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Key Points:
- Our model experiments support the established view that deep water formation does not play a major role in Baltic Sea bottom water renewal
- We implement a subgrid-scale brine rejection parameterization; its effect is too weak to impact deep water exchange in the northern Baltic
- Hypothesized bottom water ventilation through convection during cold climate phases cannot be explained by reduced air temperatures only

Supporting Information:
- Supporting Information S1

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Revisiting the Role of Convective Deep Water Formation in Northern Baltic Sea Bottom Water Renewal

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Abstract Deep water renewal and ventilation of the Baltic Sea is commonly considered to occur solely via saltwater inflows from the North Sea. However, recent analysis of geophysical and sediment proxy data suggests convective deep water formation during wintertime as a second process that has contributed to the ventilation of the northern and central Baltic Sea bottom waters during cold climate periods. Here, we investigate the role of deep water formation in the northern Baltic Sea in a regional ocean circulation model. Selecting the particularly cold winter 1986/1987 as a reference period, we perform sensitivity experiments in which the atmospheric temperature forcing is changed and/or localized brine rejection on subgrid scales is accounted for through a parameterization. We study the sinking and circulation of water masses via passive tracers introduced into the model. Generally, our model results support the established view that convective deep water formation does not play a major role in Baltic Sea deep water renewal. While a reduction of air temperatures does not qualitatively change the sinking and circulation of water masses in our experiments, brine rejection could potentially lead to localized deep water formation. However, the impact is too weak to possibly change the large-scale deep water exchange between the Baltic proper and the northern Baltic sub-basins. Although being in line with established knowledge, through consideration of the model limitations, our results provide insights on how and under which circumstances convective deep water formation could potentially have occurred during cold climate periods.

Plain Language Summary The transport of oxygen-rich water into the deep parts of the Baltic Sea is of crucial importance for the ecosystem, as under low-oxygen conditions (hypoxia), most organisms cannot survive and so-called dead zones are formed. The ventilation of Baltic Sea bottom waters is commonly considered to occur solely through inflows of dense, oxygenated water from the North Sea. However, recent analysis of sediment cores and geophysical data suggests the contribution of a second process during cold climate periods: convective deep water formation. This process is initiated through surface cooling or an increase of surface water salinity during sea ice formation (brine rejection) in the winter season. Here, we investigate the role of deep water formation in the northern Baltic Sea in a regional ocean circulation model. We test the influence of reduced air temperatures and consideration of strong, localized brine rejection by performing sensitivity experiments. Generally, our model results do not support the hypothesis of convective deep water formation contributing to the renewal of Baltic Sea bottom waters, but are in line with the established view. Nevertheless, our results provide insights on how and under which circumstances convective deep water formation could potentially have occurred during cold climate periods.

1. Introduction

The Baltic Sea (Figure 1) is a semi-enclosed, brackish sea which is connected to the North Sea via a shallow and narrow connection, the Danish straits. Limited water exchange with the North Sea and a positive freshwater budget from river discharge and net precipitation result in strong horizontal and vertical salinity gradients and the establishment of an estuarine-like circulation with outflowing brackish water at the surface and inflowing saltier water close to the seafloor. The permanent and pronounced halocline constrains an exchange between surface and bottom waters and, as a consequence, facilitates frequent occurrence of suboxic or anoxic conditions in the bottom waters. Ventilation and renewal of the deep waters of the central Baltic Sea is commonly considered to occur solely through inflows of dense and oxygenated water from the

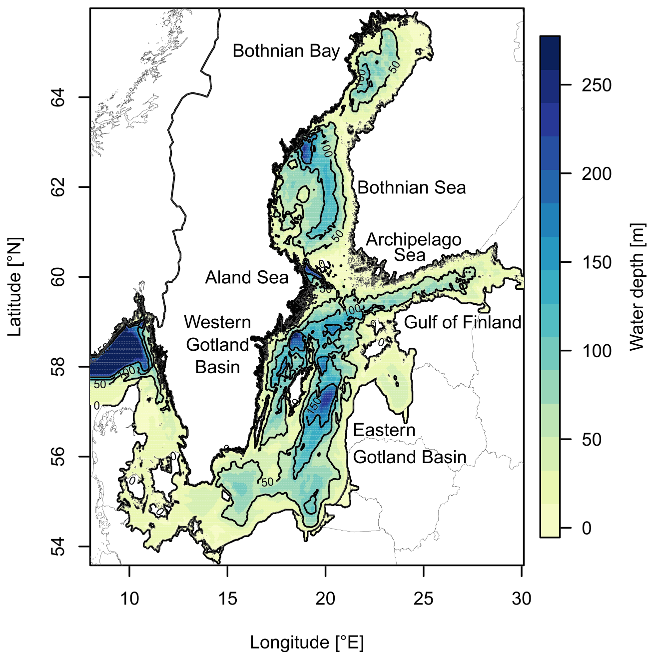

Figure 1. Map and bathymetry of the Baltic Sea. Sub-basin names relevant for this study are indicated.

North Sea (e.g., Meier et al., 2006). These inflows occur episodically in response to specific meteorological conditions (Schinke & Matthäus, 1998). An effective oxygenation of the central Baltic Sea can occur through strong inflow events, so-called major Baltic inflows (Fischer & Matthäus, 1996; Matthäus & Franck, 1992) or sequences of smaller inflows (Neumann et al., 2017). The northern central Baltic Sea is connected to two major sub-basins, namely, the Gulf of Bothnia in the north, which is further divided into the Bothnian Sea and the Bothnian Bay, and the Gulf of Finland in the northeast. The water exchange between the Baltic proper and these sub-basins is commonly considered to be of an estuarine type as well with inflows of surface water from the central Baltic Sea leading to a deep water renewal in the northern sub-basins (e.g., Andrejev et al., 2004; Elken et al., 2003; Marmefelt & Omstedt, 1993; Myrberg & Andrejev, 2006).

In addition to this well-established view, recent findings by Moros et al. (2020) suggest wintertime deep water formation as a second process that has contributed to sediment re-deposition in and bottom water ventilation of the northern and central Baltic Sea during preceding cold phases like the Little Ice Age (LIA). This hypothesis is based on evidence from numerous sediment cores (e.g., irregularities in chronologies based on radiocarbon dating results) and seismoacoustic profiles (examples are shown in the supporting information), indicating that during cold phases, local convection resulted in deep water formation and further produced strong southward-directed bottom currents causing a lateral sediment influx from the northern sub-basins to the northern Baltic proper and Landsort Deep (Moros et al., 2020). Furthermore, measurements of oxygen concentration indicate that bottom water ventilation through wintertime deep water formation might have also occurred during severe winters in the current climate state (see supporting information).

The hypothesis that convection is a key process for the deep water renewal of the northern Baltic Sea is not new and has already been raised by Granqvist (1938). Deep water formation occurs as a consequence of the creation of dense water at the ocean surface, being less buoyant than the water below, and subsequent convection. The necessary density increase or buoyancy loss can be caused either by temperature reduction (i.e., thermal convection) or salinity increase (i.e., haline convection). However, the freezing temperature in the northern Baltic is below the temperature of maximum density (TMD) due to the low salinity. That is, cooling below the TMD will increase buoyancy and temperature sensitivity of buoyancy is low close to the TMD. Therefore, although the salinity in the northern Baltic Sea is low, haline convection should be the major process controlling wintertime convection. A substantial increase of surface salinity can be generated during the sea ice period through the process of brine rejection. During the formation of sea ice, the salt gets rejected from the forming ice crystals and accumulates as highly saline water, called brine. The brine gets trapped within the interstices of the sea ice and is gradually released into the surface water through channels and holes within the ice via different mechanisms (e.g., Notz & Worster, 2009; Untersteiner, 1968). Marmefelt and Omstedt (1993) investigated the potential of thermal and haline convection for deep water renewal in the Gulf of Bothnia based on mean vertical salinity and temperature profiles from observational data. They concluded that neither deep thermal nor deep haline convection is likely to occur frequently and that deep water renewal occurs predominantly through inflows from the Baltic proper. However, they also noted distinct occasions at which deep, bottom-reaching convection could potentially have occurred, namely, in 1979 for thermal convection and in the winter 1986/1987 for haline convection, and pointed out the need for further wintertime observations. Recent ship-based, snapshot observations of sea ice, brine, and water column profiles in the ice-covered Bothnian Bay in March 2017 indicate a ventilation through lateral inflows rather than convection (Neumann et al., 2020), supporting the conclusion of Marmefelt and Omstedt (1993).

