Changes in Ocean Temperature in the Barents Sea in the Twenty-First Century

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ABSTRACT

Possible modifications to ocean temperature in the Barents Sea induced by climate change are explored. The simulations were performed with a coupled ice–ocean model (CIOM) driven by the surface fields from the Canadian Regional Climate Model (CRCM) simulations. CIOM can capture the observed water volume inflow through the Barents Sea Opening. The CIOM simulation and observations suggest an increase in the Atlantic water volume inflow and heat transport into the Barents Sea in recent decades resulting from enhanced storm activity. While seasonal variations of sea ice and sea surface temperature in CIOM simulations are comparable with observations, CIOM results underestimate the sea surface temperature but overestimate ice cover in the Barents Sea, consistent with an underestimated heat transport through the Barents Sea Opening. Under the SRES A1B scenario, the loss of sea ice significantly increases the surface solar radiation and the ocean surface heat loss through turbulent heat fluxes and longwave radiation. Meanwhile, the lateral heat transport into the Barents Sea tends to increase. Thus, changes in ocean temperature depend on the heat balance of solar radiation, surface turbulent heat flux, and lateral heat transport. During the 130-yr simulation period (1970–2099), the average ocean temperature increases from 0.8 °C to 1.8 °C in the southern Barents Sea, mostly due to increased lateral heat transport and solar radiation. In the northern Barents Sea, ocean temperature decreases by 0.4 °C from the 2010s to the 2040s and no significant trend can be seen thereafter, when the surface heat flux is balanced by solar radiation and lateral heat transport and there is no notable net heat flux change.

1. Introduction

Warm saline Atlantic water enters the Barents Sea mainly through the Barents Sea Opening (BSO; Fig. 1a; Blindheim 1989; Furevik 2001; Ingvoldsen et al. 2002; Skagseth et al. 2008, 2011; Rudels et al. 2013). On average, the mean volume and heat transports are estimated to be 2.3 Sv (1 Sv = 10⁶ m³ s⁻¹) and 70 TW (Smedsrud et al. 2013), but there is a rapid loss of heat as a result of atmospheric cooling in winter (Midttun 1985; Håkkinen and Cavaliere 1989; Zhang and Zhang 2001; Serreze et al. 2007; Arthun and Schrum 2010; Smedsrud et al. 2010; Arthun et al. 2011). The strongest heat flux is about 500 W m⁻² near the marginal ice zone in winter, which can cool the warm saline Atlantic water all the way to the ocean bottom (Håkkinen and Cavaliere 1989). Moreover, the warm Atlantic water inflow keeps the southern Barents Sea largely ice free and increases air–sea interactions (Helland-Hansen and Nansen 1909; Sandø et al. 2010; Beszczyńska-Moller et al. 2011; Arthun et al. 2012; Smedsrud et al. 2013; Onarheim et al. 2015).

The cold outflow from the Barents Sea accounts for approximately half of the Atlantic water flow into the central Arctic Ocean (Rudels et al. 1994; Schauer et al. 1997; Maslowski et al. 2004; Lien and Trofimov 2013; Rudels et al. 2013; Long and Perrie 2015). On its way through the Barents Sea, the Atlantic water is modified as a result of the mixing with surrounding water and the associated transformation, through cooling and ice formation (Midttun 1985). In the northwestern Barents Sea, the modified Atlantic water enters Nansen Basin through the Victoria Channel, and the average water volume flux is estimated to be about 0.4 Sv (Maslowski et al. 2004; Aksenov et al. 2010; Arthun et al. 2011). However, most of the Atlantic water in the Barents Sea continues to move northeastward into the Kara Sea between Franz Josef Land and Novaya Zemlya (Loeng et al. 1997; Gammelsrod et al. 2009), where it enters the central Arctic Ocean, through the St. Anna Trough (Simonsen and Haugan 1996; Schauer et al. 2002; Gammelsrod et al. 2009). Recent studies suggest that the inflow from the St. Anna Trough plays an important role...
in the heat balance in the central Arctic Ocean. Specifically, because of the reduced heat transport from the Barents Sea and increased surface heat loss, the ocean temperature associated with the Atlantic water layer tends to decrease under the SRES A1B scenario (Long and Perrie 2015).

There is a significant multidecadal variation in the ocean temperature in the Barents Sea (Levitus et al. 2009; Boitsov et al. 2012). On average, the ocean temperature was relatively colder at the beginning of the last century and in the 1970s, but warmer in the 1930s–50s and in recent decades. As part of the multidecadal variations in the Barents Sea, the temperature at 100–150 m increased by approximately 4°C from the late 1970s to the later part of the 2000–10 decade, because of a combination of oceanic and atmospheric changes (Levitus et al. 2009). For example, the ocean temperature at the Barents Sea Opening increased by about 1°C in recent decades, which enhanced the heat transport into the Barents Sea (Skagseth et al. 2008; Årthun et al. 2012). In addition, the increases in solar radiation and surface air temperature may also play an important role (Levitus et al. 2009). Overall, while the ocean temperature in the Barents Sea is dominated by multidecadal variability (Levitus et al. 2009), a positive trend has also been identified (Boitsov et al. 2012).

