Roles of Sea Ice–Surface Wind Feedback in Maintaining the Glacial Atlantic Meridional Overturning Circulation and Climate

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ABSTRACT

During glacial periods, climate varies between two contrasting modes, the interstadials and stadials. These climate changes are often associated with drastic reorganizations of the Atlantic meridional overturning circulation (AMOC). Previous studies highlight the important role of sea ice in maintaining contrasting modes of the AMOC through its insulating effect on the oceanic heat flux and the buoyancy flux (sea ice–buoyancy flux feedback); however, the effect of feedback from the atmosphere remains unclear. Here, the effect of sea ice–surface wind interactions over the North Atlantic Ocean on the AMOC is explored. For this purpose, results from comprehensive atmosphere–ocean coupled general circulation models (AOGCMs) are analyzed. Then, sensitivity experiments are conducted with the atmospheric component of the AOGCM. Last, to explore the impact of modifications in surface winds induced by sea ice on the maintenance of the AMOC, partially coupled experiments are conducted with the AOGCMs. Experiments show that the expansion of sea ice associated with a weakening of the AMOC reduces surface winds by suppressing the oceanic heat flux and increasing the atmospheric static stability. This wind anomaly then causes a weakening of the wind-driven ocean salt transport to the northern North Atlantic and maintains the weak AMOC, therefore working as a positive feedback. It is shown that, together with the sea ice–buoyancy flux and sea ice–surface wind feedback, changes in sea ice and oceanic heat flux sustain the contrasting modes of the AMOC. These results may provide useful information for interpreting the differences in the self-sustained internal oscillations of the AMOC produced by recent AOGCMs.

1. Introduction

During glacial periods, climate varies on millennial time scales and shifts between two contrasting modes, the interstadials and stadials (Dansgaard et al. 1993; Rahmstorf 2002; Capron et al. 2010; Kawamura et al. 2017). Interstadials are characterized by warm conditions over the North Atlantic Ocean and Greenland, while stadials are characterized by drastically colder conditions (Huber et al. 2006; Kindler et al. 2014). It is shown that drastic changes in sea ice over the northern North Atlantic occur along with temperature changes in Greenland (Dokken et al. 2013; Sadazki et al. 2019). As a result of the expansion of sea ice during the stadials, which suppresses the oceanic heat flux to the

atmosphere during boreal winters, the air temperature over the North Atlantic and Greenland decreases drastically and hence results in a contrasting climate different from that of interstadials (Li et al. 2010).

Both reconstruction and modeling studies show that these climate and sea ice changes are associated with modifications in the glacial Atlantic meridional overturning circulation (AMOC) between two contrasting modes, the vigorous mode and weak mode (Ganopolski and Rahmstorf 2001; Piotrowski et al. 2005; Kissel et al. 2008; Arzel et al. 2012; Peltier and Vettoretti 2014; Böhm et al. 2015). When the AMOC is in a vigorous state, it transports heat to the north and warms the northern North Atlantic, whereas the weakening of the AMOC is related to reduced oceanic heat transport, southward expansion of sea ice, and drastic cooling over the northern North Atlantic (Brown and Galbraith 2016; Kawamura et al. 2017). Previous studies explore the important processes and feedbacks that drive or maintain the AMOC and climate variability (Paul and Schulz 2002). For example, changes in the vertical structure of

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ocean temperature in the North Atlantic (Winton 1997; Kim et al. 2012; Gong et al. 2013) and the meridional salinity gradient between the subpolar and subtropical regions (Peltier and Vettoretti 2014) are shown to play a role in driving the AMOC variability, while changes in oceanic convection and the subpolar gyre are shown to maintain the AMOC by affecting the transport of salt into the deep-water formation region (Born et al. 2016). Furthermore, recent studies suggest the importance of the interaction between ocean, sea ice, and atmosphere over the North Atlantic in affecting the AMOC variability (Arzel et al. 2010; Friedrich et al. 2010; Li and Born 2019). However, the role of interactions between sea ice and atmosphere on the glacial AMOC and climate remains unclear, as described in the next two paragraphs. In this study, we focus on this aspect.

Previous studies report that modifications in sea ice can exert a substantial impact on the glacial AMOC and climate by affecting the oceanic heat flux and oceanic buoyancy flux (Gildor and Tziperman 2003; Loving and Vallis 2005; Oka et al. 2012; Dokken et al. 2013). The expansion of sea ice associated with a reduction of the AMOC causes a suppression of atmosphere–ocean heat exchange because the surface water cannot be cooled further. As a result, the buoyancy loss of the surface ocean is restricted. This suppression of the buoyancy flux then plays a role in maintaining the weak AMOC by inhibiting deep-water formation (Oka et al. 2012; Zhu et al. 2015). Hence, the expansion of sea ice associated with the weakening of the AMOC contributes to maintaining the weak AMOC (sea ice–buoyancy flux positive feedback).

On the other hand, changes in sea ice are also known to exert a large impact on the atmosphere, such as surface winds, by affecting the static stability of the atmosphere and the activity of storm tracks (Kushnir et al. 2002; Magnusdottir et al. 2004; Byrkjedal et al. 2006). This suggests a possible feedback from sea ice–surface wind interactions on the AMOC, because modeling studies show that modifications in surface winds exert a large influence on the AMOC and climate through their effect on sea surface salinity and sea ice (Timmermann and Goosse 2004; Montoya and Levermann 2008; Saenko 2009; Sherrieff-Tadano et al. 2018). In fact, a recent atmosphere–ocean coupled general circulation model (AOGCM) study implies that the modification of surface winds associated with a weakening of the AMOC and the expansion of sea ice may play a role in maintaining the AMOC in a weak state (Zhang et al. 2014). However, because of the complicated coupling process in the AOGCM, the role of sea ice in causing the surface wind anomaly, as well as the impact of the surface wind anomaly on the AMOC, remains elusive.

