Sensitivity of the Global Thermohaline Circulation to Interbasin Freshwater Transport by the Atmosphere and the Bering Strait Throughflow

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ABSTRACT
Sensitivity of the global thermohaline circulation to interbasin freshwater transport by the atmosphere and the Bering Strait throughflow is investigated by using a free-surface, coarse-resolution ocean general circulation model. The model is run by prescribing freshwater flux at the sea surface without restoring the sea surface salinity to climatology in order that effects of salinity advection are properly represented. Comparison of experiments with the open and closed Bering Strait shows that the throughflow reduces the intensity of the Atlantic deep circulation by \( \sim 17\% \), while minimally affecting the Pacific deep circulation. Increase in the atmospheric freshwater transport from the Atlantic to the Pacific intensifies both the Atlantic deep circulation and the Bering Strait throughflow. On the other hand, changes in the throughflow transport under a fixed amount of atmospheric interbasin freshwater transport are found to have a minor impact on the global thermohaline circulation. This insensitivity is realized because increased volume transport leads to increased salinity advection from low to high latitudes in the North Pacific and hence causes a salinity increase at the strait. Reducing net freshwater export from the Atlantic sea surface to nearly zero results in shutdown of the Atlantic deep circulation. The actual atmospheric freshwater transport anomaly required to shut the circulation down depends on the configuration of the Bering Strait, and the Atlantic deep circulation shows high sensitivity to the atmospheric freshwater transport around the shutdown.

1. Introduction

Oceanic circulation transports salt and freshwater, which is closely related to the atmospheric freshwater transport, to close the hydrological cycle of the climate system. In the present state of the climate, the annual-mean, basin-averaged net surface freshwater flux, composed of evaporation, precipitation, and river runoff, is positive for the Atlantic and is negative for the Pacific. This contrast is associated with the atmospheric interbasin freshwater transport from the Atlantic to the Pacific, and it helps keep sea surface salinity (SSS) in the Atlantic higher than that in the Pacific. This situation favors the present state of the global thermohaline circulation (THC), where deep water formation is taking place in the northern North Atlantic and around Antarctica. Larger atmospheric transport of freshwater from the Atlantic to the Pacific would raise SSS in the Atlantic and intensify the Atlantic deep circulation, which is associated with the southward extension of North Atlantic Deep Water (NADW).

Estimates of the total export of freshwater from the Atlantic sea surface in the present climate depend a great deal on methods or datasets, ranging from 0.1 Sv (1 Sv = \( 10^6 \text{ m}^3 \text{s}^{-1} \)) to 0.45 Sv (e.g., Broecker et al. 1990; Zaucker and Broecker 1992; Lohmann and Lorenz 2000). It is also controversial how the export is influenced by climate changes. A study with atmospheric general circulation model (AGCM) experiments suggests that the export is reduced to almost zero during the last glacial maximum (18 000 years ago) from the present value of 0.12 Sv (Miller and Russell 1990), while another such study suggests that it is enhanced from the present value of 0.177 Sv to 0.315 Sv (Lohmann and Lorenz 2000). The atmospheric hydrological cycle is expected to be enhanced by CO\(_2\)-induced global warming, which should also increase the export of freshwater from the Atlantic sea surface. An AGCM simulation of an equilibrium state under the doubled atmospheric CO\(_2\) concentration predicts a twofold increase in the export, from 0.13 Sv to 0.25 Sv (Zaucker and Broecker 1992). The atmospheric transport of freshwater from the Atlantic to the Pacific is taking place mainly in the Tropics, across Central America (Zaucker and Broecker 1992). A data analysis study indicates that climate variations associated with El Niño–Southern Oscillation accompanies fluctuations of the transport, whose amplitude is \( \sim 0.1 \text{ Sv} \) (Schmittner et al. 2000). Changes in the atmospheric interbasin freshwater transport in the Tropics by the order of 0.1 Sv are con-
sidered to have a significant impact on the global THC. Studies using zonally averaged ocean models suggest that reduction in the transport by 0.2–0.3 Sv could stop the NADW formation and the associated Atlantic deep circulation (Zaucker et al. 1994; Schmittner et al. 2000). A global warming simulation by Mikolajewicz and Voss (2000), where the freshwater export from the Atlantic sea surface increases at the rate of 0.005 Sv per decade, shows gradual weakening of the Atlantic deep circulation, while in another simulation by Latif et al. (2000), where the rate of increase is 0.015 Sv per decade, the circulation is stably maintained.