In a recent study (Long and Perrie 2015), we showed that under the SRES A1B scenario the ocean temperature in the northern Barents Sea tends to decrease, resulting from the enhanced air–sea interactions associated with the ice loss, which reduces the heat transport into the central Arctic Ocean. The objective of this study is to investigate the processes associated with the changes of the ocean temperature in the Barents Sea in the twenty-first century, focusing on the heat balance and the role of Arctic storms. Section 2 describes the model, the experimental design, and the methodology for storm tracking and heat flux computation. Section 3 shows the comparisons between the CIOM simulations and available observations. The simulated changes of ocean temperature in the Barents Sea are discussed in section 4. Section 5 discusses biases in our model simulations, as well as their impacts on our results. Section 6 presents the conclusions.

2. Model description and experiment design

A coupled ice–ocean model (CIOM) was implemented in the Arctic Ocean (Yao et al. 2000; Wang et al. 2005; Long et al. 2012), based on the Princeton Ocean Model (POM; Blumberg and Mellor 1987) and a multicategory ice model (Hibler 1979, 1980). It was successfully applied to studies of the impacts of climate change in the Arctic Ocean, including freshwater content and sea surface height (Long and Perrie 2013, 2015). A detailed description of CIOM can be found in Long et al. (2012) and Long and Perrie (2013, 2015). Only a summary is given in this section specifying the model setup.

a. Ocean model

The CIOM domain is centered on the central Arctic Ocean on a rotated spherical surface with the North Pole at 88°N, 131.5°E (Fig. 1a). Its horizontal resolution...
is $0.29^\circ \times 0.25^\circ$ with 25 vertical sigma levels. For example, for an ocean point with a depth of 3500 m, the vertical resolution is about 6 m for the upper seven layers, 380 m near the bottom, and 60 m between 200 and 400 m. To minimize numerical errors, the bottom topography is smoothed so that $\Delta H/\overline{H}$ is smaller than 0.2, where $\Delta H$ is the difference in the depths of adjacent grids while $\overline{H}$ is the mean of the grids (Mellor et al. 1994). In addition, the Neptune effect is implemented so that the model simulation can include the cyclonic rim currents associated with topographic stress (Holloway 1992). The Neptune velocity $U^*$ is estimated from the bathymetry, and the difference between ocean currents $U$ and $U^*$ $(U - U^*)$ is used to compute the viscosity term to parameterize the interactions between eddies and the bottom topography (Holloway 1992).

Because of the presence of sea ice, the vertical mixing in the Arctic Ocean is relatively weak (Fer 2009), but the increased vertical ocean heat transport due to loss of sea ice has been shown to play an important role in the future change of Atlantic water in the central Arctic Ocean (Long and Perrie 2015). The vertical mixing coefficients in CIOM are estimated using a second-order turbulence closure scheme (Mellor and Yamada 1982). In addition, the background mixing coefficients depend on local ice concentrations, ranging from 1 to $2 \times 10^{-5}$ m$^2$ s$^{-1}$ for vertical eddy viscosity and from $5 \times 10^{-6}$ to $1 \times 10^{-5}$ m$^2$ s$^{-1}$ for eddy diffusivity. An extra $3 \times 10^{-5}$ m$^2$ s$^{-1}$ is added to the background viscosity and diffusivity for the water depths above 70 m to approximately represent the effects of surface waves and eddies (Fer 2009).

The barotropic transports are specified along open boundaries, based on observations (Beszczynska-Moller et al. 2011). The inflows through Bering Strait (0.8 Sv) and the Norwegian Sea (8.5 Sv) are balanced by the outflows through the Canadian Archipelago (2.3 Sv) and along the east coast of Greenland (7.0 Sv). However, the baroclinic transports are estimated using radiation boundary conditions. As shown in section 4a, the Atlantic water inflow into the Barents Sea is mainly driven by surface wind associated with the storms in the Barents Sea (Bengtsson et al. 2004; Skagseth et al. 2008), and the slight changes that may occur in the lateral boundary conditions have no significant impacts on the changes of Atlantic water in the Barents Sea (Long and Perrie 2015).

Recent studies have shown that there are significant biases in the ocean temperature and salinity variables from the Third Generation Canadian Coupled Global Climate Model (CGCM3) simulations (Long and Perrie 2015). To provide the lateral boundary conditions for CIOM, the CGCM3 monthly temperature $T$ and salinity $S$ fields are modified using the Polar Science Center Hydrographic Climatology (PHC) data by Steele et al. (2001). In this methodology, we first computed the $T$ and $S$ anomalies of the CGCM3 outputs from their long-term means (1979–2008), and then added them to the PHC climatology to get lateral boundary boundaries for CIOM. Forced with the modified $T$ and $S$ variables, CIOM was successfully applied to study the impacts of climate change on the freshwater content in the Beaufort Sea and the Atlantic Water Layer in the central Arctic Ocean (Long and Perrie 2013, 2015). No “nudging” or restoration procedures are used.