The conclusion drawn by Marmefelt and Omstedt (1993) that deep haline convection is unlikely to occur in the Gulf of Bothnia is based on simple budget considerations, assuming that all rejected salt is homogeneously mixed within the surface layer. This approach neglects the heterogeneity of the sea ice cover and the possibility of strong localized convection. Areas with particularly high brine-induced buoyancy fluxes are openings in the sea ice cover, like leads and polynyas, as their exposure to the cold atmosphere allows for fast ice growth. Although making up only a small fraction of the total ice cover, leads account for a major fraction of the salt input into the ocean associated with brine rejection (e.g., Maykut, 1978). The convection process in leads has been extensively studied by means of observations (e.g., Morison & McPhee, 1998; Morison et al., 1992; Muench et al., 1995) and numerical modeling (e.g., Kozo, 1983; Smith & Morison, 1993, 1998; Smith et al., 2002). It is associated with the formation of convective plumes which, in case of little ice motion, sink to the bottom of the mixed layer, only partially mixing with ambient water, and then spread horizontally away from the lead center. In case of low stratification, the plumes might even penetrate through the pycnocline and sink to the bottom of the sea. Taking this localized brine rejection into account, it seems worthwhile to reconsider the results of Marmefelt and Omstedt (1993).

In this study, the role of deep water formation in the northern Baltic Sea is revisited in a model approach using a regional configuration of an ocean circulation model. The guiding question is whether a ventilation of bottom waters in the northern sub-basins and the northern central Baltic Sea due to wintertime deep water formation is present in the model. It is further subdivided into the following research questions, each addressed by different model experiments.

1. Is wintertime deep water formation occurring in the northern Baltic Sea sub-basins?
2. What is the possible effect of strong, localized brine rejection in the Baltic Sea?
3. How does the situation change in a colder climate state?

In order to represent localized brine rejection and convection on scales smaller than the model grid size, a subgrid-scale parameterization of brine rejection (Nguyen et al., 2009) is implemented into the ocean model and applied to the Baltic Sea for the first time.

2. Methods

2.1. Model Description

The numerical model used in this study is based on the circulation model Modular Ocean Model (MOM) 5.1 (Griffies, 2004) and has been adapted to the Baltic Sea with an open boundary condition to the North Sea. Vertical subgrid mixing is calculated with the non-local K-profile scheme (Large et al., 1994). Horizontal viscosity is calculated from the Smagorinsky (1963) scheme. The MOM model is complemented with a sea ice model that accounts for the formation of sea ice and the ice drift (Winton, 2000). Brine rejection is parameterized by a freshwater flux during ice formation considering a bulk ice salinity. The horizontal resolution of the model grid is one nautical mile, while vertically, the model is resolved into 152 layers, starting with 0.5-m layer thickness at the surface and gradually increasing with depth up to a layer thickness of 2 m. The model was integrated from 1950 until 2018. Meteorological forcing was derived from the coastDat2 data set (Geyer, 2014) and the runoff is based on HELCOM (www.helcom.fi) data. For this study, we select the years 1986 and 1987, representing a particularly cold winter with the largest ice extent in the current climate period (see supporting information), and perform a sensitivity study (see section 2.4).

2.2. Subgrid-Scale Brine Rejection Parameterization

Due to openings and different ice thicknesses, the surface boundary conditions in the presence of a sea ice cover are very heterogeneous. In particular, brine rejection in leads causes strong buoyancy losses on scales of only a few to a few hundred meters, that are generally not resolved in ocean circulation models. Without any special treatment, the surface buoyancy flux is averaged horizontally over an entire grid cell, suppressing the potential of localized deep mixing. The consequences are too shallow penetration depths of the convective plumes and too strong mixing and homogenization above the penetration depth, as shown for instance by Losch et al. (2006) and Jin et al. (2012).

In order to represent the subgrid-scale convection associated with brine rejection in leads, the simple concept of brine rejection parameterizations (BRPs) has been introduced (Duffy & Caldeira, 1997; Duffy et al., 1999; Nguyen et al., 2009 and follow-up studies). The basic idea is to distribute the rejected salt vertically within the mixed layer, mimicking the spreading through a convective plume, instead of placing it in the top layer and treating it via a vertical mixing scheme. This is of course an oversimplified representation of subgrid-scale convection, neglecting for instance the mixing of heat and other tracers, but it provides a first,

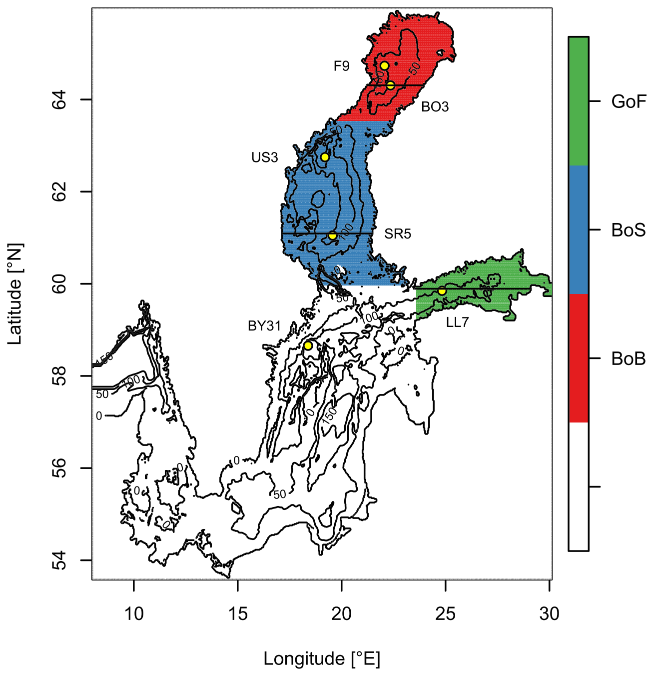

Figure 2. Mask showing the release areas of the three passive tracers, marking waters of the three northern sub-basins: the Bothnian Bay (BoB, red), the Bothnian Sea (BoS, blue), and the Gulf of Finland (GoF, green). Further shown are the stations BO3, SR5, LL7, and BY31 (marked as yellow dots) as well as three zonal transects within the sub-basins (black lines), used for evaluation purpose.

simple approach to overcome the problem of horizontally averaged buoyancy fluxes. Here, the formulation of Nguyen et al. (2009), which is based on knowledge of the convection process in leads from observations as well as laboratory and numerical experiments, is followed. The rejected salt is distributed vertically according to a power law distribution, that is,

$$s(z) = \begin{cases} Az^n & \text{if } z \leq D_{sp} \\ 0 & \text{if } z > D_{sp} \end{cases}, \tag{1}$$

where z is the depth, defined as being positive below sea level for simplicity reasons, $s(z)$ is the salt distribution function, n is the distribution power, and D_{sp} is the salt plume depth. The constant A is determined by the total amount of rejected salt S_0:

$$\int_0^{D_{sp}} s(z)dz = \int_0^{D_{sp}} Az^n dz = \frac{A}{n+1} D_{sp}^{n+1} \stackrel{!}{=} S_0 \tag{2}$$

$$\Rightarrow A = \frac{n+1}{D_{sp}^{n+1}} S_0. \tag{3}$$

The distribution power n and the salt plume depth D_{sp} are adjustable parameters. For their specific application area, the Arctic Ocean, Nguyen et al. (2009) found that setting $n = 5$ and D_{sp} to a depth corresponding to a density gradient of $d\rho/dz = 0.002$ kg/m^4 yields the lowest model-data misfits. In this study, the value of $n = 5$ for the distribution power is adopted from Nguyen et al. (2009) and two different methods for determining the salt plume depth D_{sp} are used (see section 2.4).

Note that, apart from the concept of BRPs, other approaches for taking into account subgrid-scale convection exist, for instance two- or multi-column mixing schemes (e.g., Barthélemy et al., 2016; Jin et al., 2015).

2.3. Passive Tracers

One main aim of this study is to track the pathways of water masses originating in the northern Baltic Sea sub-basins in order to verify if wintertime deep water formation is happening and whether these water masses are entering the central Baltic Sea.

Therefore, three passive tracers, marking water of the Bothnian Bay (BoB), the Bothnian Sea (BoS, including also the Åland and Archipelago Seas), and the Gulf of Finland (GoF) as shown in Figure 2, are introduced into the model. Depending on the addressed research question of the respective experiment, two different kinds of passive tracers, which differ in the method of initialization, are used in this study.

