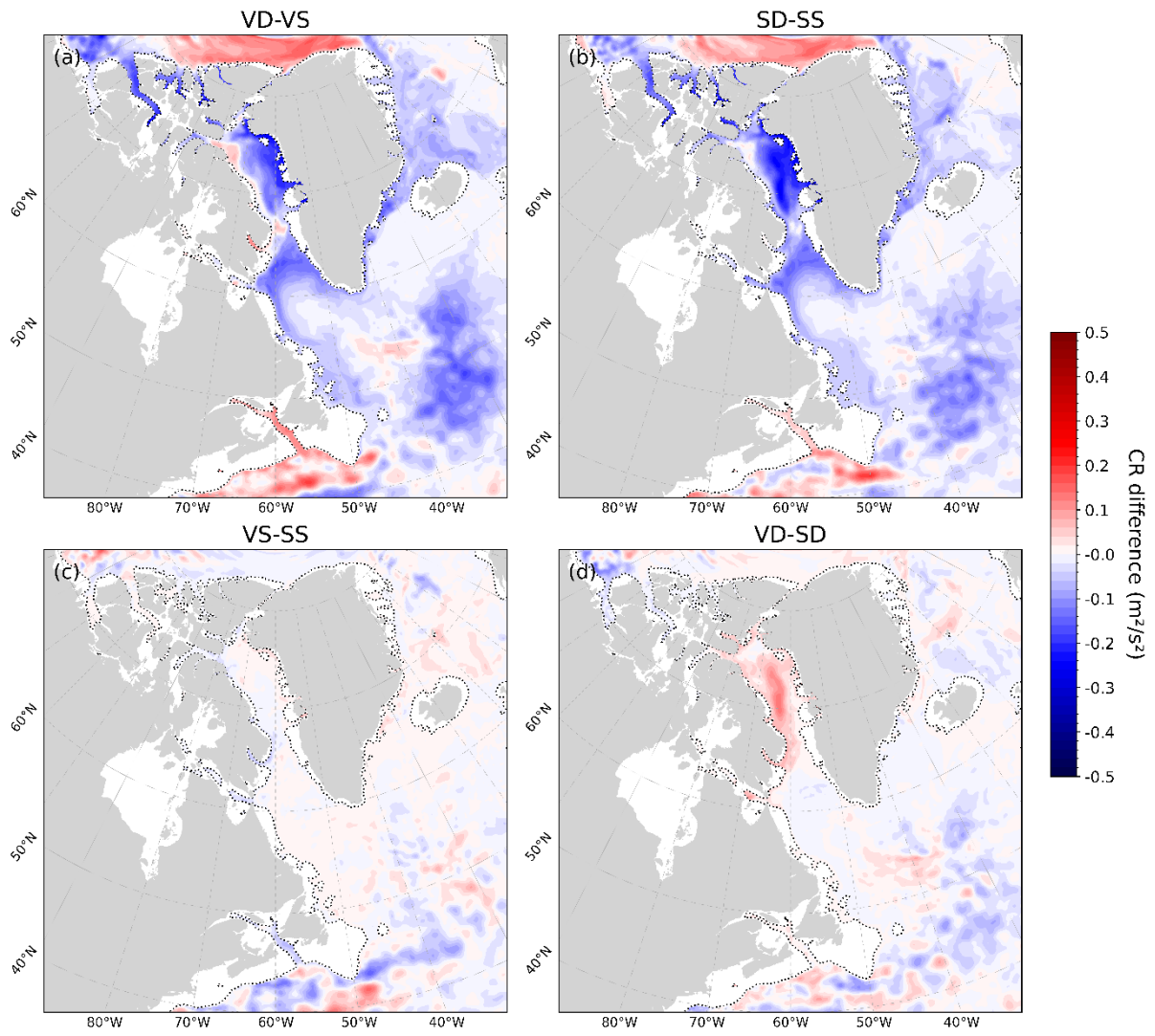


Figure 24: Differences of mean (2007-2017) convective resistance to a reference depth of 250 m between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD. Black dotted contours show the 250m isobath.