River runoff is a major freshwater source for the Arctic Ocean (Serreze et al. 2006; Long and Perrie 2013). There are 13 major rivers along the Arctic coast and the river runoff in the Barents Sea is estimated to be about 632 km$^3$ yr$^{-1}$, which is a major contribution to the low salinity Barents Seawater (Dankers and Middelkoop 2008; Schauer et al. 2002; Smelserud et al. 2010). The runoff from the rivers is prescribed as precipitation minus evaporation ($P - E$) using monthly climatological river runoff observations (Prange and Lohmann 2004) for 1970–99. In addition, to represent the impacts of climate change, a positive trend is added for the years from 2000 to 2099 with an amplitude of 0.12% yr$^{-1}$, giving a total increase of 4.15 km$^3$ (Wu et al. 2005).

b. Ice model

The sea ice component of CIOM is based on a multi-category ice thickness distribution function (Thorndike et al. 1975; Hibler 1980) and a viscous–plastic sea ice dynamics model (Hibler 1979). It has seven categories of ice thickness and is able to estimate heat, moisture, and momentum fluxes at the ice–ocean interface as determined by appropriate boundary processes (Mellor and Kantha 1989). Total downward shortwave radiation is estimated, following Shine and Crane (1984), with the cloud correction formulation of Reed (1977) and sea ice albedo is determined from snow thickness, ice surface temperature (Koltzow 2007), and ice thickness. In addition, longwave radiation is given by the Smith and Dobson (1984) parameterization, and turbulent heat fluxes and wind stress are estimated using bulk formulas. In CIOM, the surface fluxes are defined as the sum of sensible heat flux, latent heat flux, and longwave radiation, whereas solar radiation is considered separately because of its penetration process. Air–ice and air–water drag coefficients are set to $2.75 \times 10^{-3}$ (Prinsenberg and Peterson 2002) and $1.1 \times 10^{-3}$, respectively, while the ice–water drag coefficient is $5.5 \times 10^{-3}$ (Shirasawa and Ingram 1991).

The ice model provides POM with heat, moisture, and momentum fluxes, and POM provides surface temperature, salinity, and current to the ice model, as feedback.
variables. POM and the ice model exchange variables at every time step.

c. Experiment

Our objective is to investigate the processes associated with the changes in the Barents Sea temperature in the twenty-first century, particularly in regard to the heat balance and the role of storms. Because of its coarse horizontal resolution, a global climate model tends to underestimate the high wind events, such as storms (Long et al. 2009). The resolution of the atmospheric output of CGCM3 is 2.5° and for the ocean output is 1.85°, which are too coarse to simulate detailed ocean circulation features like the ice edge (Long and Perrie 2015). To simulate the storm climatology and climate change in the Arctic, the Canadian Regional Climate Model (CRCM) is used to downscale CGCM3 simulations for 1970–2099 to provide high-resolution (45 km) surface driving fields for CIOM. In this methodology, CGCM3 outputs provide CRCM with initial and boundary conditions (winds, geopotential height, temperature, specific humidity, sea ice, and sea surface temperature) and CRCM outputs provide driving fields for CIOM [surface air temperature, specific humidity, precipitation, total cloud cover, sea level pressure (SLP), and 10-m winds).

We follow the SRES A1B climate change scenario (IPCC 2007), which assumes increasing carbon dioxide emissions until about 2050, and then decreasing emissions thereafter. As shown in section 4, it is notable that a decrease in emissions after 2050 could potentially affect the eventual temperature projections for the Barents Sea. CRCM simulations can capture the basic patterns of sea level pressure and surface air temperature found in NCEP–NCAR reanalyses. For example, CRCM has a reasonable simulation of the Beaufort high in the Beaufort Sea and the low pressure system in the eastern Arctic. The low pressure system over the Barents Sea and Nordic seas is an extension of the Icelandic low and responsible for transporting warm, moist Atlantic air into the Arctic. While CRCM is able to capture the overall features of this low pressure system, it is underestimated by about 3 hPa, compared to NCEP reanalysis data (Long and Perrie 2013). Moreover, the 6-hourly surface fields from the CRCM simulations can be successfully applied in studying the impacts of climate change (Long and Perrie 2013, 2015). The related details, concerning the surface fields from the CRCM simulations and the implementation of CIOM in the Arctic Ocean, are given by Long and Perrie (2013, 2015). Forced with the surface CRCM fields, CIOM was integrated from 1970 to 2099 in this study. As part of the initialization methodology, CIOM was integrated for 60 yr to reach stability for ocean currents, sea ice, ocean temperature, and salinity, for simulation of the present climate, driven by the two rounds of the CRCM surface variables for 1970–99. There is no restoration or nudging for ocean temperature and salinity.

