Effects of Freshwater Forcing on the Atlantic Deep Circulation: A Study with an OGCM Forced by Two Different Surface Freshwater Flux Datasets

AKIRA OKA AND HIROYASU HASUMI
Center for Climate System Research, University of Tokyo, Tokyo, Japan

(Manuscript received 13 June 2003, in final form 5 January 2004)

ABSTRACT

Numerical experiments are conducted using a sea ice–coupled ocean general circulation model (OGCM) forced by two different freshwater flux datasets. These two datasets are the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis and the European Centre for Medium-Range Weather Forecasts (ECMWF)-based climatological datasets, which are widely used to force OGCMs. It is found that the strength of the simulated Atlantic deep circulation considerably differs between the two experiments. To explain the resulting difference, these two freshwater fluxes are compared and additional experiments are carried out, focusing on the difference at northern high and midlatitudes, at low latitudes, and in the Southern Ocean, separately. An examination of these experiments shows that the difference in the simulated Atlantic deep circulation comes mainly from the difference in the river runoff data, especially at the northern high latitudes. Although the amount of the difference in the river runoff data at northern high latitudes is small, compared with that of the evaporation and the precipitation in other regions, it has a considerable influence on the strength of the Atlantic deep circulation. It indicates that the strength of the Atlantic deep circulation is affected more significantly by the accuracy of the river runoff data than that of the evaporation and the precipitation data.

1. Introduction

In the present state of the earth’s climate, the deep Atlantic Ocean is filled with North Atlantic deep water (NADW), which is formed at northern high latitudes. This water mass is associated with the existence of the Atlantic deep circulation, the thermohaline circulation which is made up of downwelling at northern high latitudes and upwelling in other regions. Although currents in the deep ocean are very slow compared to those in upper layer, the Atlantic deep circulation carries a large amount of heat to high latitudes because of its large volume transport and enormous heat capacity. This poleward heat transport is believed to play an important role in the present climate.

The thermohaline circulation is driven by distribution of the buoyancy flux through the sea surface, and the downwelling occurs in very narrow regions where high-density water is formed. The density of seawater depends on both temperature and salinity. For the present Atlantic deep circulation, the deep water is formed at northern high latitudes where cooling and freshwater gain occur at the sea surface. This means that sea surface heat flux drives the Atlantic deep circulation while sea surface freshwater flux retards it. There is a negative feedback working between sea surface temperature (SST) and the sea surface heat flux where latent and sensible heat flux and upward longwave radiation depend on SST, but there is not such a feedback between sea surface salinity (SSS) and the sea surface freshwater flux. This implies that a small perturbation of surface freshwater flux may induce great changes in the salinity field and in the Atlantic deep circulation. In fact, the Younger Dryas event is considered to be caused by such freshwater flux perturbation associated with ice sheet melting at northern high latitudes (Broecker et al. 1985). Manabe and Stouffer (1999) demonstrate this by using an atmosphere–ocean coupled model. The investigation into the role of freshwater flux forcing in the Atlantic deep circulation is important for understanding the climate system.

In order to understand the present Atlantic deep circulation, we have to discuss how sea surface freshwater flux affects and determines the present salinity distribution. However, the traditional SSS-restoring boundary condition is not physically valid. In addition, the resulting freshwater flux given to an ocean general circulation model (OGCM) is not necessarily realistic. Then, the SSS-restoring boundary condition does not allow us to evaluate the role of surface freshwater flux appropriately. For understanding the role of freshwater flux forcing in the Atlantic deep circulation in the present climate system, an OGCM simulation under a realistic freshwater flux boundary condition is required.