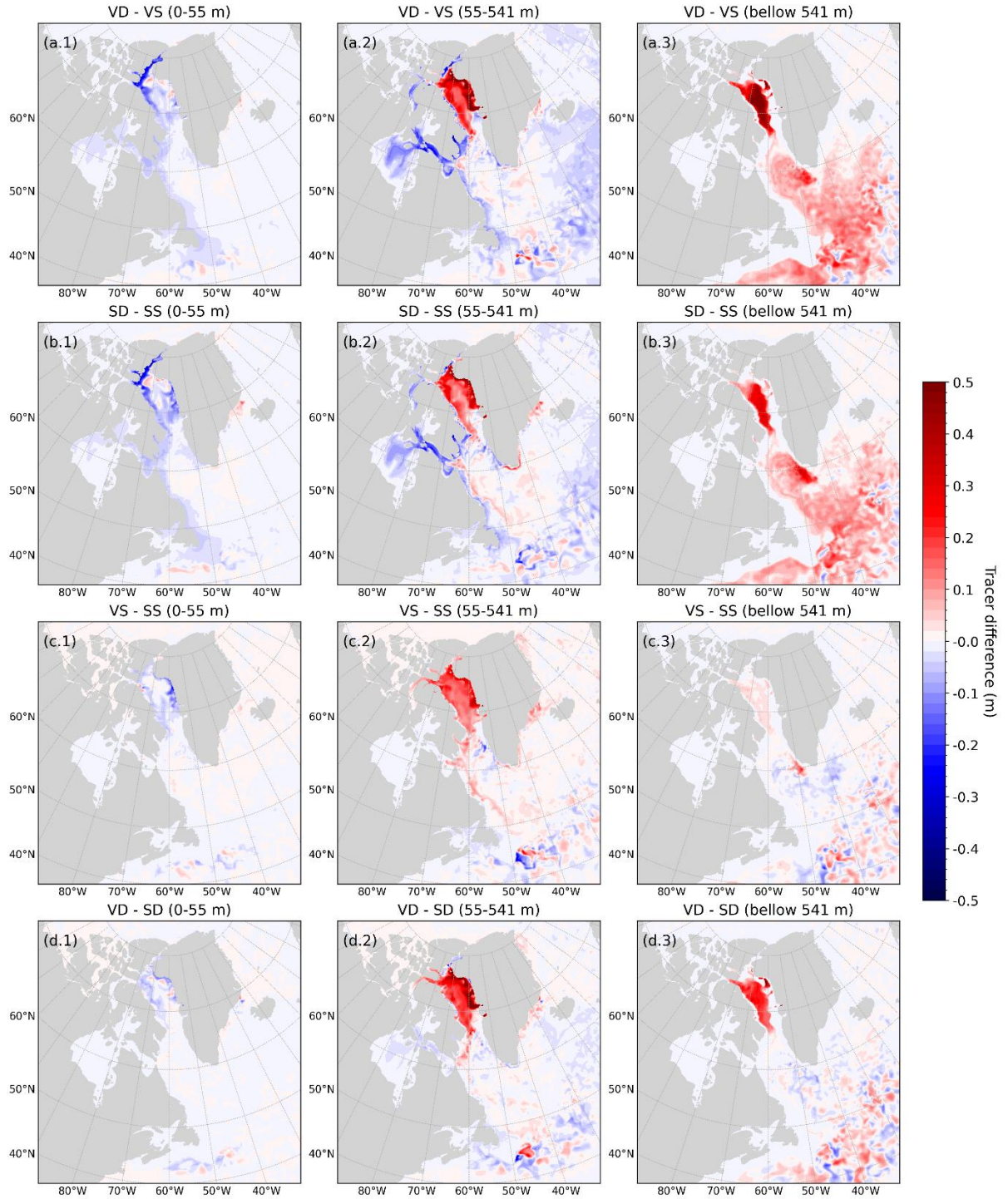


Figure 25: Differences of the final distribution (December 31st, 2017) of integrated iceberg melt tracers in the upper (0-55m) (1), subsurface (55-541m) (2) and lower (below 541m) (3) layers, between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