Another consequence of the interbasin atmospheric freshwater transport is relatively high sea level in the Pacific. The excess input of freshwater into the Pacific makes surface water less saline and hence less dense in the northern North Pacific than in the Arctic Ocean, which consequently keeps the higher sea level in the Pacific and induces the northward flow through the Bering Strait (Stigebrandt 1984). The Bering Strait throughflow injects relatively less saline North Pacific surface water into the Arctic Ocean, and eventually into the North Atlantic. This should act to reduce intensity of the Atlantic deep circulation, compared to a conceptual situation in which the Bering Strait is closed. A three-dimensional modeling study by Goosse et al. (1997) shows that intensity of the Atlantic deep circulation is reduced by 6% by opening the Bering Strait. Volume transport of the Bering Strait throughflow should also be affected by changes in the atmospheric interbasin freshwater transport. Additional freshwater input into the Pacific would raise the sea level and increase the throughflow transport, which then would affect the global THC by influencing salinity in the formation regions of NADW.

Behavior of the global THC is considered to have a great importance in various kinds of climate problems, such as differences between the present and past states of the climate (Broecker 1997) and possible future climate changes induced by global warming (Manabe and Stouffer 1994). As noted before, the atmospheric interbasin freshwater transport may differ among different states of the climate by an amount large enough to affect the global THC, while the magnitude of the difference is still largely unknown. It is then important to know the sensitivity of the global THC to the atmospheric interbasin freshwater transport to better understand the mechanisms governing climate changes in the past and the future. Three-dimensional modeling studies with idealized basins by Marotzke and Willebrand (1991) and Hughes and Weaver (1994) investigated the role of sea surface freshwater budgets in controlling the global THC. However, their studies do not address the sensitivity of the global THC to the tropical freshwater budget, which seems to have a great importance in the context of past and future climate changes. In addition, their idealized basin geometry excludes the Arctic Ocean and the Bering Strait throughflow. The Arctic Ocean receives a large amount of freshwater (~0.1 Sv) mainly through river runoff (Aagaard and Carmack 1989). The freshwater input into the Arctic Ocean, together with the Bering Strait throughflow, eventually enters the northern North Atlantic, so the Arctic Ocean and its freshwater budget could significantly influence the NADW formation, and hence the global THC.

As for the role of the Bering Strait throughflow, some studies with simple models have indicated that it significantly affects the Atlantic deep circulation and related climatic phenomena. For example, Shaffer and Bendtsen (1994) suggest that deepening the strait by 30% could stop the NADW formation, based on the result of a three-box model. Walin (1985), based on a hypothesis that the present global THC is not in a steady state, discusses how the continued freshening of the North Pacific and resulting increase of the Bering Strait throughflow could regulate climate changes of very long timescales, such as glaciation cycles, by freshening the Arctic Ocean and hence affecting the sea ice cover. However, there have been few three-dimensional modeling studies to check those simple modeling results.