In this study, we explore the role of sea ice–surface wind feedback on the glacial AMOC and climate. In particular, we assess the impact of changes in surface winds induced by modifications in the AMOC and sea ice on the maintenance of the AMOC itself. For this purpose, we first analyze results of the Model for Interdisciplinary Research on Climate, version 4 (medium resolution) (MIROC4m), AOGCM experiments (Abe-Ouchi et al. 2015; Kawamura et al. 2017), which exhibit contrasting modes of the AMOC, and explore the wind anomalies associated with modifications in the AMOC. Second, we perform sensitivity experiments with an atmospheric general circulation model (AGCM) to extract the role of sea ice and associated changes in oceanic heat flux in inducing the wind anomalies. Last, to understand the impact of changes in surface winds induced by modifications in the AMOC and sea ice on the maintenance of the AMOC, we conduct partially coupled experiments with the AOGCM (Mikolajewicz and Voss 2000; Schmittner et al. 2002; Stouffer et al. 2006; Watanabe et al. 2014). In these experiments, surface wind stresses, which are passed to the oceanic component, are replaced with a specific wind stress forcing. From these analyses and experiments, we show that the sea ice–surface wind feedback plays an important role in maintaining the contrasting modes of the AMOC and the climate. Further, we integrate the results from previous studies to show that sea ice and oceanic heat flux play a role in maintaining the contrasting modes of the glacial AMOC and climate.

This study is organized as follows. In section 2, we describe the model and numerical experiments. In section 3, we investigate the changes in surface wind over the North Atlantic associated with modifications in the AMOC by analyzing the results from the MIROC AOGCMs. Results from AGCM sensitivity experiments are also shown to explore the cause of the wind anomaly. In section 4, we explore the impact of changes in surface winds induced by the modifications in the AMOC on the maintenance of the AMOC by means of partially coupled experiments. In section 5, we discuss and compare our results with previous studies. In section 6, the conclusions of this study are presented.

2. Model and numerical experiments

a. AOGCM

Two versions of the MIROC4m AOGCM (Hasumi and Emori 2004; Table 1) are used in this study. In these two versions, the coefficient of horizontal diffusion of the isopycnal layer thickness is slightly different in the

TABLE 1. Experiments for LGM from two versions of the MIROC4m AOGCM analyzed in this study: Model A (Abe-Ouchi et al. 2015) and Model B (Kawamura et al. 2017). The LGM experimental setup follows the protocol of the PMIP3 (Braconnot et al. 2012; Abe-Ouchi et al. 2015). The strength of the AMOC differs drastically between the two experiments because of a small difference in the coefficient of horizontal diffusion of the isopycnal layer thickness, that is, the Gent–McWilliams (GM) coefficient (Chikamoto et al. 2012; Kawamura et al. 2017).

Expt name	Strength of AMOC	GM coef	Reference
Model A	25 Sv	$300 \text{ m}^2 \text{ s}^{-1}$	Abe-Ouchi et al. (2015)
Model B	7 Sv	$700 \text{ m}^2 \text{ s}^{-1}$	Kawamura et al. (2017)

ocean general circulation model (OGCM; Chikamoto et al. 2012; Kawamura et al. 2017). It is $3.0 \times 10^2 \text{ m}^2 \text{ s}^{-1}$ in the model named Model A (Gent et al. 1995; Hasumi and Emori 2004; Abe-Ouchi et al. 2015) and $7.0 \times 10^2 \text{ m}^2 \text{ s}^{-1}$ in Model B (Hirst and McDougall 1996; Chikamoto et al. 2012; Kawamura et al. 2017). These models are used extensively for modern climate (Obase et al. 2017), paleoclimate (Abe-Ouchi et al. 2007; Abe-Ouchi et al. 2013; Kobayashi et al. 2015), and future climate studies (Yamamoto et al. 2015) and also have been used in model intercomparison studies (Kageyama et al. 2013). Otherwise, the basic setup of these two models is identical. The AGCM solves the primitive equations on a sphere using a spectral method (Numaguti et al. 1997). The horizontal resolution of the atmospheric model is $\sim 2.8^\circ$, and there are 20 layers in the vertical direction. The AGCM is coupled to a land surface model. The OGCM solves the primitive equation on a

sphere, where the Boussinesq and hydrostatic approximations are adopted (Hasumi 2006). The horizontal resolution is $\sim 1.4^\circ$ in longitude and 0.56° – 1.4° in latitude (latitudinal resolution is finer near the equator). There are 44 vertical layers. It is coupled to a dynamic–thermodynamic sea ice model. Note that these models reproduce the strength of the modern AMOC (Kawamura et al. 2017), the spatial pattern of mixed layer depth (Oka et al. 2012), and surface winds reasonably well (Sherriff-Tadano et al. 2018).