2.3.1. Surface Tracer

In order to answer the first research question, whether wintertime deep water formation is occurring in the northern Baltic Sea sub-basins, the passive tracers are initialized as so-called "surface tracers." Their concentration is set to 1 m^{-3} in the surface cells within the initialization area (shown in Figure 2) at each time step during the period of November to March. Afterwards, they are advected freely. This enables to qualitatively follow the pathways of the winter surface water masses of the different basins and to observe possible deep water formation through either haline or thermal convection.

2.3.2. Brine Tracer

In order to answer the second research question, which is explicitly targeting the potential of localized brine rejection for deep water renewal in the northern Baltic Sea, the passive tracers are initialized as so-called "brine tracers." Their concentration is set to the actual salinity caused by salt rejected during ice formation. The tracer concentration has the unit g/kg and is a quantitative measure of brine rejection. The advantage of this method is that the subgrid-scale BRP, introduced in section 2.2, can be applied not only to the salinity itself but also to the brine tracer concentration. This allows us to follow the pathways of water masses at

Table 1
List of Sensitivity Experiments

Experiment	Atmospheric forcing	Brine rejection parameterization
ref	Winter 1986/1987	No BRP
BRP bottom	Winter 1986/1987	D_{sp} = bottom
BRP MLD	Winter 1986/1987	D_{sp} = mixed layer depth
Cold Climate	−2 K air temperature	No BRP
BRP bottom Cold Climate	−2 K air temperature	D_{sp} = bottom
BRP MLD Cold Climate	−2 K air temperature	D_{sp} = mixed layer depth

depths corresponding to the sinking depths of the convective brine plumes, assuming that they behave as prescribed by the BRP.

2.4. Sensitivity Experiments

In total, six experiments are conducted. An overview of the experiments is given in Table 1.

The first experiment is a reference experiment ("ref") with the atmospheric forcing of the years 1986/1987 and without a BRP. On the one hand, it aims to answer the first research question, namely, whether there is wintertime deep water formation (either through thermal or haline convection) occurring in the northern Baltic Sea. On the other hand, it serves as a reference experiment for the following sensitivity experiments.

The next two experiments address the second research question, aiming to analyze the effect of strong, localized brine rejection. Therefore, the BRP (section 2.2) is switched on. The value of $n = 5$ for the distribution power is adopted from Nguyen et al. (2009). Two experiments with different salt plume depths D_{sp}, namely, the bottom ("BRP bottom") and a variable mixed layer depth ("BRP MLD"), are performed. The mixed layer depth is determined in the MOM model as the depth at which a critical buoyancy difference of 0.0003 m/s^2 with respect to the surface is reached.

In order to answer the third research question of how the situation is changed in a colder climate state, the three experiments are repeated with a different atmospheric forcing, in which the 2-m air temperature is reduced by $\Delta T = 2\,\text{K}$ constantly in space and time ("Cold Climate" experiments). With the change in air temperature, the specific humidity needs to be adjusted as well. Assuming that the relative humidity is unchanged, the specific humidity decreases with temperature according to the Clausius-Clapeyron relation, that is,

$$s_2 = s_1 \cdot \exp\left(\frac{\Delta H_{\text{vap}}}{R}\left(\frac{1}{T_1} - \frac{1}{T_2}\right)\right), \tag{4}$$

where s_1 and T_1 are the original specific humidity and temperature, s_2 is the new specific humidity corresponding to the new temperature T_2, ΔH_{vap} is the enthalpy/heat of vaporization of water, and R is the universal gas constant. The temperature reduction by 2 K corresponds to the mean sea surface temperature (SST) difference between the LIA period and the modern warm period found by Kabel et al. (2012) based on a reconstruction from sediment proxy data. In that sense, it can be regarded as a change to LIA temperature forcing, assuming that SST and air temperature behave similarly. However, this might not be the case in the presence of an ice cover and severe winters during the LIA could have been much colder. Furthermore, in the Cold Climate experiments, we only account for differences in the temperature forcing, but not for other potential differences between different climate states, such as wind forcing or river runoff impacting the hydrographic conditions.

The "brine tracers" are introduced in all of the experiments, while the "surface tracers" are only used in the two experiments without BRP ("ref" and "Cold Climate").

3. Results

3.1. Model Validation

A reasonable model performance is the basis for confidence in the obtained model results. In the supporting information, we show a detailed validation of the simulation results by comparing sea ice, temperature, and salinity data with observations. Overall, the observations are well reproduced by the model, particularly for sea ice and temperature, but salinity biases lead to an overestimation of the stratification in the central Baltic Sea and Bothnian Sea.

3.2. Wintertime Deep Water Formation in the Northern Basins

3.2.1. Convective Sinking

The first question to answer is whether wintertime deep water formation through convection is occurring in the northern Baltic Sea sub-basins within the chosen model framework during the simulated winter

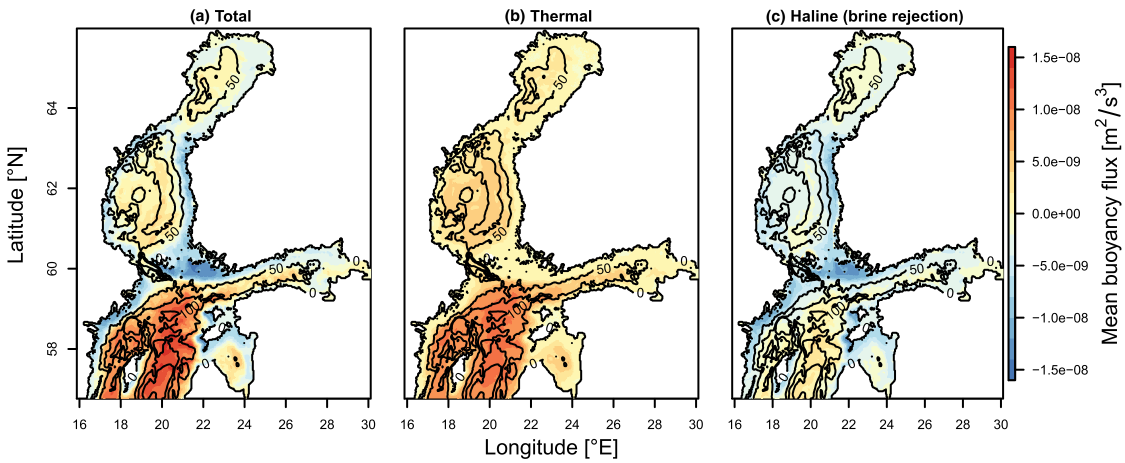

Figure 3. Winter (JFM) mean surface buoyancy flux in the reference experiment and its different contributions. (a) Total buoyancy flux, (b) thermal contribution, and (c) haline contribution from brine rejection.

1986/1987. Convection is generally forced by a loss of buoyancy at the water surface, which can either be temperature-driven (thermal convection) or salinity-driven (haline convection). A measure for the forcing of convection is the surface buoyancy flux, which can be expressed in terms of heat and freshwater fluxes as

$$B_0 = \frac{g\alpha_T}{\rho_0 c_p} Q_{\mathrm{H}} + g\beta_S S Q_{\mathrm{FW}}, \tag{5}$$

where g is the gravitational acceleration, ρ_0 is a reference density, c_p is the heat capacity of seawater, S is the surface salinity, α_T and β_S are the thermal and haline expansion coefficients, and Q_{H} and Q_{FW} are the net heat and freshwater fluxes. Defining the fluxes as being directed from the atmosphere to the ocean, a positive buoyancy flux stabilizes the water column, whereas a negative buoyancy flux gives rise to convection. Figure 3 shows (a) the winter mean surface buoyancy flux, (b) the thermal contribution, and (c) the haline contribution from brine rejection (which is about one order of magnitude larger than the contribution from precipitation and evaporation) in the reference experiment. Here and in the following, "winter mean" denotes the average over the months from January to March (JFM), which are the months of major ice formation. Due to the TMD being higher than the freezing temperature in the brackish Baltic Sea water, the thermal expansion coefficient is negative when the SST sinks below the TMD and, despite the negative heat flux, the average thermal contribution in winter is positive over the whole northern Baltic Sea. The negative, haline contribution from brine rejection is, however, dominant in the coastal areas of the northern Baltic Sea, where most of the sea ice formation takes place, and therefore, the total buoyancy flux is negative in large parts of the northern Baltic Sea. The forcing would therefore allow for potential wintertime deep water through haline convection in coastal areas of the northern Baltic Sea.