In this study, the sensitivity of the global THC to changes in the atmospheric interbasin freshwater transport and in the Bering Strait throughflow is investigated by using a three-dimensional, coarse-resolution global ocean general circulation model (OGCM) and conducting a fairly large number of numerical experiments. The freshwater flux at the sea surface is externally specified, and SSS restoration to climatology is not applied. This enables us to properly estimate effects of salinity advection in the upper layer on the global THC. The sensitivity to the atmospheric freshwater transport is examined by changing the freshwater flux in the tropical Atlantic and Pacific. Three series of sensitivity experiments are performed with different configurations of the Bering Strait: closed, open, and a fixed amount of the throughflow. Since the Bering Strait is not well represented in a coarse-resolution OGCM, the throughflow is not predicted by the OGCM dynamics, but instead is parameterized using the sea level difference across the strait.

This paper is organized as follows. In section 2, the model and forcing field used in this study and experimental design is outlined. Results are described in section 3. Summary and discussions are presented in section 4.

2. Model and forcing

a. Ocean model

The OGCM used in this study is the free surface version of the Center for Climate System Research (CCSR) Ocean Component Model (COCO). The rigid-lid version of COCO has been applied to various kinds of sensitivity studies of the global THC (e.g., Hasumi and Suginohara 1999; Oka et al. 2001). The free-surface
version model explicitly predicts changes in sea surface height by the method of Killworth et al. (1991). This model gives almost the same result as one with the rigid lid in terms of the global THC under coarse-resolution world ocean modeling. The model domain is global, including the Arctic Ocean, although the northern-most grids are treated as land. The horizontal resolution is ~2.8° both in the zonal and the meridional direction. There are 40 vertical levels, and the layer thickness is variable from 50 m for the top level to 200 m for the deepest level. The model incorporates the uniformly third-order polynomial interpolation algorithm (Leonard et al. 1993) for tracer advection. Isopycnal diffusion (Cox 1987) is applied and the diffusion coefficient is $1 \times 10^3$ m$^2$ s$^{-1}$. The vertical diffusion coefficient is prescribed as a function of depth, varying from $1 \times 10^{-3}$ m$^2$ s$^{-1}$ at the surface to $3 \times 10^{-4}$ m$^2$ s$^{-1}$ at 5500 m with sharp increase between 1000 and 3000 m [case III of Tsujino et al. (2000)]. The coefficients for the vertical and the horizontal eddy viscosity are $1 \times 10^{-4}$ m$^2$ s$^{-1}$ and $2.5 \times 10^5$ m$^2$ s$^{-1}$, respectively. To suppress computational noises appearing in the sea surface height field, horizontal diffusion is applied with a coefficient of 10 m$^2$ s$^{-1}$, and care is taken not to violate tracer conservation.

In the model geometry, the Bering Strait is represented by a four-grid-width (11.2°) gap of 50-m depth. However, the velocity across the strait is not predicted by the OGCM dynamics but is parameterized. When experiments with the closed Bering Strait are conducted, the velocity at the strait is simply set to zero. For experiments with the open Bering Strait, on the other hand, the total transport through the strait $F$ (positive northward) is parameterized by

$$F = \frac{gh}{f} \delta \eta,$$

where $g$ is the gravitational acceleration, $h$ is the depth of the Bering Strait (50 m), $\delta \eta$ is the difference in sea surface height across the strait (averaged over four grids along the strait and positive when the Pacific side level is higher), and $f$ is the Coriolis parameter, following Shaffer and Bendtsen (1994). This transport is uniformly distributed along the Bering Strait.

**b. Sea surface forcing**

The model is driven by seasonally varying sea surface boundary conditions. Wind forcing is the monthly climatology of Hellerman and Rosenstein (1983). Heat flux $Q$ (positive upward) is imposed by the method of Haney (1971),