We first analyze simulations of the Last Glacial Maximum (LGM) obtained from the two versions of the MIROC4m AOGCM. The experimental setup of the LGM follows the protocol of the Paleoclimate Model Intercomparison Project 3 (PMIP3; Braconnot et al. 2012). This protocol specifies glacial ice sheets (Abe-Ouchi et al. 2015), greenhouse gas concentrations ($\text{CO}_2 = 185 \text{ ppm}$), and orbital parameters of the LGM. Both experiments are integrated for more than 6000 years from preindustrial climate simulations and, thus, have settled into a state of quasi equilibrium. In the two experiments, the strength of the AMOC differs drastically because of a small difference in the coefficient of horizontal diffusion of the isopycnal layer thickness (Chikamoto et al. 2012; Kawamura et al. 2017). The strength of the AMOC is 25 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) for Model A and 7 Sv for Model B (Fig. 1; Table 1), and sea surface conditions also differ drastically between the experiments (Fig. 2). Whereas the majority of the northern North Atlantic remains ice free in Model A (Figs. 2a,c), the region is largely covered by sea ice in Model B (Figs. 2b,d). Although different mixing coefficients are used in the oceanic models of the two experiments, the same atmospheric

FIG. 1. Meridional streamfunction (Sv) obtained from simulations of LGM conducted with (a) Model A and (b) Model B. The annual averages of the last 100 years of the simulation are presented. The strength of the AMOC differs drastically between the two models because of differences in the coefficient of horizontal diffusion of the isopycnal layer thickness. Further information is provided in section 2.

FIG. 2. Comparisons of sea surface conditions obtained from simulations of the LGM conducted with (a),(c) Model A and (b),(d) Model B for (left) annual mean sea surface temperature ($^{\circ}\text{C}$; shading) and 50% sea ice concentration (contours) and (right) oceanic heat flux (W m^{-2} ; color) and 50% sea ice concentration (contours) averaged over the boreal winter (January, February, and March). These variables are calculated using the last 100 years of the simulations. Note that the value of 50% is selected for sea ice concentration because it corresponds to the steep gradient of the oceanic heat flux.

component is used in both experiments. Thus, the response of winds to changes in the AMOC and sea ice can be detected by comparing the results of the two models.

The weaker AMOC obtained in Model B is attributed to the larger coefficient of horizontal diffusion of the isopycnal layer thickness, which enhances oceanic mixing and decreases sea surface temperature over the Southern Ocean by suppressing open ocean convection and reducing the Deacon cell (Kawamura et al. 2017). As a result, the production of Antarctic Bottom Water increases and the AMOC weakens and becomes shallower (Otto-Bliesner et al. 2007; Buizert and Schmittner 2015; Sun et al. 2016; Jansen 2016; Klockmann et al. 2016). Once the AMOC passes a threshold, its strength decreases drastically and shifts into a weak mode (Oka et al. 2012; Kawamura et al. 2017; Klockmann et al. 2018). This implies that the AMOC in these LGM simulations are close to the threshold (Colin de Verdière and Te Raa 2010; Arzel et al. 2010; Brown and Galbraith 2016). We are further analyzing this point by performing large ensemble simulations with the AOGCM by modifying

several climatic parameters, which will be reported in a forthcoming paper.

b. AGCM

To clarify the importance of changes in sea ice and oceanic heat flux on the surface wind anomalies, four experiments are conducted with an atmospheric component of the MIROC4m (Table 2: A-Model A, A-Model B, A-SEAICE, and A-SENSEVAP). To be consistent with the AOGCM, the same resolution is used in the AGCM experiments. All experiments are performed with PMIP3 LGM boundary conditions (Braconnot et al. 2012; Abe-Ouchi et al. 2015). Experiments differ only in sea surface conditions (Fig. 2). In A-Model A, we apply sea surface temperature and sea ice from Model A. In A-Model B, we apply sea surface temperature and sea ice from Model B. In A-SEAICE, we apply the extensive sea ice cover obtained from Model B over the northern North Atlantic but apply sea ice cover obtained from Model A for the region outside of the northern North Atlantic. In the fourth experiment (A-SENSEVAP), we suppress oceanic heat flux from the ocean. Flux is suppressed in the

TABLE 2. Lists of sensitivity experiments conducted with the AGCM. The impact of modifications of sea surface conditions on the wind field is explored. The experimental setup follows the PMIP3 protocol for the LGM, except for the sea surface temperature and sea ice distribution, which are taken from the climatology of Model A and B experiments (Fig. 2). In A-SENSEVAP, both the sensible and latent heat fluxes from the ocean to the atmosphere over the northern North Atlantic are suppressed. Flux is suppressed in the model code by multiplying it by a constant value of 0.005 in each grid cell over the portion of the northern North Atlantic that is covered by sea ice in A-SEAICE but not in A-Model A (Louis 1979).

Expt name	Sea surface temperature	Sea ice	Sensible heat flux	Latent heat flux
A-ModelA	Model A	Model A	—	—
A-ModelB	Model B	Model B	—	—
A-SEAICE	Model A	Model B over North Atlantic	—	—
A-SENSEVAP	Model A	Model A	Suppressed	Suppressed

model code by multiplying it by a constant value of 0.005 in the grid cells over the northern North Atlantic that are covered by sea ice in A-SEAICE but not in A-Model A (Fig. 5; Louis 1979). The value of 0.005 is selected after several test experiments are conducted. Slight differences in the value do not affect the result. Note that, because the surface friction is unchanged, this experiment does not include the effect of changes in surface friction due to sea ice. We define the effect of sea ice expansion over the northern North Atlantic as the difference between A-SEAICE and A-Model A (Table 3), the impact of suppression of oceanic heat flux on the winds as the difference between A-SENSEVAP and A-Model A, and the effect of differences in sea surface conditions in other regions and the effect of sea ice drag as the difference between A-Model B and A-SENSEVAP (residual effect). We run the AGCM for 35 years and analyze the climatology of the last 30 years.

c. Partially coupled experiments with MIROC AOGCM

To assess the impact of changes in sea ice–surface wind feedback on the maintenance of the AMOC, partially coupled experiments (Mikolajewicz and Voss 2000; Schmittner et al. 2002; Watanabe et al. 2014) are conducted with MIROC4m AOGCMs (Table 4). In these experiments, surface wind stresses, which are passed to the oceanic component, are replaced with a specific wind stress forcing. Raw daily winds of the last

50 years of Models A and B are utilized as wind stress forcing. The climatology of these wind fields is shown in Fig. 3. Results are unchanged if daily winds of the last 100 years are used (not shown). The velocity of the surface wind, which is used to calculate the coefficient of atmosphere–ocean heat and water exchange, is unchanged. Therefore, we focus only on the role of changes in momentum flux in the partially coupled experiments.