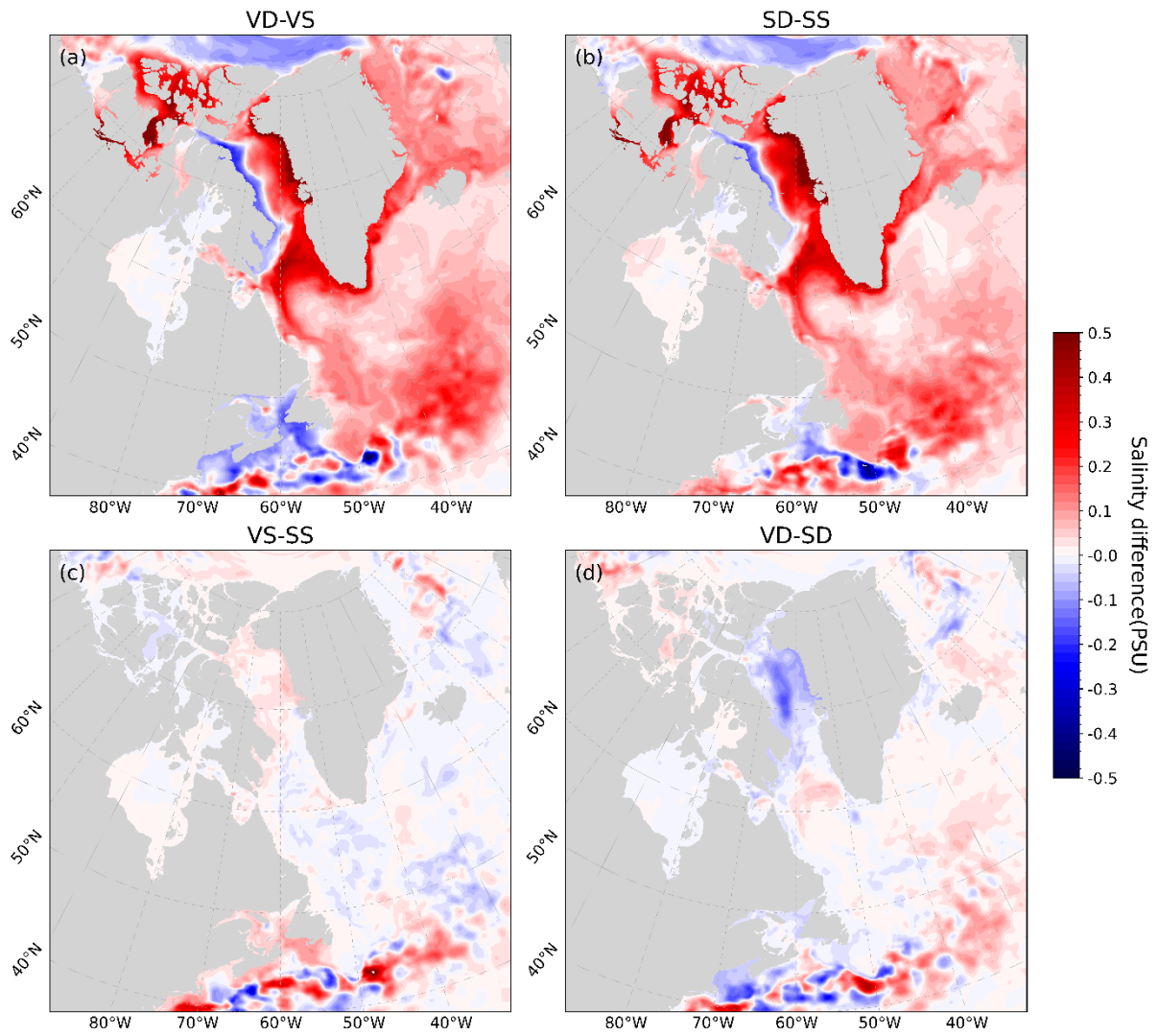


Figure 26: Differences of depth-averaged salinity for the upper layer (0-55m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

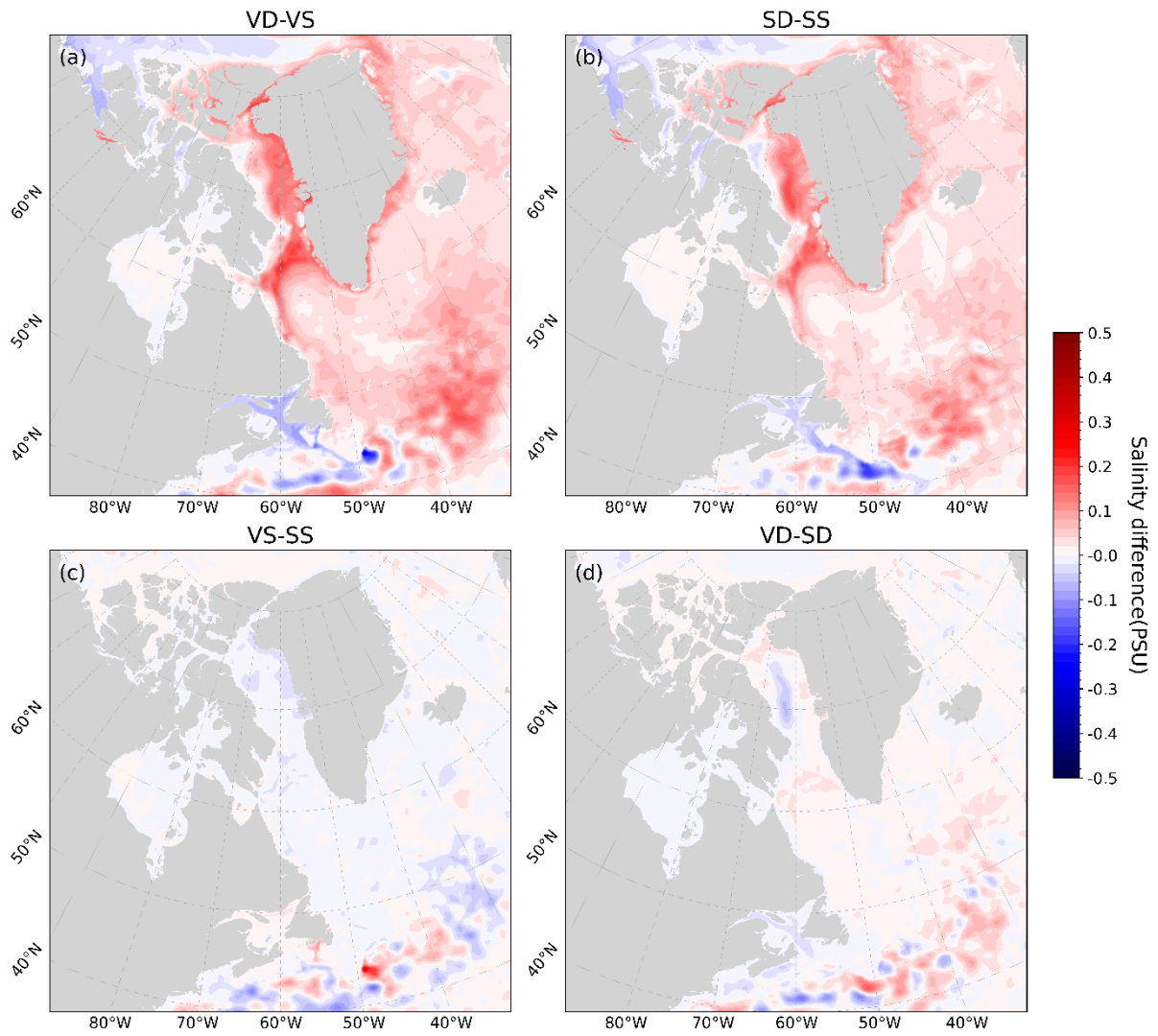


Figure 27: Differences of depth-averaged salinity for the lower layer (55-541m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