Until now, only a small number of OGCM studies have investigated the steady state of the global thermohaline circulation without employing the SSS-re-
storing surface boundary condition (e.g., Hasumi 2002; Komuro and Hasumi 2003). They use such boundary conditions to discuss how the thermohaline circulation responds to changes in sea surface freshwater flux. However, there are remaining problems in their experiments. For example, Hasumi (2002) artificially modifies sea surface freshwater flux to represent the effects of sea ice because sea ice is not incorporated in his model. Komuro and Hasumi (2003) carry out ice–ocean coupled model experiments to investigate effects of freshwater transport by sea ice on the global thermohaline circulation. However, they modify the thermal boundary condition because they cannot simulate the Atlantic deep circulation realistically without such a modification. OGCMs forced by observed freshwater flux have not yet successfully simulated the Atlantic deep circulation. There are two aspects to this problem: one is associated with OGCMs, and the other is with the sea surface flux used for the boundary condition. As for OGCMs, Oka and Hasumi (2003, manuscript submitted to J. Geophys. Res., hereafter OH) discuss how representation of water exchanges at narrow passages, such as the Iceland–Scotland passage and the Fram Strait, are important for the salinity balance in the Greenland, Iceland, and Norwegian (GIN) Seas and the dense water formation there. As for sea surface flux, there is a problem as to how reliable freshwater flux forcing data are and how significantly their uncertainty affects the strength of the Atlantic deep circulation in OGCM simulation, because there are large discrepancies among different datasets of precipitation and river runoff.

In this study, we investigate how sensitive the Atlantic deep circulation is within the range of uncertainty in the freshwater flux climatology by using an OGCM driven by two widely used datasets. We try to interpret the resulting difference in the strength of the Atlantic deep circulation from the following standpoints. Freshwater flux at northern high latitudes is known to significantly affect the Atlantic deep circulation. For example, Manabe and Stouffer (1997) demonstrate with an atmosphere–ocean coupled model that an excess freshwater discharge of 0.1 Sv (1 Sv = 10^6 m^3 s^-1) into the northern North Atlantic Ocean causes drastic weakening of the Atlantic deep circulation. Thus, we evaluate how much the strength of the Atlantic deep circulation is sensitive to surface freshwater forcing at northern high latitudes. Although many coupled atmosphere and ocean general circulation models simulate the weakening of the Atlantic deep circulation in global warming experiments, Latif et al. (2000) obtain the result that the Atlantic deep circulation does not weaken. They point out that El Niño–like warming in the eastern equatorial Pacific is simulated and the increased interbasin atmospheric freshwater transport from the Atlantic Ocean to the Pacific Ocean at equatorial regions stabilizes the Atlantic deep circulation. Hasumi (2002) quantifies the sensitivity of the Atlantic deep circulation to the amount of the atmospheric zonal freshwater transport from the equatorial Pacific Ocean to the equatorial Atlantic Ocean by conducting OGCM experiments. So, we also focus on the difference in the freshwater flux contrast between the Atlantic and Pacific Oceans at low latitudes. In addition, previous studies suggest that surface forcing in the Southern Ocean is important for the Atlantic deep circulation. Hasumi and Suginohara (1999) point out that the upwelling branch of the Atlantic deep circulation is localized in the Southern Ocean. They show that the strength of the thermohaline circulation is proportional to the contrast of the sea surface buoyancy flux between its upwelling and downwelling regions. Saenko et al. (2003) carry out sensitivity experiments where moisture eddy diffusivity is changed, especially in the Southern Ocean, by using their model of intermediate complexity. They obtain the result that enhanced poleward moisture flux in the Southern Ocean leads to the intensification of the Atlantic deep circulation. Then, the freshwater flux in the Southern Ocean is expected to have a large impact on the strength of the Atlantic deep circulation. In this study, the difference in freshwater flux in the Southern Ocean is also discussed.

This paper is organized as follows. The model and the boundary conditions used in this study are outlined in section 2. Available datasets of sea surface freshwater flux are listed and those adopted in this study are described in section 3. In section 4, sea ice–coupled OGCM experiments forced by two different freshwater flux datasets are carried out, where we discuss how the Atlantic deep circulation is influenced by the difference in the datasets of freshwater flux. Finally, a discussion and concluding remarks are given in section 5.