Apart from the surface buoyancy forcing, the stratification conditions in the water column determine whether deep-reaching convection can occur. The introduction of the passive surface tracers allows for an analysis of the sinking of winter surface water in the individual sub-basins. Figure 4 shows a map of the sinking depth, defined as the depth at which the winter mean surface tracer concentration has decreased to 0.1 m^{-3}, that is, 10% of the initial concentration at the surface, in the reference experiment. In most parts of the northern Baltic Sea, the sinking depth does not exceed the model halocline depth of around 50–60 m. Exceptions are the Åland Sea and the Bothnian Bay, where sinking depths of locally up to 140 and 110 m, respectively, are found. While for the Bothnian Bay, this means that the tracer is reaching the bottom of the basin, the largest depths of the Åland Sea (up to 300 m) are not reached by convection. Note that here the sum of all three passive tracers is used to determine the sinking depth and tracer inflowing from other areas can increase the sinking depth from pure vertical convection, for example within the inflow channel from the Åland Sea into the Bothnian Sea.

A more detailed view of the convective sinking in the individual basins can be gained from analyzing vertical profiles of surface tracer concentration. The solid lines in Figure 5 show vertical profiles of winter mean

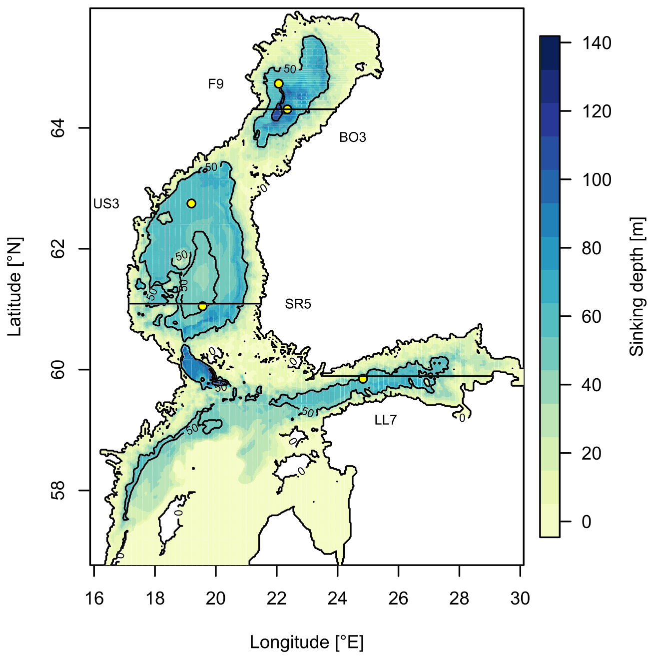

Figure 4. Sinking depth, defined as the depth at which the winter (JFM) mean surface tracer concentration has decreased to 0.1 m^{-3}, in the reference experiment.

surface tracer concentration at stations in the Bothnian Bay (F9 and BO3), Bothnian Sea (SR5 and US3), and Gulf of Finland (LL7) in the reference experiment. Furthermore, for each of the basins, a vertical cut of winter mean surface tracer concentration is shown along a zonal transect through the basin (station and transect locations are marked in Figures 2 and 4). As can be seen from these exemplary profiles, the sinking behavior of surface tracer is different in each of the basins. In the Gulf of Finland, which is the most shallow of the three sub-basins (average depth of 38 m), the surface tracer reaches very close to the bottom. The concentration gradually decreases below a depth of around 40 m, but as the bottom in most parts of the basin is not much deeper as this mixed layer depth, measurable amounts of surface tracer can be found throughout the whole water column. In the Bothnian Sea, the surface tracer is not sinking below the halocline depth of around 50 m. The concentration gradient in the halocline layer varies at different locations within the Bothnian Sea, but nowhere the convection is sufficiently strong to bring surface water down to the bottom of the basin. The narrow channel with higher sinking depths, visible in Figure 4, is a result of tracer transport from the Åland Sea and not of local convection.

In contrast to that, significant amounts of surface tracer concentration reach the basin bottom in parts of the Bothnian Bay. As can be seen from the sinking depths in Figure 4 and the profiles in Figure 5, the primarily affected region is in the more southern part of the Bothnian Bay around station BO3. The BO3 tracer profile indicates that the deep water tracer input is not a result of local convection, which would go along with a homogeneous water column, but rather the result of a lateral intrusion or inflow. The tracer concentrations along the zonal transect through the Bothnian Bay (Figure 5) reveal that the sinking of tracer occurs along the eastern coast. Two processes could potentially be responsible for the creation of a dense water mass inducing the convective sinking: mixing of Bothnian Bay surface water with (i) inflowing, more saline Bothnian Sea surface water or (ii) brine released during the formation of sea ice along the shallow coastal area. It is not possible to disentangle these two processes only from BoB tracer concentrations. However, we also find surface tracer from the Bothnian Sea at similar depths as the surface tracer from the Bothnian Bay and find qualitatively similar tracer sinking in an additional experiment where the process of brine rejection is turned off by setting the ice-ocean freshwater flux to zero (not shown). Therefore, we conclude that the dominant process for the simulated deep water formation in the Bothnian Bay is advection from the Bothnian Sea. Nevertheless, brine rejection might enhance the convective sinking.

3.2.2. Tracer Inflow Into the Central Baltic Sea

The geological findings described in Moros et al. (2020) suggest that during cold periods, not only local wintertime convection is taking place, but the deep water formed in the northern sub-basins is further transported into the deep parts of the northern central Baltic Sea. By evaluating the temporal evolution of tracer concentrations, we can follow the pathways of the winter surface water of the three sub-basins. This is shown in Figure 6 as snapshots of tracer concentrations at the bottom of the Baltic Sea in the reference experiment. In agreement to other model studies and observations (e.g., Andrejev et al., 2004; Myrberg & Andrejev, 2006), all sub-basins show a mean cyclonic (i.e., anti-clockwise) circulation. The BoB tracer leaves the Bothnian Bay southward along the western coast in the surface layer. The BoS tracer exits the Bothnian Sea partly northward, entering the Bothnian Bay bottom layer along the eastern coast, and partly southward through the Åland Sea into the surface layer of the Baltic proper. The GoF tracer escapes the Gulf of Finland at the northern part of its connection to the Baltic proper, partly entering the Bothnian Sea through the Archipelago Sea and partly flowing into the central Baltic Sea.

To evaluate the inflow of winter surface water from the northern sub-basins into the Baltic proper, we choose to analyze the tracer concentrations at Landsort Deep (station BY31). Figure 7 shows the time series of 30-day running mean tracer concentration of BoS and GoF surface tracer at station BY31. The time series of BoB tracer is not shown as the simulation period is too short for significant amounts of BoB tracer to arrive

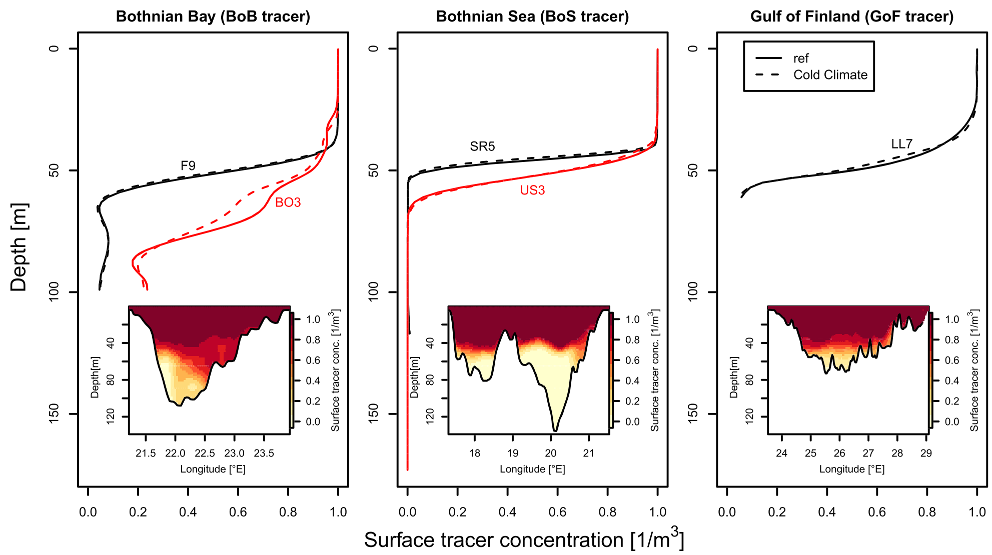

Figure 5. Winter (JFM) mean vertical profiles of the three passive surface tracers at selected stations within the respective basin of release (from left to right: Bothnian Bay, Bothnian Sea, and Gulf of Finland) in the reference experiment (solid lines) and the Cold Climate experiment (dashed lines). The insets show winter mean surface tracer concentrations in a transect across each of the basins for the reference experiment.

at Landsort Deep. The travel distances from the Bothnian Sea and the Gulf of Finland are shorter. The tracers from these basins arrive at Landsort Deep from January on, but the highest concentrations are found in the summer months from June on. Both tracers are restricted to the surface layer above the halocline at around 60 m. BoS tracer concentrations are higher than GoF tracer concentrations. In winter and spring, the tracer concentrations are vertically relatively homogeneous due to a well-mixed surface layer; in summer, the surface layer is more stratified. Concluding, there is no inflow of deep water formed during winter in the northern sub-basins into the deep parts of the central Baltic Sea in the reference experiment.