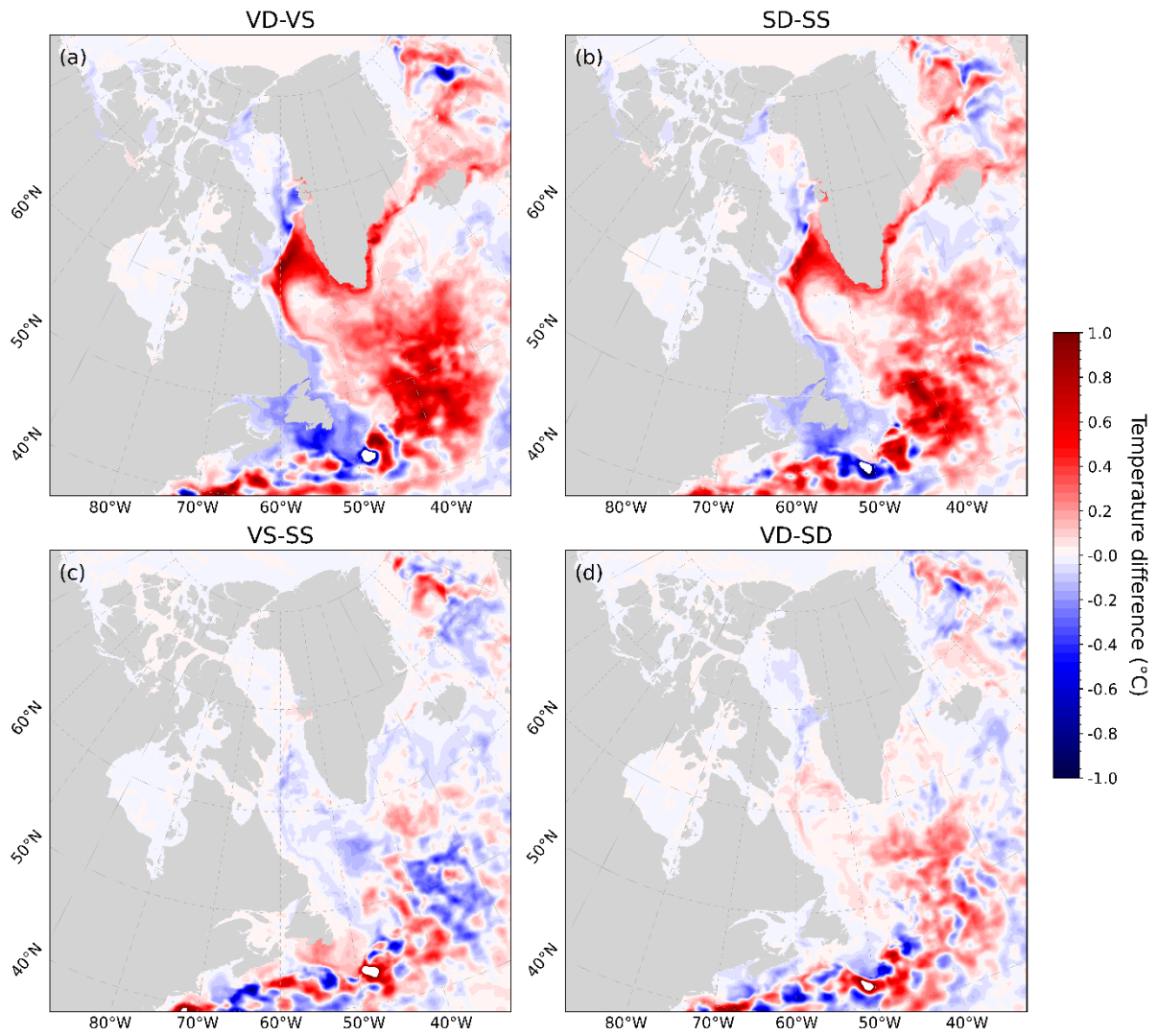


Figure 28: Differences of depth-averaged temperature for the upper layer (0-55m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