2. Model and forcing

The model used in this study is the Center for Climate System Research (CCSR) Ocean Component Model (COCO) coupled with a thermodynamic, dynamic sea ice model. The ocean model is based on that of Hasumi (2002) and is a free-surface model with isopycnal diffusion (Cox 1987), isopycnal thickness diffusion (Gent et al. 1995), and the uniformly third-order polynomial interpolation algorithm (UTOPIA; Leonard et al. 1993) for tracer advection. The vertical diffusivity and viscosity are determined by the level-2 turbulence closure scheme of Mellor and Yamada (1982). The background vertical diffusivity is 1.0 × 10^{-3} m^2 s^{-1} for the top level and increases with depth up to 2.7 × 10^{-4} m^2 s^{-1} for the bottom level, with a sharp increase around 1500 m (Tsuchino et al. 2000). The isopycnal-layer thickness diffusivity, the isopycnal diffusivity, and the background horizontal diffusivity are 5.0 × 10^2, 1.0 × 10^1, and 1.0 × 10^2 m^2 s^{-1}, respectively. The background vertical viscosity is 1.0 × 10^{-4} m^2 s^{-1}, except that this is increased to 1.0 × 10^{-2} m^2 s^{-1} near the surface. The vertical coordinate system is a hybrid of ζ (normalized depth) and z. There are 42 levels in the vertical, where
The σ coordinate is applied to the top five levels (between the free surface and 50 m below the mean sea level). The model spherical coordinate is rotated so that the North Pole is on Greenland in order to avoid the coordinate convergence in the Arctic Ocean. The model horizontal resolution is 2° both in the zonal and the meridional directions, except that the meridional resolution is locally increased to 0.5° at the northern high latitudes. The higher resolution there leads to realistic representation of salinity transport through narrow passages, such as the Iceland–Scotland passage and the Fram Strait, and is necessary for simulating the realistic Atlantic deep circulation OH. The bottom boundary layer (BBL) parameterization of Nakano and Suginoara (2002) is incorporated into the OGCM.

The sea ice model is the same as used by Komuro and Hasumi (2003). The thermodynamic part is the zero-layer model of Semtner (1976). In the dynamic part, internal ice stress is formulated by the elastic-viscous-plastic rheology (Hunke and Dukowicz 1997). The yield ellipse eccentricity is 2 and the ice strength parameter $P^*$ is $5.0 \times 10^3$ N m$^{-2}$. Two-category thickness representation is adopted, where concentration and mean thickness are predicted in each grid. Harmonic and bi-harmonic diffusion terms are added to the advection equation of concentration and thickness for the sake of numerical stability (Hibler 1979). These coefficients are $1.0 \times 10^4 \cos \theta$ m$^2$ s$^{-1}$ and $1.0 \times 10^{14} \cos^3 \theta$ m$^4$ s$^{-1}$ ($\theta$ is the latitude of the model coordinate), respectively.

The model is driven by the surface heat, freshwater, and momentum fluxes, given as monthly mean data. The momentum flux is directly taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis wind stress data (Kalnay et al. 1996). The heat flux is calculated from climatological monthly mean NCEP–NCAR reanalysis sea surface radiation fluxes and surface air properties, where bulk formulas are used for determining sensible and latent heat flux. The values of bulk coefficients are taken from the energy balance model developed by Oka et al. (2001). As for the freshwater flux, the model is directly forced by precipitation ($P$) minus evaporation ($E$) plus river runoff ($R$), and no salinity restoring is employed. For simulating the steady state of the ocean, $P - E + R$ is required to be zero when it is averaged over time and space. However, climatological freshwater flux datasets used in this study do not satisfy this requirement. Therefore, we adjust $P - E + R$ by adding or subtracting a constant value globally so that the global integral of $P - E + R$ becomes zero. Datasets for $P - E + R$ are described in the following section.

In all of the experiments, the model is integrated from temperature and salinity of the Polar Science Center Hydrographic Climatology (PHC; Steele et al. 2001) by using the acceleration method of Bryan (1984) for 5000 yr in the deep ocean. The results averaged for the last 100 yr are used for analysis.

### 3. Datasets of freshwater flux

In this section, we list available datasets of each freshwater flux component and describe the datasets used in this study. The available datasets of the precipitation, evaporation, and river runoff are summarized in Table 1. A reference for each dataset is also listed in this table.