$$Q = Q_2(T_1 - T^a),$$

where $T_1$ is the sea surface temperature (SST). The apparent atmospheric temperature $T^a$ and the coefficient $Q_2$ are obtained from an AGCM experiment with the prescribed seasonal SST climatology adopted by the Atmospheric Model Intercomparison Project (Gates 1992). AGCM results are preferred to climatological or objective-analysis datasets because of the convenience in matching the resolution of the forcing with that of the OGCM and in taking into account temporal variabilities of the atmosphere for estimation of $Q_2$. The AGCM used here is CCSR/National Institute for Environmental Studies (NIES) AGCM (Numaguti et al. 1997), and the resolution is T42 in the horizontal (equivalent with that of the OGCM used in this study) with 11 vertical levels. Monthly and zonally averaged datasets are constructed both for $T^a$ and $Q_2$. The meridional profiles of $T^a$ and $Q_2$ for February, August, and the annual-mean condition are depicted in Fig. 1. Since the present model does not include a sea ice component, overcooling of water is prevented by simply prohibiting temperature from going below $-2^\circ$C.

As for freshwater flux, the monthly dataset of evaporation ($E$) is obtained from the same AGCM experiment. Precipitation ($P$) is, however, taken from the climatology of Legates and Willmott (1990), since the AGCM estimates too much precipitation at high latitudes, resulting in collapse of the Atlantic deep circulation for control experiments. River runoff ($R$) is de-
derived by giving $E$ and $P$ to the river routing subprogram adopted in the AGCM. In the monthly dataset of net freshwater flux ($W = P + R - E$) to be used to drive the model, a constant value is globally added to make the global integral of the freshwater flux zero. In addition, $0.3 \text{ m yr}^{-1}$ is subtracted from $W$ north of $60^\circ$N and south of $60^\circ$S to represent the effect of brine rejection by sea ice formation, and a constant value is added to $W$ elsewhere to compensate for it. Annual-mean meridional distribution of $W$ for each basin of the Atlantic (including the Arctic Ocean), the Pacific, and the Indian is presented in Fig. 2. The magnitude and the meridional profile of the freshwater flux is similar to those estimated from objective-analysis datasets (Zaucker et al. 1994). The total export of freshwater from the Atlantic sea surface (north of $30^\circ$S, including the Arctic Ocean) is 0.285 Sv.

c. Experimental design

Three series of experiments are conducted: one with the closed Bering Strait, another with the open Bering Strait, and a third with a fixed amount of Bering Strait throughflow. For each series of experiments, a control experiment is conducted by using the sea surface boundary conditions described above. Sensitivity experiments are carried out by changing $W$ uniformly in the tropical Atlantic; the freshwater export from the Atlantic sea surface is increased by $\Delta$ in the latitude band between $2.8^\circ$S and $11.2^\circ$N where the original annual-mean $W$ is positive both in the Atlantic and the Pacific. This anomalous freshwater export is compensated by the uniform change in $W$ in the tropical Pacific. The series of experiments with the closed Bering Strait is hereafter referred to as the C series, and the control experiment ($\Delta = 0$) of the C series is referred to as C-0. Likewise, the series of experiments with the open Bering Strait is referred to as the O series, and its control experiment is O-0; the series of experiments with the fixed Bering Strait throughflow is referred to as the F series, and its control experiment is F-0. The prescribed transport through the Bering Strait in the F series is 0.85 Sv, which approximates the long-term mean northward transport of the case O-0. Two additional experiments are carried out with the open Bering Strait, where the Bering Strait throughflow is parameterized by (1) multiplied by factors of 0.5 and 1.5. These experiments correspond to the cases where the depth of the Bering Strait is decreased and increased, respectively, by 50%. Each case of experiment is initiated from a state of rest and constant temperature and salinity. Note that the initial global-mean salinity, specified as 34.7 psu here, does not change in the course of model integration, as the globally integrated $W$ is always zero. Time integration is continued for 3000 yr. At that time, there is no long-term trend in various physical quantities in the deep ocean. In the results shown in the next section, the data averaged over the last 300 yr are used, unless otherwise noted.