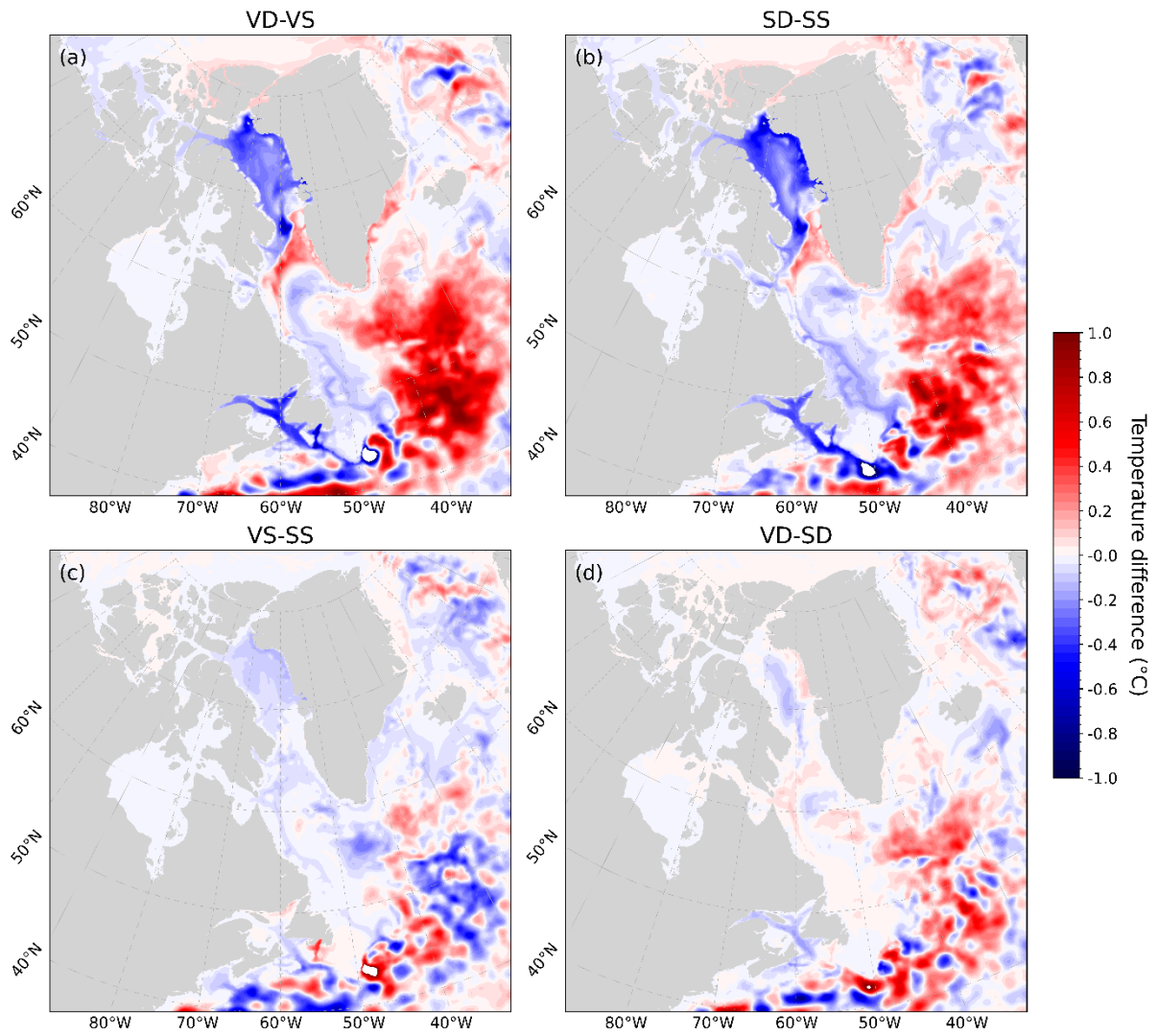


Figure 29: Differences of depth-averaged temperature for the lower layer (55-541m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

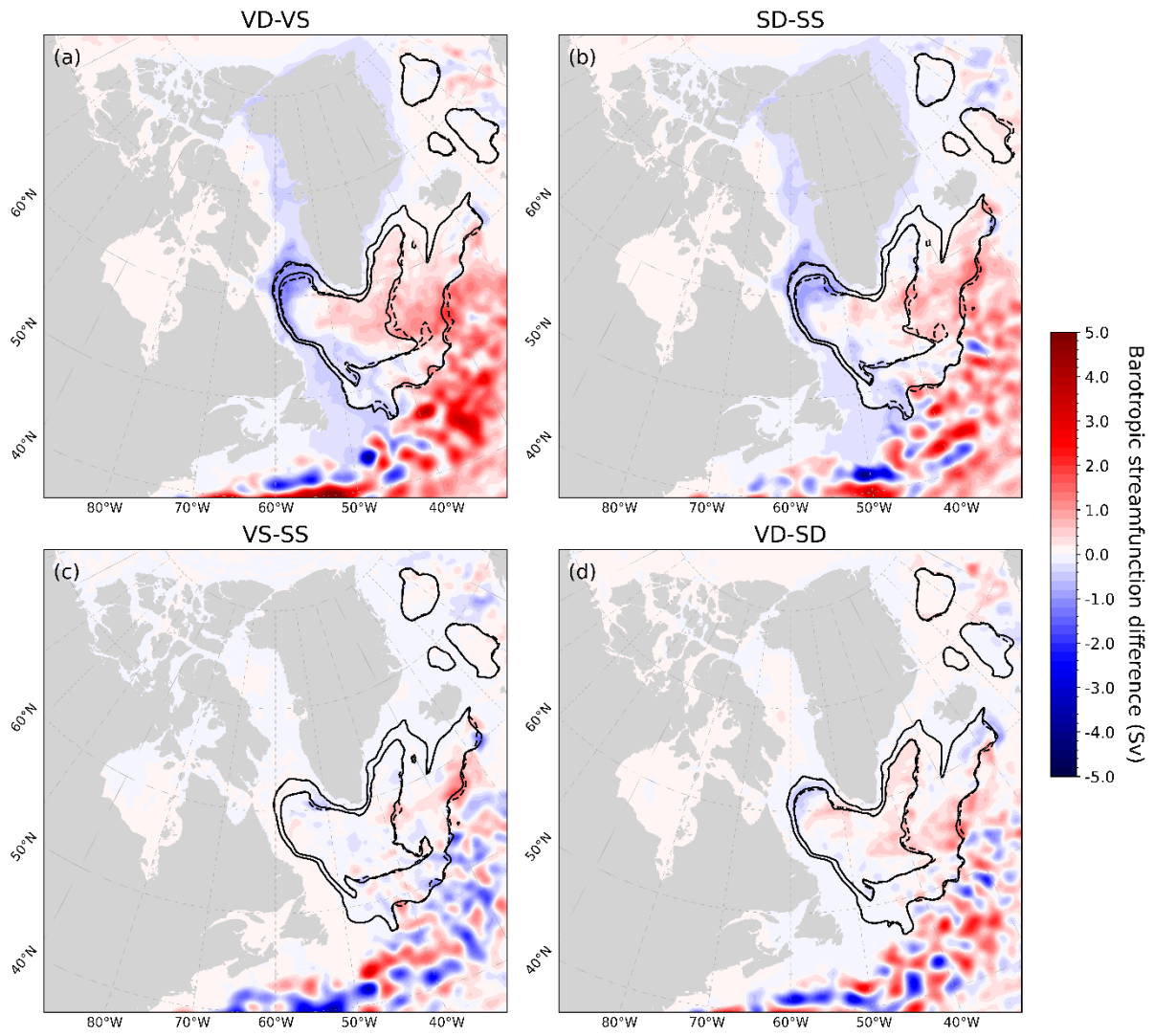


Figure 30: Differences of depth-integrated barotropic streamfunction (surface – 541m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD. Solid lines represent the -5 Sv and -10 Sv isolines of the simulation from which values are subtracted, while the dashed lines represent the -5 Sv and -10 Sv isolines of the simulation used for subtraction.