The focus of our analyses is the ice–ocean simulations of the Barents Sea (Fig. 1b) from 1979 to 2099. Although the Barents Sea is largely ice free in summer (Fig. 2a), there is significant ice cover in the northern Barents Sea in winter (Fig. 3a). Based on the observed average position of the ice edge in winter (Fig. 3a), the Barents Sea is divided into the northern Barents Sea (box A in Fig. 1b) and the southern Barents Sea (box B in Fig. 1b).

The results are not sensitive to the definition of the boxes, based on our analyses, which is discussed in section 5.

d. Cyclone activity

Our study will specifically consider how changes in the storm activity in the Barents Sea can be correlated with the water volume transport through the Barents Sea Opening. To detect and track the storms (Long et al. 2009; Long and Perrie 2012), we first make an analysis of the SLP fields to identify local minima, where a candidate SLP center must satisfy the following criteria:

1) local minimum of less than 1010 hPa within a radius of 240 km and
2) at least one closed isobar on a 4-hPa increment.

Second, we use a simple nearest neighbor search in order to track the individual storms from the preceding (6 h) time step. If a storm falls within 600 km of a storm that was identified in the preceding time step, it is assumed to be a continuation of the previous storm; otherwise, it is considered a new storm. A candidate SLP center is required to have a lifetime of at least 24 h. We use “track density” to measure the storm activity, which is defined as the number of storm tracks passing any given grid point. Results regarding annual storm-track density and water volume transport are presented in section 4.

e. Water volume and heat transports

The lateral heat transport is computed as \( \int_A (T - T_0) v \, dA \), where \( A \) is the lateral area, and \( v \) is the current normal to the area, which is positive (negative) if the water flows into (out of) one of the selected boxes shown in Fig. 1b. Here, \( T \) is ocean temperature and \( T_0 \) is the reference temperature, which is set to 0°C. Since the lateral boundaries for boxes A and B are closed, the lateral heat transport is not sensitive to \( T_0 \). For the Barents Sea Opening, the lateral heat transport is
FIG. 2. Ice concentration in September, showing averages (1980–2009) for (a) HadISST, (b) CIOM results, and (c) the differences between CIOM results and HadISST.

FIG. 3. As in Fig. 2, but for March.
computed along the section between Svalbard and Norway, and there may be a bias in the estimate of heat transport because of its dependence on $T_0$ (Montgomery 1974; Schauer and Beszczynska-Möller 2009; Årthun et al. 2012). However, the temperature changes in the Barents Sea are discussed based on the heat budget in the boxes A and B, and the bias will not affect the conclusion. The water volume for the Barents Sea Opening is computed for the whole water column as $\int_A \nu dA$.

3. Present climate

a. Heat fluxes

The ocean heat transport through the Barents Sea Opening is a major heat source for the Barents Sea, which can affect the ice cover and air–sea interactions and plays an important role in the Barents Sea climate variability (Smedsrud et al. 2010; Sandø et al. 2010; Årthun et al. 2012; Onarheim et al. 2015). The observed mean water volume transport through the Barents Sea Opening is about 2.3 Sv, and the associated heat flux is about 70 TW (Smedsrud et al. 2013). However, the CIOM simulations suggest about 2.3-Sv Atlantic water volume transport and about 55-TW heat flux into the Barents Sea during the period 1980–2000 (Fig. 4). Therefore, although CIOM simulates the magnitudes of the water volume transport well, it underestimates the ocean heat transport through the Barents Sea Opening.

The heat entering the southern Barents Sea is largely lost as turbulent heat fluxes and longwave radiation (Sandø et al. 2010; Smedsrud et al. 2010; Beszczynska-Möller et al. 2011). In terms of model evaluation, during the period 1980–2000, the simulated net heat transport into the southern Barents Sea through lateral boundaries is about 40 TW (Fig. 5b). Both CIOM simulations and previous studies (Smedsrud et al. 2010) suggest that solar radiation contributes about 25 TW (Fig. 5b). In addition, CIOM simulations suggest that 60 TW is lost to the atmosphere in the southern Barents Sea (Fig. 5b), while about 15 TW is lost to atmosphere in the northern Barents Sea (Fig. 5a), which is consistent with previous studies (Smedsrud et al. 2010, 2013).

In recent decades, observations also show increased Atlantic water inflow and associated heat transport into the Barents Sea through the Barents Sea Opening (Skagseth et al. 2008; Årthun et al. 2012). The annual upward trend in heat flux is 2.5 TW, which is due to a positive trend in the water volume transport combined with an increase in ocean temperature (Skagseth et al. 2008; Årthun et al. 2012). Although the CIOM simulation shows an increase in the heat transport through the
Barents Sea Opening, its magnitude is significantly underestimated. The simulated heat transport increases from about 58 TW in the late 1990s to about 63 TW in the mid-2010s with an annual increase of about 0.5 TW (Fig. 4b).