Two partially coupled experiments are conducted with Model A. The first experiment (PC-A-A) is used to validate our method. In PC-A-A, the daily surface winds of the last 50 years of Model A are applied cyclically to the same model. Hence the wind forcing, which affects the oceanic component, is almost identical to that in Model A. We regard our method as valid when the AMOC and other variables in PC-A-A remain similar to those in Model A. In the second experiment (PC-A-B), daily surface winds of the last 50 years of Model B are applied to Model A. Therefore, in PC-A-B, the changes in surface wind associated with modifications in the AMOC are removed (we refer to this as removal of the sea ice–surface wind feedback). Two other experiments

TABLE 4. Lists of partially coupled experiments conducted with the MIROC4m AOGCM. These experiments are conducted with both Models A and B. In PC-B-A, the surface wind stress of Model A is applied to Model B. In PC-B-B, the surface wind stress of Model B is applied the same model. In PC-A-B, the surface wind stress of Model B is applied to Model A. In PC-A-A, the surface wind stress of Model A is applied to the same model. Experiments PC-A-A and PC-B-B are used to validate our method. The effect of sea ice–surface wind feedback can be evaluated by comparing PC-A-B with PC-A-A or Model A and by comparing PC-B-A with PC-B-B or Model B.

Effect	Diff between expts	Expt name	Model	Wind stress	Surface wind feedback
Sea ice over North Atlantic	A-SEAICE minus A-ModelA	PC-A-A	Model A	Model A (strong wind)	On
Oceanic heat flux over North Atlantic	A-SENSEVAP minus A-ModelA	PC-A-B	Model A	Model B (weak wind)	Off
Residual	A-ModelB minus A-SENSEVAP	PC-B-A	Model B	Model A (strong wind)	Off
		PC-B-B	Model B	Model B (weak wind)	On

TABLE 3. Relationship between the effect of the changes in individual sea surface conditions and the sensitivity experiments in Table 2.

FIG. 3. (left) Surface (1000-hPa) winds (arrows; m s^{-1}) and (right) 500-hPa winds (arrows; m s^{-1}) obtained from the MIROC4m AOGCM. Thick green and dashed green contours in (a)–(d) show the sea ice edge (50% sea ice concentration). The climatology of boreal winter (January, February, and March) is presented. These variables are calculated using the last 100 years of the simulations. Color shading shows zonal winds (m s^{-1}). Shown are results for (a),(e) Model A, (b),(f) Model B, and (c),(g) differences between Model B and Model A. (d),(h) The impact of sea ice expansion over the northern North Atlantic on the wind fields, which are assessed based on AGCM sensitivity experiments (differences between A-SEAICE and A-Model A; Table 2).

FIG. 4. Results from MIROC4m AOGCM experiments. The impact of the weakening of the AMOC on the vertical profile of the westerly wind (m s^{-1}) over the North Atlantic (differences between Model B and A) is shown. The westerly wind is averaged zonally from 50° to 20°W . The climatology of January, February, and March is shown. This variable is calculated using the last 100 years of the simulations.

are conducted with Model B. In PC-B-A, surface winds of the last 50 years of Model A are applied cyclically to Model B, while in PC-B-B, surface winds of the last 50 years of Model B are applied to the same model. Sea ice–surface wind feedback is removed in PC-B-A, and PC-B-B is used for validation. By comparing the differences in the AMOC between Model A or PC-A-A and PC-A-B and between PC-B-A and PC-B-B or Model B, the impact of sea ice–surface wind feedback on the maintenance of the AMOC can be evaluated. Each experiment is integrated for more than 1000 years.

3. Response of winds to modifications in AMOC, expansion of sea ice, and suppression of oceanic heat flux

To understand the changes in the general pattern of the wind field associated with the modifications in the AMOC, the simulated surface and 500-hPa winds from the two versions of the MIROC4m AOGCM (Models A and B) are analyzed (Fig. 3). The 100-year climatology of boreal winter (January, February, and March) is presented, because deep water forms and anomalies of surface wind are largest during this season. When the AMOC is in the vigorous state (Model A; Figs. 3a,e), a vigorous surface wind of 8 m s^{-1} is present over the North Atlantic, which is associated with the topography and surface cooling of the North American ice sheet (Cook and Held 1988; Liakka 2012; Sherreff-Tadano et al. 2018). However,

the wind pattern is drastically different when the AMOC is in a weak state (Model B; Figs. 3b,f). Compared with the situation under a vigorous AMOC, surface wind is substantially reduced, whereas the westerly at 500 hPa is intensified under a weak AMOC. The change in the wind is more evident in the anomaly field (Figs. 3c,g). The weakening of the surface wind is associated with an anticyclonic circulation anomaly near the surface. In contrast, the strengthening of the westerly aloft is associated with a cyclonic circulation anomaly. Therefore, the response of winds depends on height. The anomaly changes sign around 900 hPa (Fig. 4).