3.3. Effect of Localized Brine Rejection

In the reference experiment, bottom-reaching deep water formation is occurring only in parts of the Bothnian Bay. It might, however, be possible that localized brine rejection on scales that are not resolved by the model (e.g., in sea ice leads) actually leads to stronger deep water formation. In order to test the potential effect of such localized brine rejection, we conducted two sensitivity experiments with a subgrid-scale BRP. In the "BRP bottom" experiment, all salt rejected during ice formation is mixed down to the bottom, which sets an upper limit for the potential effect of localized brine rejection. In the "BRP MLD" experiment, the rejected salt is mixed down to some mixed layer depth as it is done for instance in Nguyen et al. (2009) for the Arctic Ocean. Although the reasoning for introducing the BRP into the model is based on physical grounds, it is not clear whether its application is justified and leads to more realistic model results in the present case. However, the goal here is not necessarily to improve model results but to test how sensitive the model results react to a vertical redistribution of rejected salt. We here compare brine tracer concentrations, which correspond to the salinity of the rejected salt and are subject to the vertical distribution of the BRP, allowing to follow the pathways of water masses at the prescribed salt plume depth.

The effect of the BRP can be seen from Figure 8, showing the mean BoS brine tracer concentration in March along the zonal transect in the Bothnian Sea in the reference experiment and the two BRP experiments. While in the reference experiment, the tracer is vertically homogeneously distributed within the mixed layer, in the BRP bottom experiment, the tracer concentration is gradually increasing with depth and is maximal at the bottom. The differences between the reference experiment and the BRP MLD experiment are less obvious. In both cases, the tracer spreads only within the surface mixed layer, but in the BRP MLD experiment, stronger vertical gradients are found, and at the eastern coast, some tracer sinks below the mixed layer depth. The weakness of these differences is due to a rather conservative choice of MLD criterion, meaning

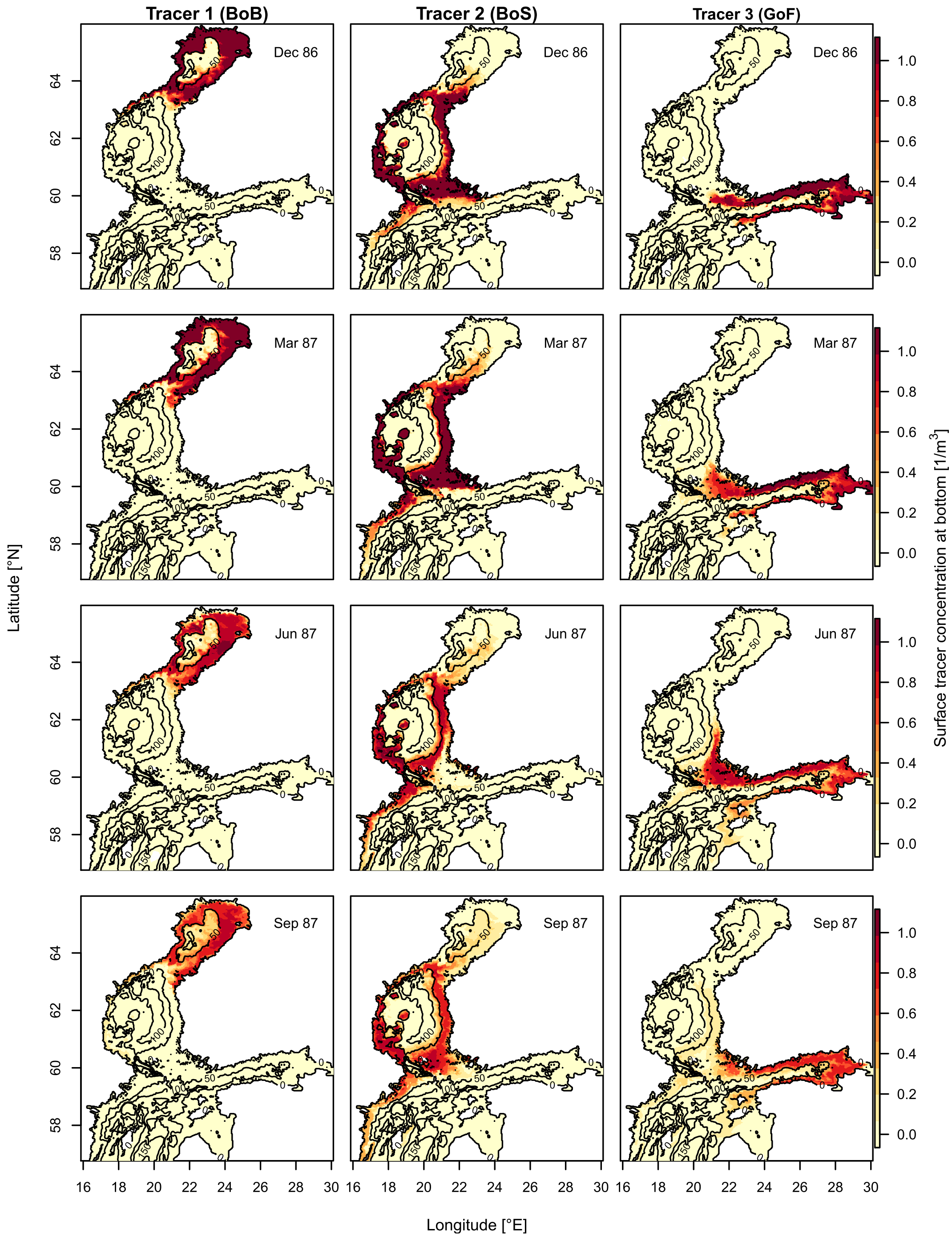

Figure 6. Snapshots of the tracer concentrations of the three passive surface tracers (left column: BoB tracer, middle column: BoS Sea tracer, right column: GoF tracer) at the bottom of the Baltic Sea at the end of the indicated month in the reference experiment.

that the tracer is only mixed down to the upper boundary of the MLD and is subject to surface mixing in a similar manner as in the reference experiment.