Consistent with the observed temperature increase at the Fugløya–Bear Island section (73°N, 20°E), the average temperature along the Barents Sea Opening in the CIOM simulation does increase from about 3.7°C in 1980 to about 4.0°C in 2015 (Fig. 4c); but the magnitude of the temperature increase is notably underestimated, compared to observations (Skagseth et al. 2008; Årthun et al. 2012). Moreover, while the simulated temperature at the Kola section (71.5°N, 33.5°E) shows the observed increase in 1980–2015, the CIOM simulation underestimates the ocean temperature by about 1°C (Fig. 4d). In addition, the average ocean temperature along the Barents Sea Opening shows significant decadal variations (Fig. 4c). For example, both the simulation (Fig. 4c) and the observations (Drinkwater et al. 2014; Levitus et al. 2009, Smersrud et al. 2013; Zhang 2015) suggest a temperature decrease in the early 1990s and an increase in the late 1990s.

b. Ocean temperature and salinity

Warm Atlantic waters enter the southern Barents Sea through the Barents Sea Opening and rapidly lose heat through intense air–sea interactions and surface heat fluxes (Figs. 6–8). For example, the ocean temperature at 80 m is 5°–6°C near the Barents Sea Opening, but rapidly decreases from 4°–5°C near the northern Norwegian coast to from −1° to 0°C in the northeastern Barents Sea (Fig. 7). Similar patterns can be seen in the ocean temperatures at 150 m and at the surface (Figs. 6 and 8). While the CIOM simulation shows the warm Atlantic water in the southern Barents Sea and cold water in the northeastern Barents Sea, the CIOM overestimates the ocean temperatures at depths of 80 and 150 m by about 1°–2°C near the St. Anna Trough, and underestimates temperatures in the southern Barents Sea (Figs. 7 and 8). In addition, the CIOM simulations can reproduce the seasonal variation of sea surface temperature in PHC data (Fig. 6d), but significantly underestimate the sea surface temperature in the southern Barents Sea (Fig. 6c).

Moreover, observations show that the ocean temperatures in the southern Barents Sea tend to increase in recent decades, but the magnitude of temperature increase is underestimated in the CIOM simulations (Levitus et al. 2009; Figs. 9a,b). The area-averaged ocean temperature in the CIOM simulations linearly increases from about −0.1°C in the 1980s to about 0.2°C in the 2010s in the southern Barents Sea (Fig. 9b). However, there is no significant change in the northern Barents Sea (Fig. 9a). In the southern Barents Sea, the simulated increases in the ocean temperature are mainly associated with the enhanced heat transport through the Barents Sea Opening, decreased surface heat flux, and increased solar radiation (Figs. 4 and 5).

Because of the Atlantic water inflow through the Barents Sea Opening, the sea surface salinity is relatively high in the southern Barents Sea. The sea surface salinity is about 34.5–35 psu near the Barents Sea Opening, and gradually decreases eastward, reaching about 33.5–34 psu in the northern Barents Sea. While the CIOM simulation shows the salinity decrease from the southern Barents Sea to the northeastern Barents Sea, it underestimates the salinity values by 0.5–1.5 psu in the southern Barents Sea and along the Norwegian coast (Fig. 10), suggesting an underestimated vertical mixing in the Barents Sea.

c. Ice concentration

The presence of sea ice effectively modifies the exchange of heat, moisture, and momentum between the
FIG. 6. Annual ocean temperature (°C) at surface, showing averages (1980–2009) for (a) CIOM results, (b) PHC data, and (c) the differences between CIOM results and HadISST. (d) Seasonal variations of sea surface temperature for CIOM results (black) and PHC data (red), averaged over Fig. 1b.

FIG. 7. As in Figs. 6a–c, but at 80 m.
atmosphere and the ocean, compared to open water conditions (Andreas et al. 2010). When the ocean is fully covered with sea ice, the presence of ice reduces the exchanges of heat, moisture, and momentum between the atmosphere and the ocean compared to conditions in the marginal ice zone (MIZ). When ice forms, it influences the salinity through brine rejection; the formation of dense bottom water associated with the brine rejection is a regular phenomenon in the Barents Sea (Midttun 1985; Årthun et al. 2011). Furthermore, the cold bottom water moves into the central Arctic Ocean through the Victoria Channel and St. Anna Trough, affecting the heat balance at intermediate layers, and playing an important role in the Arctic Ocean ventilation (Long and Perrie 2015).