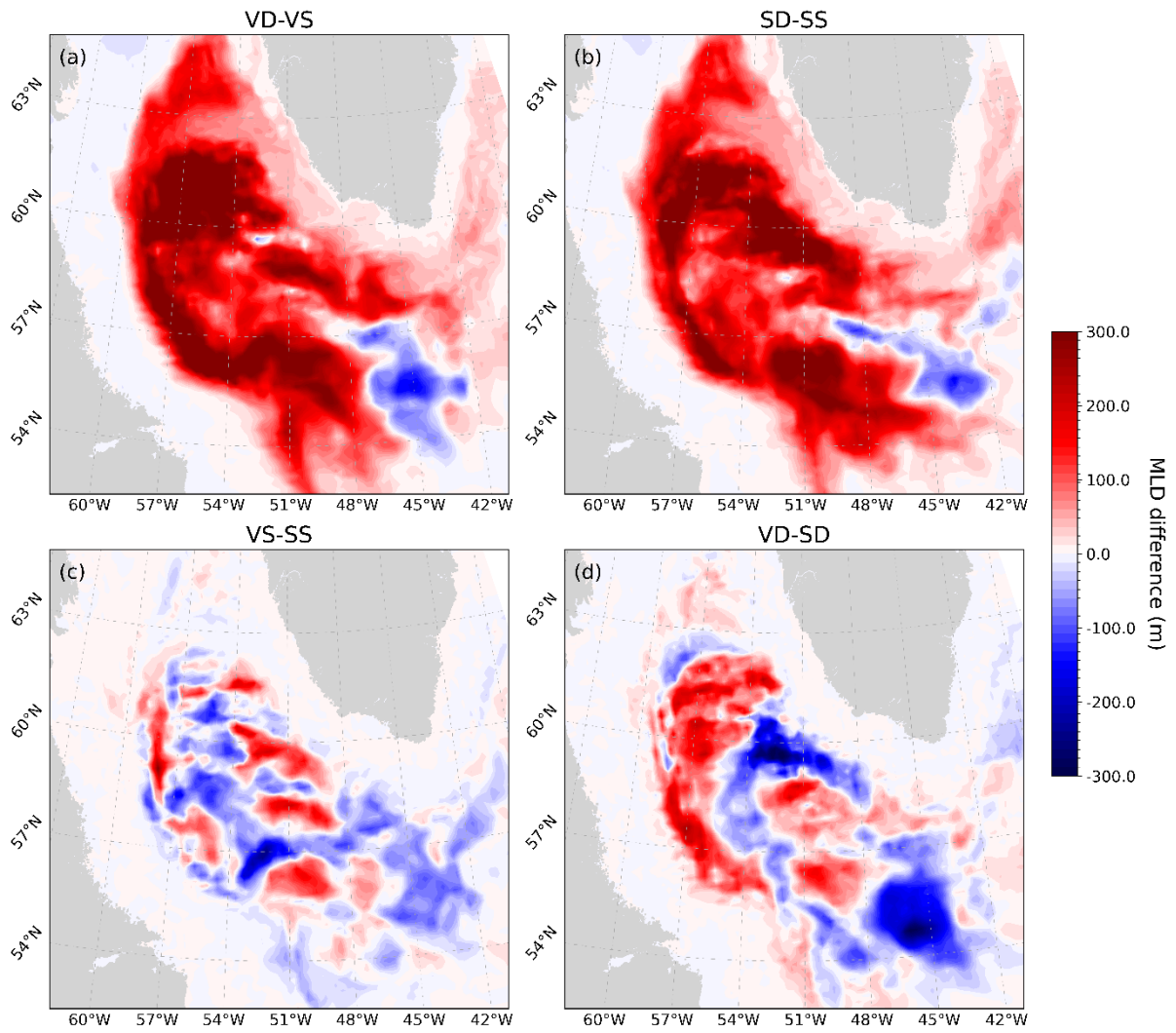


Figure 31: Differences of winter average mixed layer depth between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