**a. Precipitation**

The precipitation datasets, which cover the entire globe, are compared in Fig. 1. The Global Precipitation Climatology Project (GPCP) and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) are compiled from various observational data such as rain gauge values and indirect estimates from satellite data. Legates and Willmott (1990, hereafter Legates) compiled direct observations from ships for their data. Oberhuber (1988, hereafter Oberhuber) also compiled data from ship observations, but the data are missing in some regions, especially in the Southern Ocean. The NCEP–NCAR reanalysis data are not pure observation, but are dependent on the atmosphere model that is used. The Ocean Model Intercomparison Project (OMIP) dataset is an European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis–based climatology. The disagreement seen in Fig. 1 is especially large in the Tropics, which seems to be due to the fact that the precipitation associated with cumulus convection is difficult to be estimated or modeled.

### Table 1. The datasets of the sea surface freshwater flux. The name of the datasets, components ($P$: precipitation, $E$: evaporation, $R$: river runoff), the method of the dataset construction, and the reference are shown.

<table>
<thead>
<tr>
<th>Data</th>
<th>Components</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMAP</td>
<td>$P$</td>
<td>Various</td>
<td>Xie and Arkin (1997)</td>
</tr>
<tr>
<td>GPCP</td>
<td>$P$</td>
<td>Various</td>
<td>Huffman et al. (1997)</td>
</tr>
<tr>
<td>Legates</td>
<td>$P$</td>
<td>Ship observation</td>
<td>Legates and Willmott (1990)</td>
</tr>
<tr>
<td>NCEP</td>
<td>$P$, $E$</td>
<td>Reanalysis</td>
<td>Kalnay et al. (1996)</td>
</tr>
<tr>
<td>OMIP</td>
<td>$P$, $E$, $R$</td>
<td>Based on ECMWF reanalysis</td>
<td>Röske (2001)</td>
</tr>
<tr>
<td>Perry</td>
<td>$R$</td>
<td>Observation at estuaries</td>
<td>Perry et al. (1996)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Data</th>
<th>Components</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMAP</td>
<td>$P$</td>
<td>Various</td>
<td>Xie and Arkin (1997)</td>
</tr>
<tr>
<td>GPCP</td>
<td>$P$</td>
<td>Various</td>
<td>Huffman et al. (1997)</td>
</tr>
<tr>
<td>Legates</td>
<td>$P$</td>
<td>Ship observation</td>
<td>Legates and Willmott (1990)</td>
</tr>
<tr>
<td>NCEP</td>
<td>$P$, $E$</td>
<td>Reanalysis</td>
<td>Kalnay et al. (1996)</td>
</tr>
<tr>
<td>OMIP</td>
<td>$P$, $E$, $R$</td>
<td>Based on ECMWF reanalysis</td>
<td>Röske (2001)</td>
</tr>
<tr>
<td>Perry</td>
<td>$R$</td>
<td>Observation at estuaries</td>
<td>Perry et al. (1996)</td>
</tr>
</tbody>
</table>

---

The precipitation datasets, which cover the entire globe, are compared in Fig. 1. The Global Precipitation Climatology Project (GPCP) and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) are compiled from various observational data such as rain gauge values and indirect estimates from satellite data. Legates and Willmott (1990, hereafter Legates) compiled direct observations from ships for their data. Oberhuber (1988, hereafter Oberhuber) also compiled data from ship observations, but the data are missing in some regions, especially in the Southern Ocean. The NCEP–NCAR reanalysis data are not pure observation, but are dependent on the atmosphere model that is used. The Ocean Model Intercomparison Project (OMIP) dataset is an European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis–based climatology. The disagreement seen in Fig. 1 is especially large in the Tropics, which seems to be due to the fact that the precipitation associated with cumulus convection is difficult to be estimated or modeled.
b. Evaporation

The datasets for evaporation are shown in Fig. 2. The overall disagreement is smaller than that of precipitation. However, there is a significant difference at low latitudes and in the Southern Hemisphere.