On a seasonal time scale, the ice cover in the Barents Sea is strongly correlated with air temperature, and shows significant seasonal variations (Fig. 11). While the northern Barents Sea is covered with sea ice in winter, the southern Barents Sea is dominated by open water (Fig. 3), influenced by the warm Atlantic water inflow. The minimum in ice cover occurs in August–September when the Barents Sea is almost ice free (Fig. 2), mostly due to increased solar radiation and reduced surface heat loss. Compared to observations such as represented by HadISST (Rayner et al. 2003), CIOM simulates the September ice cover (Fig. 2) and seasonal variation of ice cover reasonably well (Fig. 11a), but notably overestimates the March ice concentration in the central Barents Sea (Fig. 3). Figure 11a gives the time series of ice area averaged over the domain shown in Fig. 1b. While the model simulation shows a reasonable seasonal variation, the simulated minimum ice cover occurs in August, one month earlier than in observed data from HadISST (Fig. 11). In addition, CIOM significantly overestimates the ice cover from October to June.

4. Impacts of climate change in the Barents Sea

a. Water volume and heat transport through the Barents Sea Opening

We explore the correlation between the changes in the storm activity in the Barents Sea and the water volume transport through the Barents Sea Opening. Under the SRES A1B scenario, the Arctic warms faster as a result of positive feedback, namely because of the reduced surface albedo associated with the loss of snow and ice at high latitudes. Concomitantly, the north–south temperature gradient in the Northern Hemisphere tends to decrease causing the northeastward shift of the storm tracks (Long et al. 2009). Although the changes in the storm tracks are subject to competing
thermodynamic influences, our simulations suggest that the reduced temperature gradient dominates the process, consistent with Shaw et al. (2016).

Using the methodology outlined in section 2d, the simulated annual storm-track density in the Barents Sea is about 16–18 storms per year in 1980–99 and there is a significant increase after the 2000s (Fig. 12), consistent with previous observational studies (Zhang et al. 2008; Benestad et al. 2016). In addition, the changes in the storm-track density in the Barents Sea are significantly correlated with the water volume transport through the Barents Sea Opening (Fig. 13), because the westerly winds associated with the cyclonic atmospheric circulation enhance the wind-driven oceanic inflow through the Barents Sea Opening (Bengtsson et al. 2004; Ingvaldsen et al. 2004; Skagseth et al. 2008). However, it is noteworthy that the change in storm intensity is weak, and there is no significant correlation between the water volume transport through the Barents Sea Opening and the storm intensity in the Arctic.

The maximum increases in the storm activity in the Barents Sea occur in 2000–19, and the increases are relatively weak thereafter, showing multidecadal variations (Fig. 12). Correspondingly, the simulated water
volume transport through the Barents Sea Opening increases from 2.3 Sv in the 1980s to 2.7 Sv in the 2020s, but decreases after the 2030s, and there is no significant trend (Fig. 4a). The associated ocean heat transport has a similar change, but shows significant multidecadal variability (Fig. 4b), mostly due to variations of ocean temperature (Fig. 4c), as shown in Drinkwater et al. (2014) and Levitus et al. (2009). The correlation between the water volume and heat transports is 0.67. Although other recent studies suggest a decrease in northward heat transport in the northern North Atlantic (Yeager et al. 2015), the heat transport through the Barents Sea Opening tends to increase in 2000–19, because of the increased storm-track density. In contrast to the decrease of the water volume transport after 2030s, there are no significant changes in the heat transport, which is mostly due to the temperature increase in the Barents Sea Opening (Fig. 4c), consistent with recent studies (Smedsrud et al. 2013; Sandø et al. 2014). The average heat transport after the 2030s is about 65 TW, which is about 10 TW higher than the average during 1980–2000.

b. Ocean temperature

The average ocean temperature in the northern Barents Sea decreases from about −0.2°C in the 2010s to about −0.5°C in the 2030s, and there is no significant change thereafter (Fig. 9a). For example, the temperature at 80 m is from about −1°C to 0°C in 1980–99, but slowly decreases after 2000–19, and the decrease reaches −1°C in the northwestern Barents Sea by the end of the century, in 2080–99 (Fig. 14). While the changes at a depth of 150 m show similar patterns (Fig. 15), the sea surface temperature in the northern Barents Sea tends to increase, and the maximum increase is about 2°C in the northwestern Barents Sea in the period of 2080–99 (Fig. 16).

In the southern Barents Sea, the average ocean temperature tends to increase from the 1980s to the 2090s. The average temperature increases from about 0°C in the 2000s to about 1°C in the 2090s (Fig. 9b). Meanwhile, the warm water at 80 m is limited to coastal areas during 1980–99, but slowly expands northeastward, and dominates the southern Barents Sea in 2080–99. The maximum increase in ocean temperature is about 3°C in 2080–90 (Fig. 14). Similar patterns can be seen at 150 m, but the warming is relatively weaker than at 80 m (Fig. 15). The maximum increase in ocean temperature occurs at the surface with a magnitude of 5°C in the central Barents Sea in 2080–99 (Fig. 16). Consistent with previous studies, the average ocean temperature in the southern Barents Sea shows significant multidecadal variability (Drinkwater et al. 2014).
c. Heat balance in the Barents Sea

The temperature decrease in the northern Barents Sea from the 2010s to the 2030s is mainly associated with the increased surface heat flux (Fig. 5a). Under the climate change scenario, the average surface heat flux increases from 15 TW in 2000s to about 30 TW in the 2090s. However, the solar radiation and lateral heat transport tend to increase the ocean temperature. The average solar radiation gradually increases from about 10 TW in the 2000s to about 20 TW in the 2090s (Fig. 6a).