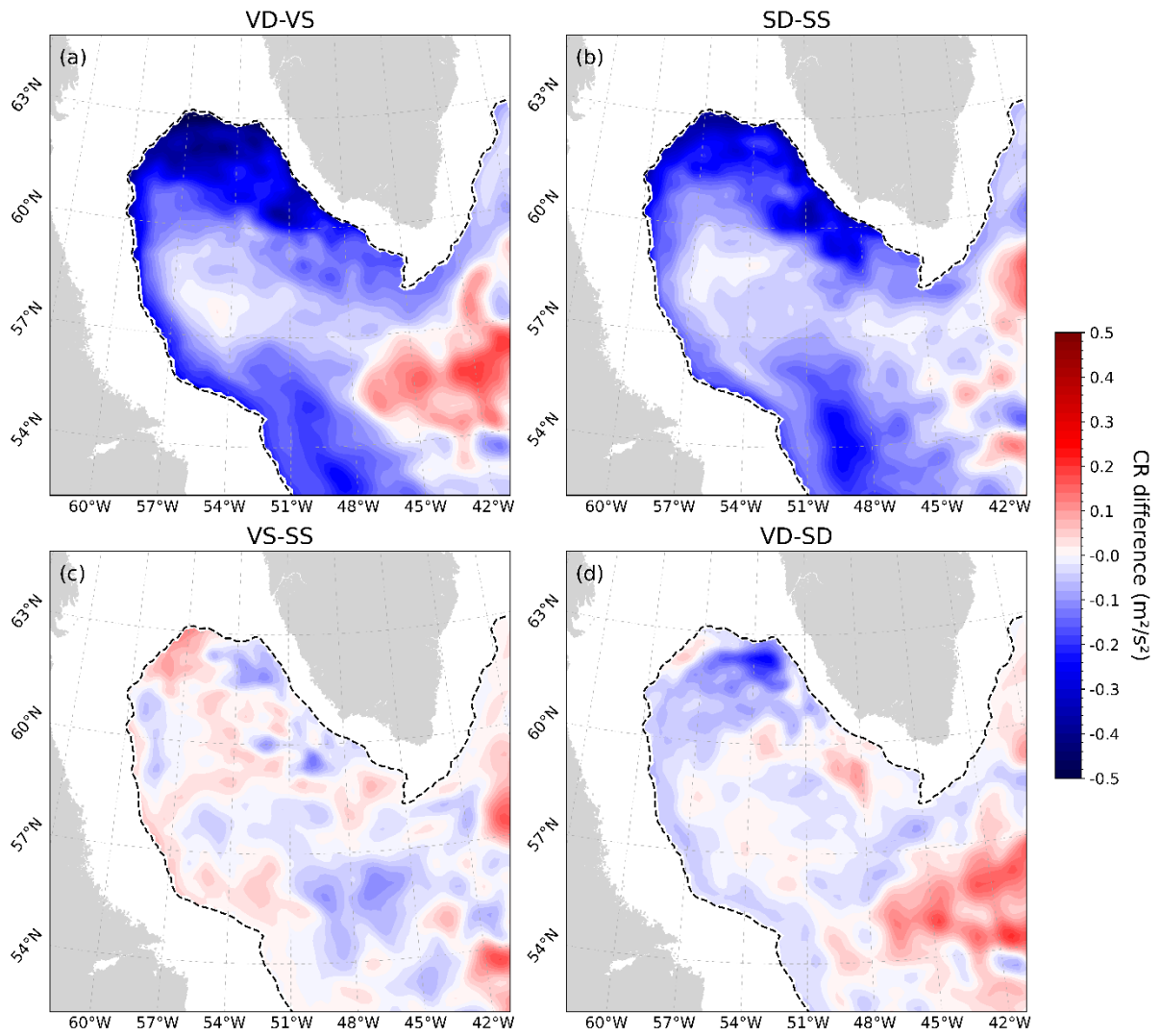


Figure 32: Differences of winter average convective resistance (to a reference depth of 2000 m) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD. Black dashed contour displays the 2000 m isobath.