3. Results

a. Control cases

Figure 3 shows the zonally averaged, annual-mean SST and SSS for the case O-0. The modeled meridional profiles are compared with the climatology of Levitus and Boyer (1994) and Levitus et al. (1994). Note that the climatological data are averaged over the top 50 m. Meridional profiles of the modeled SST are in good agreement with the climatology. The modeled SSS is, on the other hand, significantly lower in the Southern Hemisphere and the northern North Atlantic. The difference in the northern North Atlantic is mostly due to the Labrador Sea. The model fails to reproduce intrusion of surface water into it. Apart from that difference, reproduction of the SSS in the Northern Hemisphere looks good, and it seems possible to carry out meaningful sensitivity experiments.

The zonally integrated overturning circulation in the
Atlantic is shown in Fig. 4 for the cases C-0 and O-0. The overall pattern of the circulation (i.e., sinking takes place in the northern North Atlantic and a bottom water comes from the south), is consistent with that of previous coarse-resolution modeling (e.g., Tsujino et al. 2000). The time series of 10-yr-mean cross-equatorial transport of NADW, which is defined here as the maximum value of the Atlantic meridional overturning streamfunction at the equator, for the last 300-yr period are plotted in Fig. 5. In both the cases, the system reaches an almost steady state by the end of the integration, with the overturning circulation oscillating with an amplitude of <1 Sv. The cross-equatorial transport of NADW averaged over the last 300 yr is 16.3 Sv for the case C-0 and 13.5 Sv for the case O-0. On the other hand, the Pacific deep circulation differs little between these cases as shown in Fig. 6, although the shallow overturning circulation in the northern North Pacific existing in the case O-0 is not found in the case C-0.

Horizontal distribution of sea level for the case O-0 is show in Fig. 7. The mean sea level difference across the Bering Strait is 24 cm, and the mean volume transport through the strait is 0.845 Sv, which is within observational and modeling estimates (e.g., Overland and Roach 1987; Coachman and Aagaard 1988). The salinity averaged along the Bering Strait is 32.64 psu, which is also close to observations (Coachman and Aagaard 1988). Since the Bering Strait throughflow transports less saline water from the northern North Pacific to the northern North Atlantic, the opening of the Bering Strait reduces the intensity of the Atlantic deep circulation. This result is consistent with that obtained by Goosse et al. (2000).
et al. (1997). However, the rate of reduction in this study, 17%, is significantly larger than the 6% obtained by them. There are several possible factors causing this difference, such as freshwater forcing and model setup. Some part of the difference may be associated with the way in which the surface freshwater boundary condition is imposed. In addition to the precipitation minus evaporation flux, Goosse et al. imposed weak restoring of SSS to the climatology to keep the simulated salinity field close to the reality, which could dampen the freshening effect of the throughflow on the Atlantic. Another important difference is that a sea ice component is included in their model. In their result, the opening of the Bering Strait enhances sea ice transport through the Fram Strait, reduces melting of sea ice in the Greenland, Iceland, and Norwegian (GIN) Seas by 0.01 Sv, and increases melting in the northern North Atlantic by 0.02 Sv. Reduced melting of sea ice in the GIN Seas suppresses the weakening of the NADW formation, while increased melting in the northern North Atlantic enhances it. It is difficult to say at this stage which effect is more important. Answering the question requires experiments with and without sea ice under the same model setup. Figure 8 shows zonally averaged salinity in the Atlantic for the C-0 and O-0 cases. The salinity of NADW is slightly higher in the C-0 case. For example, salinity averaged along the equator at 2600-m depth, where the maximum is located, is 35.02 for the C-0 case, and 34.98 for the O-0 case.