To clarify the role of sea ice and the suppression of oceanic heat flux in the North Atlantic on the wind anomaly, sensitivity experiments are conducted with the AGCM (Table 2). Figures 3d and 3f and Fig. 5 show responses of the wind field to the expansion of sea ice (differences between A-SEAICE and A-Model A; Tables 2 and 3) and the suppression of the oceanic heat flux (differences between A-SENSEVAP and A-Model A; Tables 2 and 3), respectively. They show the critical role of sea ice and the oceanic heat flux in inducing the wind anomaly throughout the troposphere. Approximately 90% of the weakening of the surface westerly at midlatitude is explained by the expansion of sea ice and more than 80% of the weakening of the surface westerly at midlatitude induced by sea ice expansion is explained by the suppression of oceanic heat flux (Fig. 6). In contrast, it appears that changes in sea surface conditions in other regions and the increase in friction due to sea ice play a secondary role in the creation of the wind anomaly (residual effect, Fig. 6). Hence, these experiments confirm that the weakening of surface winds associated with the reduction of the AMOC is mostly caused by the expansion of sea ice and the suppression of the oceanic heat flux over the North Atlantic.

Mechanism by which sea ice modifies the wind field

Previous studies show that changes in sea ice modify the wind field by affecting the static stability at the surface or affecting storm-track activity (Kushnir and Held 1996; Kushnir et al. 2002; Hoskins and James 2014). These studies show that the dominant processes inducing the wind anomaly are detectable by analyzing its vertical structure (Kushnir et al. 2002). When the changes in static stability are important, the wind anomaly has a baroclinic structure, while the wind anomaly becomes barotropic when the changes in storm-track activity play an important role. Considering the baroclinic structure of the wind anomaly (Fig. 4), the expansion of sea ice seems to weaken the surface wind by affecting the surface air temperature and the static

FIG. 5. Response of atmospheric variables to a suppression of oceanic heat flux over the northern North Atlantic obtained from AGCM experiments (difference between A-SENSEVAP and A-Model A). The 30-year climatology of boreal winters is presented (January, February, and March). The oceanic (a) sensible and (b) latent heat fluxes (W m^{-2}) are suppressed in the AGCM by multiplying the flux by a constant value of 0.005 in each grid cell over the northern North Atlantic that is covered by sea ice in A-SEAICE but not in A-Model A. (c) Surface (1000-hPa) and (d) 500-hPa wind anomalies (arrows; m s^{-1}). Color shading in (c) and (d) shows 1000- and 500-hPa zonal wind anomalies (m s^{-1}), respectively.

stability of the lower troposphere, rather than affecting storm-track activity (Kushnir and Held 1996; Kushnir et al. 2002; Hoskins and James 2014). To verify this hypothesis, changes in surface air temperature, static stability of the lower troposphere, and eddy momentum fluxes, which are associated with changes in storm-track activity, are analyzed. Results are shown in Fig. 7. Near the surface, air temperature decreases drastically due to the suppression of sensible heat flux under sea ice expansion (Fig. 7a). The amplitude of the temperature anomaly decreases with height, and hence the static stability increases in the lower troposphere (Fig. 7b). This very large static stability anomaly then suppresses the growth of low pressure systems and generates an anticyclonic circulation anomaly near the surface (Byrkjedal et al. 2006), which is consistent with the wind anomaly obtained from AOGCM and AGCM experiments. With respect to storm-track activity, changes in sea ice decrease northward eddy momentum transport around 50°N (Fig. 7c). This forces an easterly surface wind anomaly north of 50°N and a westerly surface wind anomaly south of 50°N (Brayshaw et al. 2009; Merz et al. 2015); hence, a southward migration of the surface westerly. However, in the AOGCM and the AGCM, the surface westerly decreases south of 50°N due to the

expansion of sea ice (e.g., Fig. 6), which is different from the wind anomaly predicted from the changes in storm-track activity. Therefore, changes in storm-track activity do not seem to play a role in causing the surface wind anomalies in these experiments. The larger importance of static stability on the wind anomaly

FIG. 6. Decomposition of the surface westerly anomaly (m s^{-1}) over the North Atlantic ($50^{\circ}\text{--}10^{\circ}\text{W}$, $40^{\circ}\text{--}50^{\circ}\text{N}$) induced by the weakening of the AMOC based on AGCM sensitivity experiments: (a) effects of weakening of the AMOC (A-Model B minus A-Model A), (b) expansion of sea ice (A-SEAICE minus A-Model A), (c) suppression of oceanic heat flux (A-SENSEVAP minus A-Model A), and (d) residual (A-Model B minus A-SENSEVAP) for the 30-year climatology of January, February, and March.

FIG. 7. Responses of atmospheric variables to the sea ice expansion over the northern North Atlantic obtained from AGCM experiments (differences between A-SEAICE and A-Model A). The climatology of boreal winter (January, February, and March) is presented. Shown are (a) surface air temperature (K), (b) static stability at 925 hPa (s^{-1}), and (c) momentum flux by transient eddies (m^2s^{-2}) vertically averaged over 850–250 hPa. The band-pass filter of Murakami (1979) is applied to the wind anomaly in calculating the momentum flux to extract the changes in storm-track activity.

obtained from our model may be attributed to the large amplitude of the surface temperature anomaly. This temperature drop is so large ($\sim -30^\circ\text{C}$) that the temperature contrast between the relatively cold continent and the warm ocean during boreal winter, which is considered to be an important ingredient for the climatological cyclonic surface circulation over the northern North Atlantic (Icelandic low; Held 1983; Byrkjedal et al. 2006), almost disappears. As a result, changes in the static stability in the lower troposphere are likely to be important for inducing the wind anomaly in these experiments.