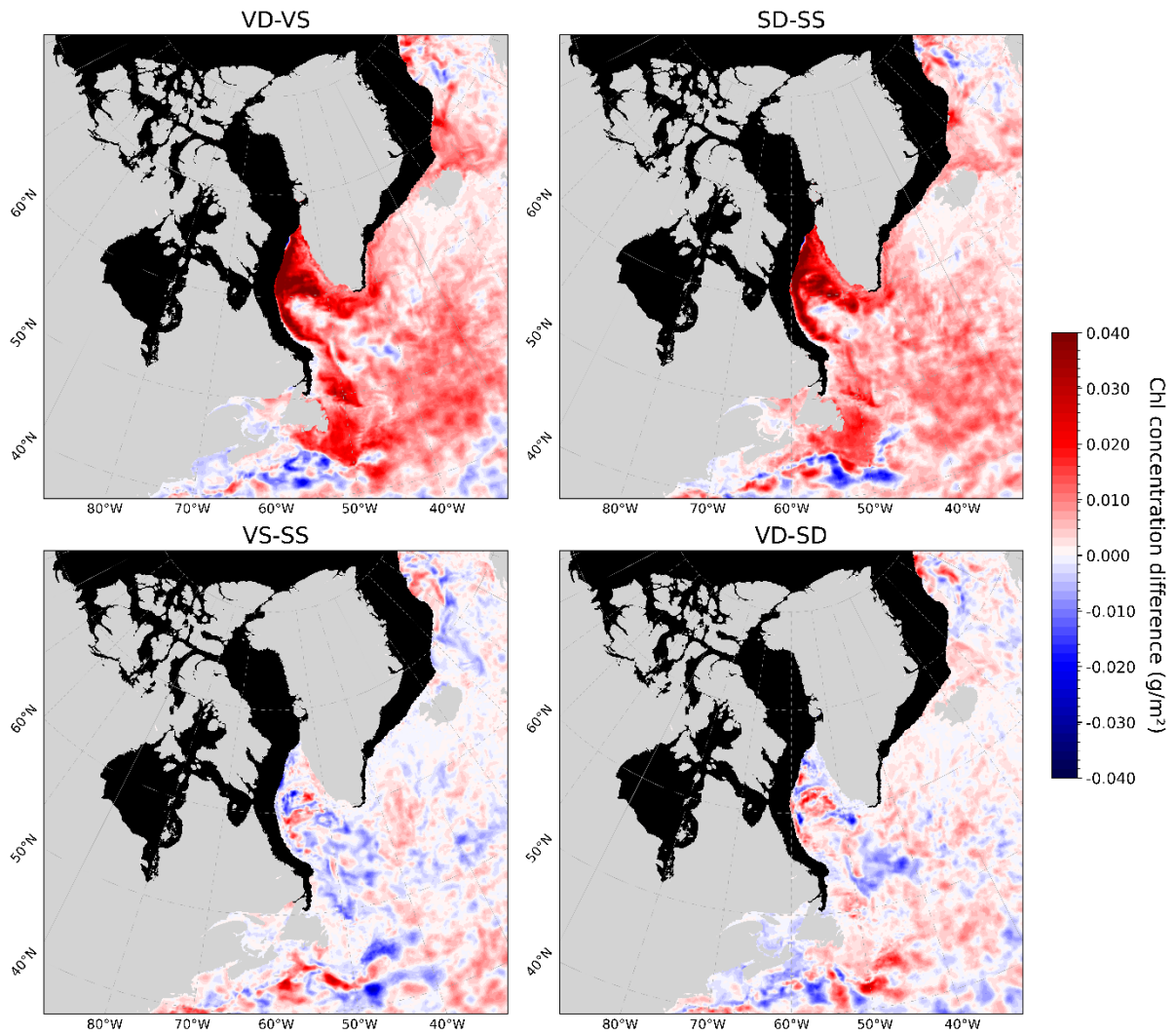


Figure 33: Differences of spring average depth-integrated (surface-bottom) chlorophyll concentration ($\text{g}\cdot\text{m}^{-2}$) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD. Black shading covers the region where spring average sea ice cover > 10%.

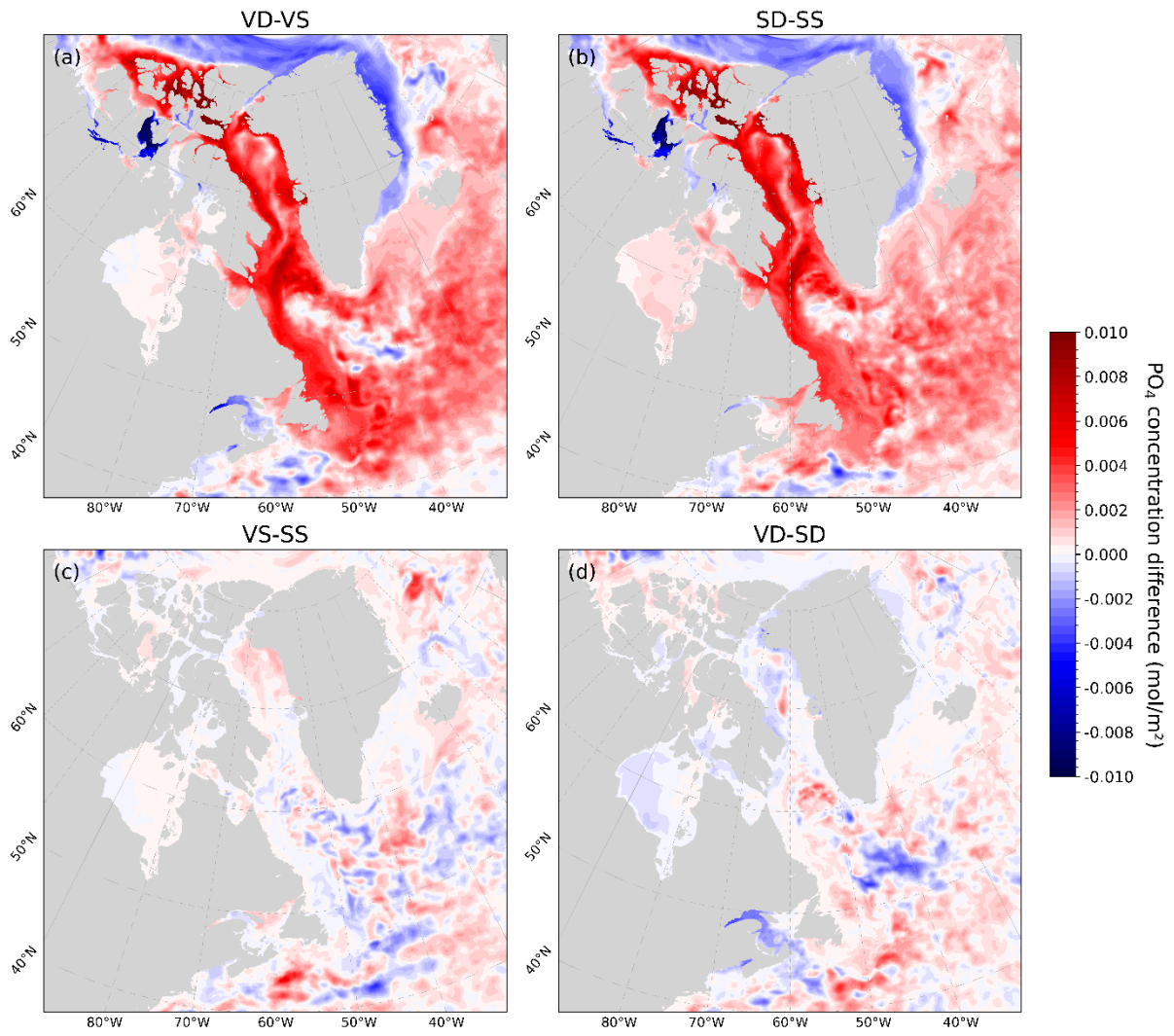


Figure 34: Differences of spring average depth-integrated (0-55m) phosphate concentration (mol.m^{-2}) between (a) VD and VS, (b) SD and SS, (c) VS and SS, and (d) VD and SD.

4. Contributions of Authors

GAW conceived the study, performed the data analysis, and led the writing of the manuscript. JMM designed and carried out the simulations, corrected the model setup, and supervised the project together with GAW, contributing to the interpretation of results and revision of the manuscript. PGM provided supervision during the development of the simulations, contributed to the interpretation of results, and revised the manuscript. All authors contributed to the improvement of the study.