To answer the question whether localized brine rejection in sea ice leads might be able to form deep water within the northern sub-basins which then enters the deeper parts of the central Baltic Sea, we analyze the brine tracer concentration at Landsort Deep. The solid lines in Figure 9a show the mean profiles of brine

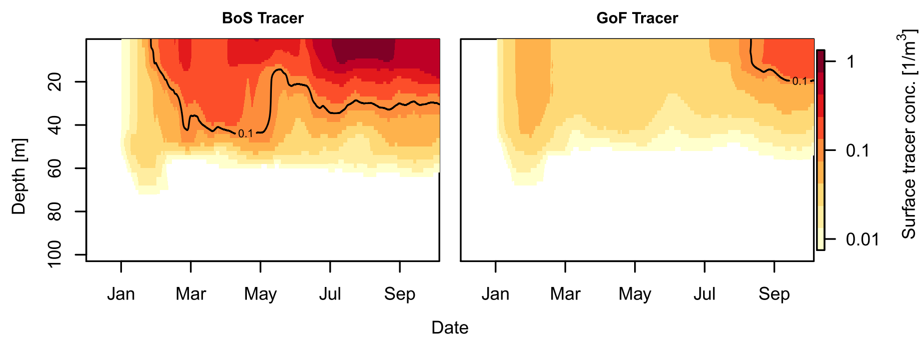

Figure 7. Time series of 30-day running mean tracer concentration of BoS and GoF surface tracers at Landsort Deep (station BY31) in the reference experiment. The BoB tracer is not shown here as no significant amounts of tracer (concentrations higher than 0.01 m^{-3}) reach Landsort Deep within the time scale of the simulation period. Note that a logarithmic scale is chosen for better visualization.

tracer concentration (sum of all three tracers) at station BY31 averaged over the period from June to September (i.e., the period of strongest tracer inflow) in the reference and the two BRP experiments. In all experiments, the highest tracer concentrations are found near the surface. While the concentration profile in the BRP MLD experiment is hardly changed compared to the reference experiment, the surface inflow of tracer concentration is reduced by 40% in the BRP bottom experiment. This is expected as, due to the downward distribution of brine tracer, the tracer concentration in the surface water transported into the Baltic proper is reduced. An interesting feature appearing in the BRP bottom experiment is the increase of tracer concentration in a depth range of 60–90 m, which is shown in the inset of Figure 9a. A closer look at the contributions from the individual tracers, shown in Figure 9b, reveals that the origin of this deeper intrusion of tracer-marked water is the Gulf of Finland. From the tracer concentrations alone, it is not possible to distinguish whether the inflow in this greater depth range is a unique feature of the BRP bottom experiment as a result of the formation of more dense bottom water or if it is also present in the other experiments and only becomes visible by the marking of bottom water with tracer. However, a comparison of the temporal behavior of the inflow and the prevailing wind forcing (see supporting information) suggests that the inflow is most likely caused by a wind-driven reversal of estuarine circulation in the Gulf of Finland, which is possible if strong southwesterly winds with a wind stress exceeding 0.02–0.04 N/m^2 persist (Elken et al., 2003). One could, however, claim that even if the formation of denser bottom water through the BRP is not causing the inflow, surface water from the Gulf of Finland is transported into intermediate depths of the central Baltic Sea, assuming that it is brought down to the bottom beforehand by strong localized brine rejection.

Despite the appearance of the additional deeper inflow of bottom water from the Gulf of Finland in the BRP bottom experiment, a contribution to the Landsort Deep bottom waters from the northern sub-basins, as hypothesized by Moros et al. (2020), is not found. For this to be possible, two requirements must be met: first,

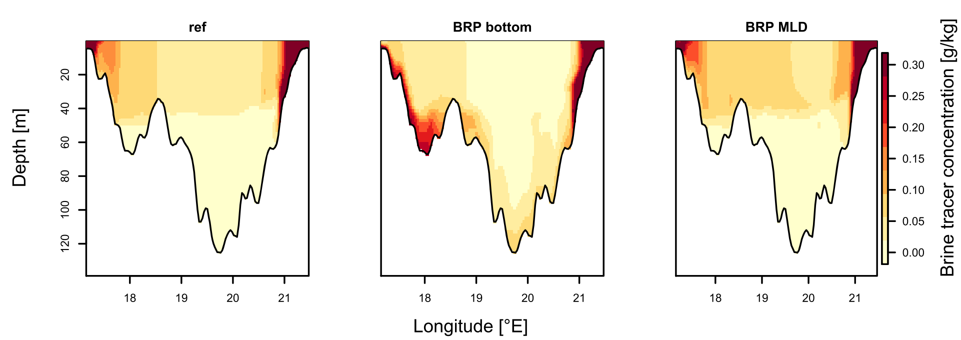

Figure 8. BoS brine tracer concentration in the Bothnian Sea transect averaged over the month of March (i.e., toward the end of the freezing period) in the reference experiment and the BRP experiments with the bottom and MLD as salt plume depths.

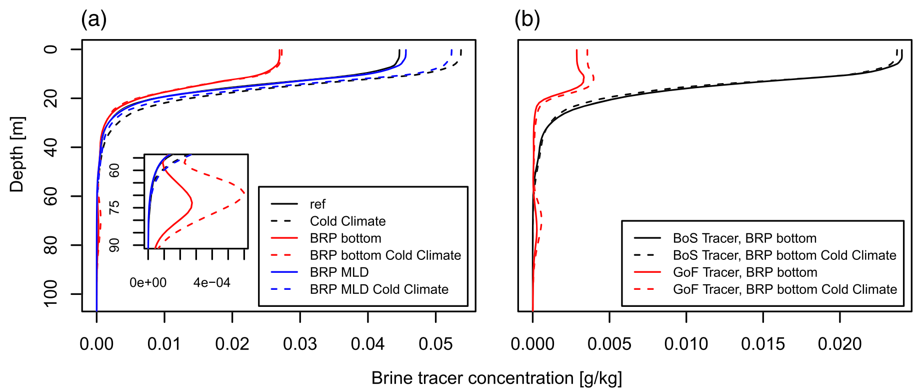

Figure 9. Vertical profiles of mean brine tracer concentration (averaging period: June–September) at Landsort Deep (station BY31). (a) Sum of all three tracers in all model experiments. (b) Contribution of BoS and GoF tracer in BRP bottom experiments.

a water mass with a density high enough to be layered into the central Baltic Sea bottom waters must be formed in the sub-basins, and second, the circulation conditions must be such that this water mass is transported into the central Baltic Sea. Even though in the model, no inflow of bottom water from the Gulf of Bothnia into the central Baltic Sea is occurring, we can test whether the increase in bottom water salinity in the BRP bottom experiment would in principle be sufficient to form a water mass dense enough to contribute to Landsort Deep sub-halocline waters. Figure 10 shows the distribution of winter mean bottom salinity differences between the BRP bottom and the reference experiment in the grid cells of the Bothnian Bay, Bothnian Sea, and Gulf of Finland. On average, the winter bottom salinity is increased by 0.088 g/kg in the Bothnian Bay, 0.083 g/kg in the Bothnian Sea, and 0.054 g/kg in the Gulf of Finland due to the BRP. Locally and on shorter (daily instead of seasonal) time scales, the increase of bottom salinity can be much higher with up to 3 g/kg. However, this is only the case in shallow coastal areas, such as the Archipelago Sea. Assuming that a water mass which could potentially contribute to Landsort Deep bottom water ventilation would require a salinity of at least 9 g/kg, in both reference and BRP bottom experiment, only the Gulf of Finland bottom water is eligible. Considering the salinity bias of the model at Landsort Deep (see supporting information), the inflow of Gulf of Finland bottom water (Figure 9) could occur at greater depths in reality than it is the case in the model. Nevertheless, the process of localized brine rejection itself is too weak to form a dense water mass in the northern sub-basins which could potentially ventilate the northern central Baltic Sea.

3.4. Cold Climate Conditions

In order to answer the question whether the situation changes in a colder climate state, we tested the sensitivity of the results to the atmospheric forcing by repeating the previous experiments with air temperatures

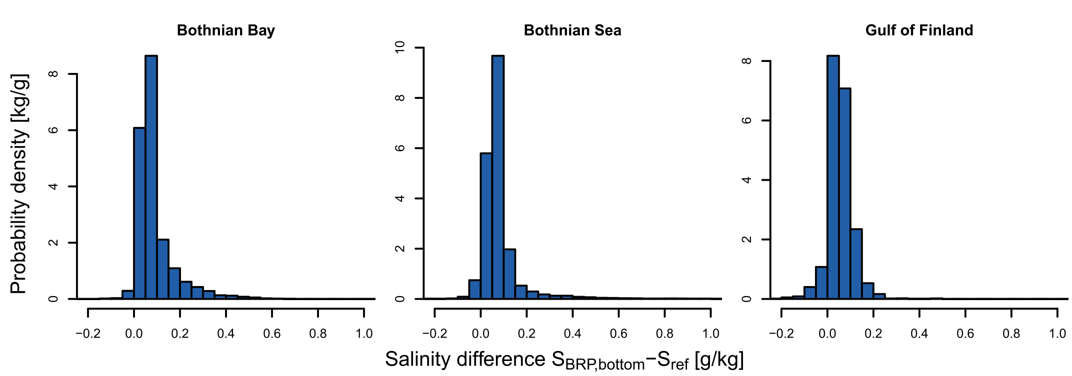

Figure 10. Distribution of bottom salinity differences in winter (JFM mean) between the BRP bottom and the reference experiment in the grid cells of the Bothnian Bay, Bothnian Sea, and Gulf of Finland sub-basins.

reduced by 2 K. The temperature forcing modifies the surface buoyancy flux due to additional cooling of the surface water as well as increased ice growth and associated brine rejection.