Meridional overturning circulation for the F-0 case is almost identical to that for the O-0 case, both in the Atlantic and the Pacific. In the O-0 case, volume transport through the Bering Strait seasonally varies by 10%. Salinity of the throughflow also varies with seasonality, both in the O-0 and F-0 cases. However, its amplitude is as small as ~0.1 psu, so the seasonal variation in the freshening effect of the throughflow on NADW is not significant.

b. Sensitivity to atmospheric interbasin freshwater transport

In all of the three series of experiments, the Atlantic deep circulation is intensified with increasing Δ, anomalous atmospheric freshwater transport from the equatorial Atlantic to the Pacific (Fig. 9a). The salinity of NADW also increases with Δ (Fig. 9b). In the O series experiments, larger Δ results in larger northward volume transport through the Bering Strait (Fig. 9c), as is naturally expected. However, the responding change in the throughflow transport is larger than the change in Δ.

Figure 10 shows the zonally averaged salinity distribution in the Pacific for the O-0 case and its difference between the cases Δ = 0.1 Sv and Δ = 0 Sv (former minus latter) for the O series. The anomalous freshwater input to the Pacific induces freshening in the upper 1000 m of the entire Pacific by ~0.1 psu. Density difference between 10°C, 34.5-psu water and 10°C, 34.4-psu water is, for example, ~0.07 kg m⁻³. If this density difference were to persist in the upper 1000 m, the difference in dynamic height would become ~7 cm. Here, the difference in dynamic height between the cases is almost equivalent to that in sea level, because the wind forcing is identical among the cases and there is little difference in the barotropic flow. According to (1), the sea level rise of 7 cm augments the throughflow transport by 0.25 Sv. Sea level actually rises by several centimeters in the North Pacific and the sea level difference across the strait is enhanced to the same degree, which explains the enhanced response of the throughflow transport. That is, anomalous freshwater input to the Pacific (positive Δ) raises sea level in the Pacific by
freshening the upper layer, which eventually increases the throughflow transport.

There is no distinct difference in the sensitivity of the Atlantic deep circulation to changes of $\Delta$ among the three series of experiments, at least when $\Delta$ is not smaller than $-0.2$ Sv. Furthermore, there is little difference in the intensity of the Atlantic deep circulation itself between the O series and the F series, despite the fact that the throughflow transport significantly differs when $\Delta \neq 0$. In the O series, one might expect that a part of the increase in the atmospheric interbasin freshwater transport is canceled by the increase in the throughflow transport, thus reducing the sensitivity compared with the C series. In the results, however, the increase of the throughflow transport is accompanied by higher salinity at the strait (Fig. 11a), despite the mean salinity in the upper layer of the North Pacific, which decreases as $\Delta$ increases. The higher salinity at the Bering Strait under the stronger throughflow is caused by the advection of more saline water from the south. Therefore, the freshening effect of the throughflow on the Atlantic, which is defined as the product of the northward volume transport of the Bering Strait throughflow and the salinity anomaly from 34.7 psu (the global-mean salinity specified by the initial condition), is not so much affected by $\Delta$ as the volume transport when $\Delta$ is not smaller than $-0.2$ Sv (Fig. 11b). In the F series, salinity of the strait drops as $\Delta$ increases because the throughflow transport is kept constant. As a result, the freshening effect is very close between the O series and the F series, so the sensitivity of the Atlantic deep circulation is almost identical between them.