4. Impact of sea ice–surface wind feedback on the AMOC

The influence of sea ice–surface wind feedback on the maintenance of the AMOC is explored by means

of partially coupled experiments (Table 4). Figure 8a shows the time series of the AMOC obtained from partially coupled experiments conducted with the weak AMOC (Model B), which are compared with results from the fully coupled experiment (Model B, black curve). In the experiment in which the surface wind of Model A is applied to Model B (red curve, PC-B-A), the weak AMOC can no longer be maintained and the AMOC intensifies drastically. In a second experiment designed to validate our method, the surface wind of Model B is applied cyclically to the same model (blue curve, PC-B-B), and the AMOC remains weak as in Model B. Hence, these experiments show that the weakening of surface wind associated with the expansion of sea ice and the reduction of the AMOC plays an important role in maintaining the weak AMOC in Model B. Partially coupled experiments with the vigorous AMOC (Model A) also confirm these results

Impact of sea ice-wind feedback on the AMOC

FIG. 8. Results of partially coupled experiments conducted with Models (a) B and (b) A. Time series of the maximum strength of the AMOC (Sv) is shown. The original experiments are shown as black curves, and the results of partially coupled experiments are shown as colored curves. In the experiments indicated by blue curves, the surface wind stress of Model B is applied. In the experiments shown by red curves, the surface wind stress of Model A is applied. In (a), the black curve is Model B, the red curve is PC-B-A, and the blue curve is PC-B-B. In (b), the black curve is Model A, the red curve is PC-A-A, and the blue curve is PC-A-B. By comparing the differences in the AMOC between PC-A-A (with sea ice–surface wind feedback) and PC-A-B (without sea ice–surface wind feedback), and between PC-B-B (with sea ice–surface wind feedback) and PC-B-A (without sea ice–surface wind feedback), the impact of sea ice–surface wind feedback on the AMOC can be evaluated.

that the vigorous surface wind associated with a small extent of sea ice and extensive release of oceanic heat flux plays an important role in maintaining the vigorous AMOC in Model A (Fig. 8b). Hence, partially coupled experiments reveal that the changes in surface winds induced by modifications in the AMOC and sea ice are crucial for the maintenance of both the vigorous AMOC (Model A) and weak AMOC (Model B).

Application of surface winds modifies the AMOC by changing the sea surface salinity and sea ice over the deep-water formation region (Figs. 9c,d). In PC-B-A, after modification of the surface winds, the sea surface salinity increases over the deep-water formation region compared to Model B (black curve) and PC-B-B (blue curve) (Fig. 9c). This increase of sea surface salinity is associated with the increase of the transport of salt into the deep-water formation region due to the strengthening of Ekman upwelling (Fig. 10a; Sherriff-Tadano et al. 2018), as well as the strengthening of the subpolar gyre (Fig. 10b; Born et al. 2016). At the same time, the strengthening of the cyclonic surface wind as well as the vertical transport of heat causes a divergence of sea ice and an increase of sea ice melting at the deep-water formation region, respectively (not shown), which decrease the sea ice thickness (Fig. 9d), and as a result, weakens the insulating effect of sea ice (Fig. 9f). Once the surface salinity and sea ice

reach a threshold, deep-water formation and associated convective mixing initiate (Figs. 9b,e). This triggers a drastic release of the heat (Fig. 9f) stored in the subsurface ocean and induces further melting of sea ice. As a result, surface water loses its buoyancy and deep-water formation accelerates, which further strengthens the subpolar gyre via the thermal wind balance (Kleppin et al. 2015), and increases the salt transport into the deep-water formation region (Montoya et al. 2011; Born et al. 2016). This strong positive feedback then intensifies the AMOC abruptly (Fig. 9a). After the shift of the AMOC into the vigorous mode, the continuous formation of deep water, as well as the salt transport supported by the intense wind, maintain the vigorous AMOC for the duration of the simulation.

Similar processes operate in PC-A-B (Fig. 8b; Fig. 11, blue curve). In PC-A-B, after surface winds weaken, the sea surface salinity decreases over the deep-water formation region (Fig. 11a), which is associated with the weakening of the Ekman upwelling, subpolar gyre, and convective mixing (Figs. 12a,b; Fig. 11c; Oka et al. 2001; Montoya et al. 2011; Born et al. 2016; Sherriff-Tadano et al. 2018). At the same time, a southward expansion of sea ice (Fig. 11b) causes an increase in sea ice melt over the deep-water formation region (not shown) and a further decrease in the surface salinity. Once the surface salinity reaches a threshold, sea ice expands

FIG. 9. Results of partially coupled experiments conducted with Model B, focused on the first 5 yr. (a) Time series of the maximum strength of the annual mean AMOC (Sv), where the black curve is Model B, the red curve is PC-B-A, and the blue curve is PC-B-B. (b) Spatial map of sea ice concentration (colors) and deep-water formation region (red contour is the frequency of convective adjustment at 600 m) at the onset of the first deep-water formation event in PC-B-A [January of year 1; see also the vertical pink-shaded region in (d) and (e)]. Also shown are time series of (c) surface salinity (psu = practical salinity units), (d) sea ice thickness (cm), (e) frequency of convective adjustment at 600 m (unitless), and (f) oceanic heat release (K day^{-1}) at the deep-water formation region ($35^{\circ}\text{--}30^{\circ}\text{W}$, $55^{\circ}\text{--}60^{\circ}\text{N}$). In (a), the annual mean is presented, whereas monthly values are presented in (c)–(f).

drastically and deep-water formation and associated convective mixing ceases (Figs. 11b,c). As a result, the AMOC decreases abruptly within several decades. After the shift to the weak AMOC, the suppression of oceanic heat flux by sea ice (Fig. 11d), as well as the weak salt transport by the weak surface wind, maintains the weak AMOC.