The Cold Climate experiment is equivalent to the reference experiment except for the atmospheric forcing. From comparing the sinking of surface tracer between these two experiments, we can draw conclusions on possible changes in wintertime deep water formation in the northern sub-basins. The dashed lines in Figure 5 show vertical profiles of winter mean surface tracer concentration at selected stations in each of the sub-basins, similar to the solid lines for the reference experiment. In the Gulf of Finland and the Bothnian Sea as well as at station F9 in the Bothnian Bay, the profiles are almost unchanged with respect to the reference experiment. At station BO3, which was the only station with ground-reaching deep water formation in the reference experiment, the profile is modified, exhibiting lower tracer concentrations at sub-halocline depths except for the bottom layer where tracer concentrations are slightly increased. Thus, apart from some modulation due to the different buoyancy forcing, the emergence of deep water formation is qualitatively unchanged in the Cold Climate experiment.

Apart from the changes in local deep water formation, we are interested in changes in the inflow of tracer-marked water from the northern sub-basins into the Baltic proper. The dashed lines in Figure 9 show the mean brine tracer concentration at Landsort Deep in the Cold Climate experiments compared to the equivalent winter 1986/1987 experiments (solid lines). In all of the Cold Climate experiments, more brine tracer is arriving at Landsort Deep than in their respective reference experiments. This is because, due to the lower air temperatures, more sea ice is formed, and hence, more brine tracer is created. The depth ranges of brine tracer inflow are, however, unchanged in the Cold Climate experiments and the main inflow is still within the surface layer. In the BRP bottom experiment, where the brine tracer is brought to the bottom of the sea, the brine tracer inflow in the surface layer is increased only very little by the Cold Climate forcing. The deeper inflow observed in this experiment is still present and more than doubled in strength with the Cold Climate forcing. However, it is not found at greater depths, as one could expect due to a formation of denser water, but peaks even at slightly lower depths.

Summarizing, despite the increased sea ice formation and modified buoyancy forcing, the Cold Climate experiments yield qualitatively similar results to their corresponding winter 1986/1987 experiments.

4. Discussion

The model experiments conducted in this study suggest that the process of convective wintertime deep water formation plays no major role for bottom water ventilation in the northern and central Baltic Sea, which is in line with the established view (Marmefelt & Omstedt, 1993). By analyzing deep water formation in the individual northern Baltic Sea sub-basins through passive tracer experiments, the southern central part of the Bothnian Bay (in the vicinity to station BO3) was identified as a region where winter surface water from the same basin is brought down to the bottom of the sea. Although brine rejection along the eastern coast might have enhanced this deep water formation, it is primarily attributable to an inflow of denser Bothnian Sea surface water, mixing with Bothnian Bay surface water and sinking to the bottom following the basin topography. This finding is in agreement with the conclusions drawn by Neumann et al. (2020) based on ship-based measurements. In the Bothnian Sea and the deep parts of the Gulf of Finland, no bottom water renewal by winter surface water could be detected in the reference experiment. Hence, there is also no lateral inflow of deep water formed during wintertime in the northern basins ventilating the northern central Baltic Sea, as hypothesized to occur in cold climate periods by Moros et al. (2020). Instead, the inflow from the northern basins into the Baltic proper is restricted to the surface layer above the halocline at around 60 m. These results are robust to changing the atmospheric forcing to colder climate conditions by reducing the air temperature by 2 K. The introduction of a BRP, mixing all salt released during ice formation to the ground (BRP bottom experiment), is likely to overestimate the actual effect of localized brine rejection but leads to deep water formation in all northern sub-basins. As a result, deep water formed in the Gulf of Finland ventilates the Landsort Deep in a depth range of about 60–90 m. Due to a salinity bias at Landsort Deep (see supporting information), the inflow of Gulf of Finland bottom water might occur at even greater depths in reality. Although only appearing in the BRP bottom experiment, the inflow can most likely be attributed to a wind-driven estuarine circulation reversal (Elken et al., 2003) and not to a circulation change due to brine rejection. The effect of localized brine rejection has no major impact on the deep water exchange between the Gulf of Bothnia and the Baltic proper and is too weak to create a dense water mass which could potentially contribute to a ventilation of Landsort Deep bottom waters.

Similar results were found by performing comparable simulations with the ocean circulation model GETM (Burchard & Bolding, 2002) by Giesse (2018). The robustness of the results throughout various sensitivity experiments and models strongly suggests that convective wintertime deep water formation is not a relevant process for the ventilation of the northern and central Baltic Sea, at least in present climate conditions. Still, there is clear geological evidence for the occurrence of this process during cold climate periods (Moros et al., 2020). It could, therefore, be that the chosen approach, based on an ocean circulation model, is not suitable to fully and realistically simulate the involved processes or that further factors, not taken into account in our simulations, are crucial for deep water formation in the northern Baltic Sea. We discuss important limitations of our approach in the following.

4.1. Salinity Bias

As the convective sinking of water masses is mainly determined by the stratification conditions in the water column, a good representation of the hydrographic conditions in the model is crucial for this study. While the MOM ocean circulation model reproduces observed temperatures and surface salinities well, bottom water salinities are overestimated in the Bothnian Sea and the northern Baltic proper, in particular at Landsort Deep, where modeled bottom salinities exceed the observed salinities by approximately 2 g/kg (see supporting information). This, in turn, leads to unrealistically high stratification in the respective basins. As a consequence, the depth of convective sinking might be underestimated at these locations and also inflows into the Baltic proper might layer into too shallow depths due to the inaccurate density profile. However, we could show that, even with ground-reaching brine rejection, the salinity increase in the northern basins is not sufficient to reach salinities comparable to observed salinities in the great depths of Landsort Deep. We therefore do not expect the salinity biases to qualitatively change the general conclusions drawn from the model results. Furthermore, one should note that the usage of a different ocean circulation model would not solve the issue, as other models have similar difficulties in realistically representing salinity and other tracer concentrations in the northern Baltic Sea, as pointed out by Placke et al. (2018).

4.2. Bathymetry and Sill Depths

Circulation patterns and water exchange between different basins of the Baltic Sea are strongly impacted by the seafloor topography, especially by bathymetric sills (e.g., Laanearu & Lundberg, 2003; Lass & Mohrholz, 2003). Prominent bathymetric features can be localized on scales much smaller than the model grid size. As a consequence, those features are smoothed out and replaced by an average depth per grid cell in the model bathymetry.

The deep water exchange between the northern Baltic proper and the Bothnian Sea is of particular interest for our study. It occurs through the Åland Sea, which consists of two basins and three sills with depths of 100 (northern Åland sill) and 70 m (middle and southern Åland sills) (Leppäranta & Myrberg, 2009). Jakobsson et al. (2019) provide a detailed comparison of the Åland Sea bottom topography in high-resolution multibeam data from the R/V *Electra* and two digital bathymetric models (DBMs): IOWTOPO (resolution of 2×1 arcmin, Seifert et al., 2001) and the more recent EMODnet (resolution of $1/16 \times 1/16$ arcmin, EMODnet Bathymetry Consortium, 2018). While the main morphology of the Åland Sea is well represented in EMODnet compared to the high-resolution data from R/V *Electra*, the coarser resolved IOWTOPO DBM, on which our model bathymetry is based, only gives a rough picture of the highly complex Åland Sea bottom topography. Deviations in depth of up to 100 m are found in some locations. The sills are located nearly at the same locations as in EMODnet. However, they are substantially shallower. The northern sill has a depth of 57 m in IOWTOPO (compared to 88 m in EMODnet and 100 m reported by Leppäranta & Myrberg, 2009) and the middle and southern sill have a depth of 49 m in IOWTOPO (compared to 60 m in EMODnet and 70 m reported by Leppäranta & Myrberg, 2009). Furthermore, the middle sill is represented by only one deep passage in IOWTOPO instead of three passages identified in EMODnet. The depths of the sills separating the Bothnian Sea and the Bothnian Bay are likely to be equally underestimated in IOWTOPO. The Gulf of Finland, however, is not separated from the Baltic proper by any sills (Leppäranta & Myrberg, 2009).