The salinity at the Bering Strait abruptly drops around $\Delta = -0.2$ Sv (Fig. 11a). In the O-0 case, there is a sharp minimum in the zonally averaged SSS in the North Pacific to the south of the strait (Fig. 3b). This is due to the runoff from the Yukon River, and the correspond-
ing feature is also recognized in the climatology of Levitus et al. (1994), although the averaging over the top 50 m in Fig. 3b blurs it. When there is relatively large throughflow transport, salinity advection from the south keeps the salinity high away from the Alaskan coast. However, when the transport becomes very low, as realized at $\Delta = -0.2$ Sv (Fig. 9c), the low salinity water covers the Bering Sea, which causes the abrupt drop of the salinity of the strait and results in the breakdown of the insensitivity of the freshening effect (Fig. 11b). Further reduction of $\Delta$ eventually results in shutdown of the Atlantic deep circulation (Fig. 9a). The threshold value of $\Delta$ depends on the configuration of the Bering Strait: $-0.24$ Sv for the F series, $-0.26$ Sv for the O series, and $-0.29$ Sv for the C series. The intensity of the Atlantic deep circulation shows high sensitivity to the atmospheric freshwater transport around the shutdown. In addition, the shutdown takes place when the net freshwater export from the Atlantic sea surface approaches zero ($\Delta = -0.285$ Sv). While the O series and the F series look very similar to each other when $\Delta$ is not so small, they exhibit different behavior around the shutdown. In the O series, the throughflow transport is nearly zero when the shutdown of the Atlantic deep circulation takes place. Although there is a compensating effect between changes in the throughflow transport and salinity of the strait, difference in the freshening effect of the throughflow becomes significant as the throughflow transport in the O series gets considerably small. This difference results in the higher sensitivity of the Atlantic deep circulation around the shutdown for the F series.

The Pacific deep circulation is relatively insensitive to changes in $\Delta$, compared with the Atlantic deep circulation (Fig. 12). Over the presented range of $\Delta$, the cross-equatorial transport of NADW changes by more than 10 Sv, whereas the deep water transport in the Pacific changes by less than 1 Sv. For the C series and the O series, the cross-equatorial northward transport of the deep water in the Pacific gradually decreases as $\Delta$ decreases, and it takes a value around 9.2 Sv when the Atlantic deep circulation stops. For the F series, on the other hand, the thermohaline circulation becomes different in shape when the Atlantic deep circulation ceases, shallow overturning circulation with downwelling at the northern boundary of the North Pacific, as seen in Fig. 6b, extends deeper and farther southward.

c. Sensitivity to the depth of the Bering Strait

The comparison between the O series and F series in the previous subsection indicates that an absolute
amount of volume transport through the Bering Strait seems to have little significance under the same value of $\Delta$, as long as $\Delta$ is not too small. To further investigate this consequence, results of the two additional experiments, where the parameterized Bering Strait throughflow $F$ is multiplied by factors of 0.5 (the shallow-strait case) and 1.5 (the deep-strait case), are presented here. The throughflow volume transport is enhanced by 27% when the strait is deepened and reduced by 43% when the strait is made shallower. However, the intensity of the Atlantic deep circulation is fairly insensitive to these changes: the change in the cross-equatorial transport of NADW is $\sim$1% for both the cases. The stronger (weaker) Bering Strait throughflow is accompanied by the higher (lower) salinity at the Bering Strait, and the change in the freshening effect of the throughflow on the northern North Atlantic (the anomalous salt transport defined in the previous subsection) is limited only to $\sim$1%. In these results, the change of salinity at the Bering Strait works as a negative feedback against that of the throughflow transport induced by different configurations (depth) of the strait. Thus, the Atlantic deep circulation is kept almost unchanged.

4. Summary and discussion

In this study, sensitivity of the global THC to interbasin freshwater transport by the atmosphere and the Bering Strait throughflow has been investigated by using a free-surface coarse-resolution ocean general circulation model. The Bering Strait throughflow transports less-saline northern North Pacific water to the northern North Atlantic, thus working to reduce the intensity of the Atlantic deep circulation, while the Pacific deep circulation is almost unaffected. The weakening of the Atlantic deep circulation due to the presence of the Bering Strait throughflow obtained in this study is much larger than a previous estimate (Goosse et al. 1997). Increase in the atmospheric freshwater transport from the Atlantic to the Pacific intensifies both the Atlantic deep circulation and the Bering Strait throughflow. When the atmospheric interbasin freshwater transport is kept unchanged, difference in the throughflow transport, caused by different depths of the strait or by artificially fixing the transport, is found to have little significance on the global THC, as long as the Atlantic deep circulation exists with certain high intensity. This is realized due to the compensation between the volume transport and the salinity of the strait; larger transport induces larger advection of relatively saline water to the Bering Strait and raises salinity there, thus suppressing changes in the freshening effect of the throughflow on NADW. Reducing the net freshwater export from the Atlantic sea surface to nearly zero results in shutdown of the Atlantic deep circulation. The actual atmospheric freshwater transport anomaly required to shut the circulation down depends on the configuration of the Bering Strait, and the Atlantic deep circulation shows high sensitivity to the atmospheric freshwater transport around the shutdown.