Over the simulation period, the AMOC in PC-B-A (red curve in Fig. 8a) settles into the vigorous mode. However, the AMOC in PC-A-B (blue curve in Fig. 8b) exhibits internal variability; it returns to the vigorous mode 600 years after the drastic reduction, remains vigorous for less than 200 years, and then returns to the weak mode. This result shows that removal of the sea ice–surface wind feedback alone cannot maintain the weak AMOC in Model A. This is because feedback from other processes affects stratification over the deep-water formation region and promotes recovery of the AMOC

(Fig. 11e, e.g., an increase in subsurface ocean temperature; Vettoretti and Peltier 2016; Brown and Galbraith 2016). (Detailed analysis of the mechanism of the internal oscillation of the AMOC is beyond the scope of this study.) After the shift to the vigorous mode, the continuous release of heat from the sea surface feeds the formation of deep water and promotes the maintenance of the vigorous AMOC (Figs. 11c,d). However, without the feedback from surface winds, deep-water formation cannot be sustained because of insufficient surface salinity in the deep-water formation region (Fig. 11a). As a result, the AMOC shifts to the weak mode and the duration of the vigorous AMOC becomes short compared to that of the weak AMOC. Therefore, this result suggests that sea ice–surface wind feedback plays a role in increasing the periodicity of the internal variability of the AMOC. The significance of the sea ice–surface wind feedback can

FIG. 10. Time series of convergence of (a) vertical oceanic salt flux over the shallowest 100 m (psu day^{-1} , where a positive value corresponds to upward salt flux) and (b) barotropic streamfunction ($10^6 \text{ m}^3 \text{ s}^{-1}$) averaged over the deep-water formation region ($35^\circ\text{--}30^\circ\text{W}$, $55^\circ\text{--}60^\circ\text{N}$). The black curves are Model B, and the red curves are PC-B-A. Monthly means are presented in (a), and annual means are presented in (b). In (a), the vertical transport of salt by convective mixing is excluded. In (b), the barotropic streamfunction is averaged over $60^\circ\text{--}5^\circ\text{W}$, $48^\circ\text{--}65^\circ\text{N}$, following the method of Born et al. (2016). The increase in the negative value of the streamfunction corresponds to the strengthening of the subpolar gyre.

be confirmed from the fact that the absence of the sea ice–surface wind feedback causes the quasi-stable vigorous AMOC in Model A to shift into a condition in which the vigorous AMOC can last only for 200 years.

5. Discussion

The results discussed above demonstrate the significant role of sea ice–surface wind feedback on the maintenance of the AMOC. The weakening of surface winds induced by the weakening of the AMOC and the expansion of sea ice play a role in maintaining the weak AMOC. The strengthening of surface winds induced by a stronger AMOC and sea ice contraction has the effect of maintaining the strong AMOC. These results are summarized in Fig. 13, which considers an expansion of sea ice over the northern North Atlantic associated with the weakening of the AMOC. This induces a weakening of the surface wind, which is caused by the suppression of the oceanic heat flux. The suppression of oceanic heat flux reduces the surface wind by increasing the static stability of the atmosphere near the surface. Subsequently, this weakening of surface wind reduces the transport of salt into the deep-water formation region by weakening the wind-driven ocean circulation. In addition, the weakening of surface wind causes an increase of sea ice over the deep-water formation region. Due to

the decrease in surface salinity and the increase in sea ice, the weak AMOC is sustained. Thus, we have shown that the sea ice–surface wind feedback works as a positive feedback (Fig. 13). Together with the sea ice–buoyancy flux (Gildor and Tziperman 2003; Oka et al. 2012) and sea ice–surface wind feedback, we find that changes in sea ice and oceanic heat flux play a critical role in maintaining the contrasting modes of AMOC.

Several previous modeling studies report responses of surface winds to modifications in the AMOC. For instance, Zhang et al. (2014) and Brandefelt and Otto-Bliesner (2009) show a weakening of surface winds due to the reduction of the AMOC, which is consistent with the present study. In addition, similar results are also observed in an experiment conducted with the Geophysical Fluid Dynamics Laboratory climate model, version 2, with the Modular Ocean Model, version 4p1, at coarse resolution (CM2Mc) model under low- CO_2 and low-obliquity conditions (not shown; Galbraith and de Lavergne 2019). However, we should also note that some other comprehensive climate models and Earth system models of intermediate complexity show different responses of surface wind to a weakening of the AMOC. Swingedouw et al. (2009) and Pausata et al. (2011) show small changes in surface winds, whereas Timmermann et al. (2005) and Arzel et al. (2008) show a strengthening of surface winds as the AMOC weakens. These results suggest that the strength of this feedback, or perhaps its qualitative effect, may be model dependent. Differences in the response of surface winds to a weakening of the AMOC among models are still unclear and require additional study, though we speculate that the difference in sea ice behavior and atmospheric models may cause the largest discrepancies among models. For example, whereas Zhang et al. (2014, their Fig. 4) and Brandefelt and Otto-Bliesner (2009, their Fig. 2) show a drastic expansion of sea ice over the North Atlantic in response to the weakening of the AMOC, changes in sea ice are relatively small in Pausata et al. (2011, their Fig. 7). With respect to the atmospheric model, differences in the response of static stability over the low troposphere to modifications in sea ice among atmospheric models may cause model discrepancies. For example, in Arzel et al. (2008), as they use a simple atmospheric model, the static stability of the atmosphere is fixed and the role of the increase in static stability in reducing the surface wind is not taken into account. Because the sea ice–surface wind interaction plays a significant role on the maintenance of the AMOC, the cause of the discrepancy among models should be clarified in the future.