In addition, the increase in the lateral heat transport is relatively weak and increases from about 5 TW in the 2000s to about 10 TW in the 2090s. Therefore, the changes in solar radiation and lateral heat transport tend to partly offset the impacts of the enhanced surface heat flux. In terms of net heat flux, the northern Barents Sea loses heat from the 2000s to the 2030s, but reaches equilibrium thereafter (Fig. 5a). Correspondingly, the average ocean temperature decreases from the 2000s to the 2040s, and there is no significant trend from the 2040s to the 2090s (Fig. 9a).

By comparison, in the southern Barents Sea, the increased lateral heat transport and solar radiation play...
important roles in the ocean temperature increases (Fig. 5b). Under the SRES A1B scenario, the solar radiation increases from about 20 TW in the 2000s to about 50 TW in the 2090s. Meanwhile, the lateral heat transport increases from 40 to about 60 TW in the 2020s, but there is no significant trend thereafter. Similar to the changes in the northern Barents Sea, the surface heat flux increases from about 60 TW in the 2000s to about 100 TW in the 2090s, which partly offsets the impacts resulting from the increased lateral advection and solar radiation, and the net heat flux tends to increase ocean temperature (Figs. 5b and 9b).
Therefore, changes in the ocean temperature in the Barents Sea depend on the heat balance among lateral ocean heat transport, solar radiation, and surface heat flux (Sandø et al. 2010; Smedsrud et al. 2010). Under the SRES A1B scenario, more heat is lost to the atmosphere in the northern Barents Sea before the 2040s, and the ocean temperature tends to decrease. However, the southern Barents Sea receives more heat through solar radiation and ocean heat transport than is lost to the atmosphere through the turbulent heat flux and longwave radiation, and thus the ocean temperature tends to increase.

d. Air–sea interactions

The changes in solar radiation and surface heat flux (Fig. 5) are mainly related to the loss of sea ice in the Barents Sea. Because of the increases in surface air temperature in the Barents Sea and the lateral heat transport (Fig. 5), there is a decreasing linear trend in the ice cover in the Barents Sea (Smedsrud et al. 2013; Sandø et al. 2014; Figs. 3b and 11). For example, the northern Barents Sea is mostly covered with sea ice in November in the 2000s but is largely ice free in the 2090s (Fig. 17), and the total ice cover decreases from about $2 \times 10^6$ to about $1 \times 10^6 \text{km}^2$ in November (Fig. 11b). Furthermore, the loss in sea ice significantly reduces surface albedo and increases the solar radiation absorbed by the ocean (Fig. 5), but increases the ocean heat loss through the turbulent heat fluxes and longwave radiation in the Barents Sea (Fig. 5).

5. Discussion

The CIOM simulations for the present climate have been discussed in detail in section 3. Overall, CIOM underestimates the heat transport through the Barents Sea Opening and its increase in recent decades. While the sea surface salinity is underestimated by 0.5–1.5 psu in the southern Barents Sea, the CIOM sea surface temperatures are underestimated by 2°–4°C in the southern Barents Sea and 0.5°–1.5°C in the northern Barents Sea. Moreover, the ice concentration is overestimated in the southern Barents Sea, compared to observations.

The biases in the CIOM simulations are consistent with previous studies. For example, Sandø et al. (2014) used ROMS to downscale GISS-AOM (ROMS-G) and NCAR CCSM3 (ROMS-N) simulations, following the IPCC SRES A1B scenario (IPCC 2007). Compared to observations, ROMS-G underestimates the ocean salinity (temperature) but significantly overestimates the ice cover in the southern Barents Sea. While ROMS-N has a reasonable simulation of water volume transport through the Barents Sea Opening, it underestimates the associated heat transport, as shown in the CIOM simulation.

It is noteworthy to point out that CIOM has a reasonable simulation of the ocean response to atmospheric surface forcing. Compared to observations, it can reproduce seasonal variations of ice cover (Fig. 11a) and sea surface temperature (Fig. 6d) in the Barents Sea. Consistent with previous studies (Bengtsson et al. 2004; Ingvaldsen et al. 2004; Skagseth et al. 2008), the simulated water volume transport through the Barents Sea Opening shows a strong correlation with the storm-track density in the Barents Sea (Fig. 13). In particular, CIOM has a reasonable simulation of water volume transport through the Barents Sea Opening and ice concentration in September. Therefore, CIOM can be a useful tool to simulate the ocean response to the changes in atmospheric surface forcing under the SRES A1B scenario (Long and Perrie 2013, 2015).