In the two-dimensional modeling of Schmittner et al. (2000), where an energy–moisture balance atmospheric model is coupled, the Atlantic deep circulation ceases when the atmospheric interbasin freshwater transport from the equatorial Atlantic to the equatorial Pacific is reduced by 0.2 Sv. In the present study, on the other hand, the Atlantic circulation is not shut down, though it is significantly weakened, when the atmospheric freshwater transport is reduced by that amount, regardless of the configuration of the Bering Strait (open, closed, or when the transport is fixed at the present state value). It is not easy to say which result is more realistic, as there are a number of differences in these two studies. It is very likely that the freshwater export from the Atlantic in the control case is different, although the value is not noted by them. Response of the Atlantic deep circulation to changes in the atmospheric hydrological cycle is discussed in various contexts, such as global warming (Latif et al. 2000) and palaeoclimate (Broecker 1997), and shutdown of the Atlantic deep circulation is of primary importance in those discussions. This study indicates that the threshold for shutdown of the Atlantic deep circulation is sensitive to the representation of the Bering Strait throughflow, so proper treatment for it is necessary to examine the behavior of the global THC under climate changes.

In the three-dimensional modeling with idealized basins by Marotzke and Willebrand (1991), the present-style global THC is maintained under zero freshwater transport from the Atlantic to the Pacific. This may be caused by the difference in dealing with the Arctic Ocean. In this study, the Arctic Ocean is explicitly represented and there is $\sim$0.1 Sv of freshwater input to the Arctic Ocean, while it is not represented in their idealized basin geometry. As the series of experiments shown here indicate, such an amount of difference in the freshwater forcing could significantly affect the global THC. If the Arctic Ocean is removed from the
current model, the Atlantic circulation would be maintained with significant intensity even when the freshwater export from the Atlantic sea surface is zero. In this sense, the present results are consistent with theirs and show the importance of the Arctic Ocean and its freshwater budget for the global THC.

In this study, the depth of the Bering Strait little affects the global THC, which disagrees with the result of Shaffer and Bendtsen (1994). In their simple model, salinity in the North Pacific is represented by a single parameter. On the other hand, it is essential to the present result that SSS spatially varies. However, although the present three-dimensional modeling is more realistic than their method, a coarse-resolution OGCM is far from perfect in terms of reproducing SSS distribution.

The relative insensitivity of the global THC, found in this study, to changes of the Bering Strait depth seems qualitatively valid, but a quantitative discussion needs further elaboration.

The result that a steady state is obtained in the present modeling goes against Walin’s (1985) hypothesis, although the present modeling does not necessarily give a decisive conclusion. Though Walin discusses the stability of the global THC, he only shows the instability of equatorially symmetric THC under the fixed freshwater boundary condition, which does not mean that the present-style global THC is unsteady. In addition, even if the real global THC is not steady and the North Pacific is undergoing freshening as Walin hypothesizes, the insensitivity of the freshening effect of the throughflow indicated by this study should affect the suggested mechanism of glaciation cycles.

The sensitivity discussed in this study is likely to be affected by surface forcing and the model’s performance, especially with regard to reproduction of the surface salinity field in the North Pacific. Therefore, further investigation is necessary to conclude the problem discussed here by applying higher resolution and better forcing fields. Furthermore, as stated in the comparison with the result of Goosse et al. (1997), sea ice could significantly affect the sensitivity, so modeling studies with a comprehensive sea ice model are also required.

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