FIG. 11. Results of partially coupled experiments conducted with Model A. The black curves are Model A, and the blue curves are PC-A-B. Shown are time series of areal averaged annual mean oceanic variables over the deep-water formation region (35° – 30° W, 55° – 60° N) for (a) surface salinity (psu), (b) sea ice thickness (cm), (c) frequency of convective adjustment at 600 m (ND), (d) oceanic heat release (K day^{-1}), and (e) subsurface ocean temperature ($^{\circ}\text{C}$).

Furthermore, we should note that the effect of changes in sea ice transport may depend on models (Zhu et al. 2014; Sherriff-Tadano et al. 2018). For example, while this study shows that the strengthening of surface wind reduces the sea ice transport

into the deep-water formation region, other studies report different results (Arzel et al. 2008; Arzel and England 2013); the strengthening of surface wind increases sea ice transport into the deep-water formation region. The reason behind this model discrepancy remains unclear,

FIG. 12. Time series of convergence of (a) vertical oceanic salt flux over the shallowest 100 m (psu day^{-1} , where a positive value corresponds to upward salt flux) and (b) barotropic streamfunction ($10^6 \text{ m}^3 \text{ s}^{-1}$ averaged over the deep-water formation region ($35^\circ\text{--}30^\circ\text{W}$, $55^\circ\text{--}60^\circ\text{N}$). The black curves are Model A, and the blue curves are PC-A-B. Annual averages are presented in these plots. In (a), the vertical transport of salt by convective mixing is excluded. In (b), the barotropic streamfunction is averaged over $60^\circ\text{--}5^\circ\text{W}$, $48^\circ\text{--}65^\circ\text{N}$, following the method of Born et al. (2016). The decrease in the negative value of the streamfunction corresponds to the weakening of the subpolar gyre.

though it can cause differences in the effect of the sea ice–wind feedback among models. Hence, for these reasons, it is important to compare the impact of sea ice–surface wind feedback with other coupled models. This may also provide useful information for the interpretation of the differences in the structure of self-sustained internal oscillations produced by recent comprehensive coupled models (Peltier and Vettoretti 2014; Brown and Galbraith 2016; Klockmann et al. 2018).

Last, it is important to note the future tasks of our research. While this study focuses on the impact of sea ice–surface wind feedback on the maintenance of

the AMOC under one glacial condition (LGM), it is necessary to check the role of this effect under different background climates, such as warm (Yamamoto et al. 2015) and midglacial climates (Guo et al. 2019). This is important because the changes in sea ice associated with modifications to the AMOC can differ under different background climates, which will also affect the response of surface wind. We are planning to assess this effect using freshwater hosing experiments performed under preindustrial and midglacial climates by Kawamura et al. (2017). The second point is related to the time scale of the sea ice–surface wind interactions. In this study, we switch the surface winds of Models A and B, which have contrastingly different AMOC and sea ice conditions. In this way, we are able to assess the role of the sea ice–surface wind feedback on the maintenance of the AMOC on a long time scale (more than 100 years). However, in this method, we miss the interaction of sea ice and surface wind on the decadal time scale, which may be important in understanding the timing of the abrupt transition of the AMOC (Kleppin et al. 2015) or the speed of the transition. This is also an intriguing question and should be explored. Last, previous studies also suggest the importance of changes in freshwater flux associated with modifications in sea ice on the stratification of the ocean and the AMOC (Krebs and Timmermann 2007; Peltier and Vettoretti 2014; Jensen et al. 2016, 2018). The relative importance of changes in freshwater flux and surface winds on the AMOC should also be clarified in the AOGCM.

FIG. 13. Schematic of processes over the deep-water formation region by which sea ice maintains the AMOC and climate. The signs on the arrows are linked to the impact of the forcing on the target. Black arrows show results from previous studies, and red arrows show the result from this study.

6. Conclusions

In this study, we explore the role of the sea ice–surface wind feedback on the maintenance of the AMOC, which remains elusive in previous studies because of the complicated nature of AOGCMs. For this purpose, we first explore the response of surface winds to modifications in the AMOC and changes in sea ice by analyzing results from two different versions of the MIROC4m AOGCM. Second, we perform sensitivity experiments with an AGCM to clarify the role of changes in sea ice and oceanic heat flux on the wind anomaly. Last, to assess the role of sea ice–surface wind feedback on the maintenance of the AMOC explicitly, we conduct partially coupled experiments with the MIROC4m AOGCM. The results of the present study can be summarized as follows:

- The weakening or strengthening of the AMOC causes a reduction or enhancement, respectively, of surface winds over the North Atlantic.
- AGCM experiments show that the suppression of the oceanic heat flux due to the sea ice expansion decreases the surface wind by increasing the static stability of the atmosphere near the surface over the North Atlantic. The release of oceanic heat due to sea ice contraction has the opposite effect.
- Partially coupled experiments show that the weakening of surface wind reduces the transport of salt into the deep-water formation region by weakening the wind-driven ocean circulation. As a result, the surface salinity can be sufficiently low at the deep-water formation region, which inhibits deep-water formation, and hence sustains the weak AMOC. On the other hand, the strengthening of surface wind enhances salt transport by strengthening wind-driven ocean circulation, and then increases sea surface salinity at the deep-water formation region, which accelerates deep-water formation and sustains the vigorous AMOC.

Together with the sea ice–buoyancy flux (Gildor and Tziperman 2003; Oka et al. 2012) and sea ice–surface wind feedback, changes in sea ice and oceanic heat flux play an important role in maintaining the AMOC (Fig. 13). Last, it is important to check the robustness of our results with other coupled models. This may also provide useful information for the interpretation of the differences in the structure of self-sustained internal oscillations produced by recent comprehensive coupled models (Brown and Galbraith 2016; Klockmann et al. 2018; Vettoretti and Peltier 2018). It can also improve our understanding of the dynamics of the millennial-scale climate variability of the glacial period.

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