Under the SRES A1B scenario, the overestimated ice concentration in the southeastern and central Barents Sea can cause overestimates in ice loss and changes in surface solar radiation and turbulent heat flux. Since the net effect of surface solar radiation and surface heat flux is heat loss from the ocean (Fig. 5), the overestimated ice loss in the southeastern and central Barents Sea may cause overestimates in the net heat loss and underestimates in the temperature increase in the southern Barents Sea. In addition, temperature changes near the bottom are mainly located near the Norwegian coast (Fig. 15) where there is no significant sea ice presence (Figs. 2 and 3). Indeed, both CIOM simulations (Fig. 16) and previous studies (Sandø et al. 2014) show increased sea surface temperatures in the southern Barents Sea.

In this study, the Barents Sea is divided into the northern Barents Sea (box A in Fig. 1b) and the

FIG. 13. Correlation coefficient between water volume transport through the Barents Sea Opening and annual storm-track density. Only the coefficients that have significance at the 99% level are shown.
southern Barents Sea (box B in Fig. 1b), based on the average observed ice edge in March for the present climate (Fig. 3a). To understand the sensitivity of our results to the definition of these two areas, boxes A and B (Fig. 1b), the northern Barents Sea is extended southward so that it includes most of the central Barents Sea (Fig. 18c). The resulting time series of ocean temperature averaged over the subdomains in Fig. 18c are similar to those shown in Fig. 9, and the correlation coefficients are 0.98 for box A and 0.81 for box B. For example, the temperature averaged over box A decreases from −0.2°C in the 1990s to −0.6°C in the 2030s, and no significant change can be seen after the 2040s. However, the temperature averaged over box B...
gradually increases from about 0°C in the 1980s to about 1.5°C in the 2090s.

6. Conclusions

A coupled ice-ocean model (CIOM) is implemented in the Arctic Ocean to simulate the impacts of climate change on the ocean temperature in the Barents Sea. Compared to observations (Rayner et al. 2003; Skagseth et al. 2008; Smedsrud et al. 2010; Rudels et al. 2013), the CIOM simulations can capture the observed seasonal variations of the sea ice cover in the Barents Sea as well as the observed magnitudes of water volume through the Barents Sea Opening. Over the present climate period (1980–2009), both the CIOM simulation and observations suggest a positive trend in the Atlantic water volume and associated

![Figure 15](image-url)
heat transports through the Barents Sea Opening, because of enhanced storm-track density in the region. However, most of the heat from the Barents Sea Opening is lost in the southern Barents Sea through the air–sea interactions, whereas the surface heat flux in the northern Barents Sea is relatively weak, as suggested in the observations. Therefore, CIOM is capable of simulating oceanic responses to the changes in atmospheric surface forcing. It is noteworthy to point out that there are significant biases in the CIOM simulations, with overestimates in the ice concentration and underestimates in the ocean temperature in the southern Barents Sea. Moreover, although CIOM can reproduce the water volume transport through the Barents Sea Opening, it underestimates the associated heat transport. Therefore, the conclusions from this study are more qualitative rather than quantitative.

We investigated the future changes in ocean temperature through the heat balance among the lateral heat...
FIG. 17. Ice concentrations in November.
transport, surface heat flux, and solar radiation. In terms of the impacts of climate change in the Barents Sea, sea ice plays an important role in the heat budget. Under the SRES A1B scenario, there is a decreasing trend in ice cover in the Barents Sea. While the reduced albedo associated with the ice loss significantly increases the solar radiation absorbed by the ocean, the ice loss increases the ocean heat loss through the enhanced turbulent heat fluxes and longwave radiation. In addition, there is an increasing trend in the lateral heat transport into the Barents Sea through the Barents Sea Opening during the period from the 1980s to the 2020s, which stabilizes thereafter.

The changes of the ocean temperature in the southern Barents Sea show a different pattern from that in the northern Barents Sea. In the southern Barents Sea, the average ocean temperature tends to increase under the SRES A1B scenario, because of the increased lateral heat transport and solar radiation. Although there is an increase in the heat loss through the turbulent heat flux, the net impacts of the heat balance tend to increase the ocean temperature. However, in the northern Barents Sea, the heat loss through the turbulent heat flux and longwave radiation is dominant from about the 1980s to the 2030s, and the average ocean temperature decreases from about −0.2°C in the 2010s to about −0.6°C in the 2040s, but stabilizes thereafter. In addition, the changes in the artic storm activity play an important role in the increased lateral heat transport through the Barents Sea Opening.

The results in this study are obtained from one-member simulation, and there are still some significant uncertainties. As an example, although both the CIOM simulations and previous studies suggest reduced ice cover in the northern Barents Sea and increased ocean temperature in the southern Barents Sea (Sandø et al. 2014), the changes in the ocean temperature in the northern Barents Sea are still uncertain (Smedsrud et al. 2013; Sandø et al. 2014).

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