It is of course not possible to represent the bottom topography on scales smaller than the model grid size. However, the loss of structure and the underestimation of sill depths could influence and possibly restrict the deep water exchange in the model. It is conceivable that in reality, small-scale bottom currents that are highly dependent on topographic features can form and exchange deep water between the Gulf of Bothnia and the central Baltic Sea. Those bottom currents could be too small-scaled to be resolved in a typical regional ocean circulation model but would imprint to the seafloor morphology, for instance as erosion channels or drift deposits as found by Moros et al. (2020) and also in the high-resolution multibeam data from R/V *Electra* presented in Jakobsson et al. (2019).

4.3. Unresolved Processes

A crucial point in determining the potential of brine rejection in causing deep water formation is the degree of mixing of the rejected salt with the ambient water. Does the rejected highly saline brine sink down to the bottom or does it mix within the surface layer? When performing simulations with an ocean circulation model, one has to rely on the implemented vertical mixing scheme to realistically capture the mixing and convection. Here, the KPP scheme (Large et al., 1994) is used; in analogous GETM simulations (Giesse, 2018), a k-ϵ-turbulence closure model (Burchard et al., 1999) and vertically adaptive σ-coordinates are used (Burchard & Beckers, 2004; Hofmeister et al., 2010). While the mixing schemes can be assumed to work sufficiently well for homogeneous surface boundary conditions leading to widespread convection, difficulties arise when the surface boundary conditions are heterogeneous; that is, the buoyancy flux varies on scales smaller than the horizontal resolution of the model (e.g., Losch et al., 2006). The averaging of the buoyancy flux over the grid cell reduces the actual potential of deep-reaching, local convection. Here, the localized convection has been accounted for by performing sensitivity experiments with a subgrid-scale BRP adopted from Nguyen et al. (2009). This parameterization has some shortcomings. For example, it neglects the mixing of heat, momentum, and tracers other than salt and should not be applied under certain conditions (e.g., for widespread convection or high ice-ocean relative velocities) (Barthélemy et al., 2015). It is therefore not clear whether the introduction of a BRP leads to improved model results in this particular case. Irrespective of this, the sensitivity experiments with BRP show that the degree of mixing of rejected brine with ambient water does not impact the large-scale circulation and deep water exchange between the northern sub-basins and the Baltic proper in the model.

Nevertheless, the fact that current, state-of-the-art ocean circulation models tend to have difficulties in realistically representing the hydrographic conditions in the northern Baltic Sea (Placke et al., 2018) supports the hypothesis of processes being not or not well represented in the models for this particular region. The issue of a low-resolution bathymetry, as discussed above, could also add to or be the primary reason for the model biases. As the modeling of convection and deep water formation has proven difficult also in other regions in the world (e.g., the northwestern Mediterranean Sea; Somot et al., 2016), a better understanding of the processes taking place in the northern Baltic Sea might be needed. High-resolution and non-hydrostatic model simulations might give more conclusive insights.

4.4. Restricted Simulation Period

The analysis and the sensitivity experiments are restricted to the period of one particular winter and the following months. We chose the year 1986/1987 because it has been one of the coldest winters in the more recent climate period, with the largest ice extent since the simulation start in 1950. We would therefore expect the largest impact of brine rejection in this winter. However, when it comes to circulation changes, other factors such as the wind forcing, which can vary substantially from year to year, also have a strong impact. Moreover, the sensitivity experiments lack a spin-up period and it might take much longer for the circulation to fully adapt to the different atmospheric forcing or the application of a BRP. Longer simulations might be more conclusive, but would also be computationally more expensive.

4.5. Paleosalinity

Our model simulations are based on the present climate state. In the Cold Climate experiments, we have modified the air temperature forcing only. However, also other forcing factors which impact the hydrographic conditions could have been different during colder climate periods like the LIA. Paleoenvironmental reconstructions indicate that the salinity of the Baltic Sea was lower during the LIA compared to the present climate period (Emeis et al., 2003; Leipe et al., 2008), which might have gone along with a less pronounced halocline separating the surface and bottom water masses. Giesse (2018) has shown that in the extreme scenario of an unstratified Baltic Sea, with homogeneous initial conditions for temperature and salinity, brine rejection would indeed lead to deep water formation in the northern sub-basins and cause an inflow of these dense bottom waters into the Baltic proper. Although the scenario of an unstratified Baltic Sea is only hypothetical, it is very likely that the actual hydrographic conditions, in particular during the LIA period, were less stratified than represented in the model, which could have facilitated deep water formation.

5. Summary and Conclusions

In this study, we revisited the role of convective deep water formation in Baltic Sea deep water renewal by analyzing passive tracer pathways in several sensitivity experiments based on the severe winter 1986/1987 with the ocean circulation model MOM. In particular, we assessed the potential effect of localized brine rejection by applying a BRP (Nguyen et al., 2009) to the Baltic Sea for the first time. Generally, our model results support the established view that convective deep water formation does not play a major role in Baltic Sea deep water renewal. We find no evidence for deep water formation caused by thermal or haline convection in the northern sub-basins and, hence, also no ventilation of the northern Baltic proper through deep water formed in the sub-basins during wintertime. These results are robust to changing the atmospheric forcing to colder climate conditions by reducing the air temperature. Localized brine rejection could potentially lead to deep water formation. However, the effect is too weak to impact the large-scale circulation and deep water exchange between the Baltic proper and the sub-basins and to create a substantial dense water mass in the Bothnian Sea, which could possibly ventilate the northern Baltic proper. Solely, bottom water from the Gulf of Finland is found to contribute to the ventilation of the Landsort Deep at sub-halocline depths, caused, however, by a wind-driven reversal of the estuarine circulation.

The used model approach has some shortcomings, such as a salinity bias leading to overestimated stratification in the northern Baltic Sea, non-resolved bathymetric features influencing the deep water exchange, and a general limitation in representing small-scale processes due to resolution. Moreover, we have restricted the simulation period to one specific winter as a case study and have not taken into account potentially different hydrographic conditions during cold climate periods like the LIA. Nevertheless, we would like to stress that models with different vertical coordinates and different mixing schemes lead to the same conclusion that under present climate, deep water formation would be too weak to counteract the estuarine circulation in the northern Baltic Sea (Giesse, 2018).

Still, geological findings suggest wintertime deep water formation to have contributed to deep water renewal in the northern Baltic Sea during cold climate phases (Moros et al., 2020) and there is clear evidence for a better ventilation during the LIA (e.g., Kabel et al., 2012; Leipe et al., 2008). Considering the model limitations, we can gain insights on how and under which circumstances convective deep water formation could potentially have occurred and ventilated the Baltic Sea bottom waters. It is conceivable that in reality, deep water formation is associated with the formation of small-scale bottom currents, which cannot be resolved in the model. These bottom currents could leave the Gulf of Bothnia through the western Åland Sea following the bathymetry. Larger-scale ventilation during the LIA could be explained by not only reduced temperatures but also reduced salinity going along with weaker stratification conditions, which would be in agreement with paleoenvironmental studies such as Emeis et al. (2003) and Leipe et al. (2008). A different wind forcing and increased storminess during the LIA could further impact the ventilation (Meier, 2005; Zillén & Conley, 2010).

Bringing together the geological evidence for ventilation of the northern Baltic Sea through wintertime deep water formation and ocean model simulations would be an important contribution to improve process understanding in the northern Baltic Sea in different climate periods. Since wintertime observations in the northern Baltic Sea are scarce and information about past climate periods that can be retrieved from geological and paleontological studies is limited, the role of models for process understanding is crucial. The fact that various state-of-the-art ocean models have difficulties to realistically represent the hydrographic conditions in the northern Baltic Sea (Placke et al., 2018) emphasizes the need for improved process understanding and representation in models. The question of ventilation processes in the Baltic Sea is particularly interesting in the context of hypoxia in different climate periods (e.g., Jilbert et al., 2015; Papadomanolaki et al., 2018; Zillén et al., 2008). Having shown that, in our model approach, accounting for localized brine rejection and reduced air temperature is not sufficient to cause a ventilation of the northern Baltic Sea through wintertime deep water formation, further research is needed. Sensitivity experiments with different salinity and stratification conditions for the LIA and/or high-resolution process studies analyzing, for example, the deep water exchange through the complex bathymetry of the Åland Sea could bring more insights.

Data Availability Statement

The model data generated and analyzed in this study are available in a repository (https://thredds-iow.io-warnemuende.de/thredds/catalogs/publications/thomas/catalog_giesse2020.html).

Acknowledgments

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