c. River runoff

Table 2 shows the area-integrated river runoff over major basins for data of Perry et al. (1996, hereafter Perry) and OMIP. Perry is compilation of measurements for 981 rivers. On the other hand, in OMIP, the river runoff is estimated from the precipitation and the evaporation on the land, assuming that the excess precipitation flows into the ocean. The values of OMIP are larger than those of Perry in all of the basins. Perry is considered to underestimate them because small rivers are not included. However, this does not indicate that OMIP is more accurate because of the uncertainties in precipitation and evaporation over the landmass.

d. Freshwater flux used in this study

If we try to use the same data source for precipitation, evaporation, and river runoff, the only available dataset is OMIP. As for precipitation and evaporation, Oberhuber and the NCEP–NCAR reanalysis are available but Oberhuber has data-missing regions. The ECMWF and
NCEP–NCAR reanalyses have been used as freshwater flux forcing by many OGCM studies (e.g., Vialard et al. 2002; Komuro and Hasumi 2003; Gulev et al. 2003), and OMIP is to be used in the worldwide Ocean Model Intercomparison Project. Perry includes the most plentiful observational records for 981 rivers among published global climatologies of river runoff. Therefore, we decide to assess the impact of the difference between the following two freshwater flux datasets: one consists of the evaporation and precipitation of the NCEP–NCAR reanalysis and river runoff of Perry (labeled as NP flux, hereafter), and the other consists solely of OMIP.

4. Result

a. Experiment forced by NP flux

In this section, we show the results of the experiment forced by NP freshwater flux (EXP-NP). The simulated temperature, salinity, and meridional overturning circulation in the Atlantic Ocean are shown in Fig. 3. The model realistically simulates two deep circulation cells associated with NADW and Antarctic bottom water (AABW). Because the strength of the Atlantic deep circulation, 12.26 Sv at the equator, is a little weak compared with an observational estimate (14 Sv; Schmitz 1995), the temperature and the salinity in the deep ocean are a little lower than the observed data (Fig. 4). Unrealistic open-ocean convection in the Southern Ocean also contributes to this cooling and freshening in the deep ocean (figure not shown). However, the overall structures of temperature and salinity are well repro-
The model does not seem to simulate effects of brine rejection on the salinity realistically. In reality, saline waters formed by brine rejection are considered to flow down on the continental slopes and make the salinity of the deep ocean high (Aagaard et al. 1985). The model does not reproduce such downsloping flow, and saline waters spread out horizontally at shallow depths. This makes the salinity high at the sea surface but low at the depths compared with the observed data.

The observation suggests that the deep convection in both the Greenland Sea and the Labrador Sea contributes to the formation of the NADW (Dickson and Brown 1994). In the model, the deep convection realistically takes place in the Greenland Sea, and dense waters flow over the Iceland–Scotland ridge and discharge into the Atlantic Ocean as seen in the observation (Fig. 6). On the other hand, there is no deep convection in the Labrador Sea. This seems related to the fact that the model simulates a low salinity in the Irminger basin compared with observation. This less saline water spreads around 50°N in the Labrador Sea and the Atlantic Ocean. This is caused by the large amount of sea ice transport through the Denmark Strait (see following discussion). Another possible reason is that there is a large maximum (about 400 cm yr⁻¹ by annual average) near the Irminger basin in the NCEP–NCAR reanalysis precipitation. Although such a maximum is also seen in Oberhuber and OMIP (about 240 and 180 cm yr⁻¹, respectively), the maximum of the NCEP–NCAR reanalysis is the largest and might be overestimated.

Figure 7 shows the sea ice velocity and net annual production in EXP-NP. The model simulates the characteristic pattern of the sea ice velocity field observed from the satellite (Emery et al. 1997), such as the anticyclonic Beaufort Gyre in the Canadian basin and the Transpolar Drift, which is directed to the Fram Strait.
FIG. 7. Annual mean sea ice (a) velocity and (b) net production in EXP-NP. (a) Unit vector is 30 cm s\(^{-1}\). (b) Contour interval is 50 cm yr\(^{-1}\) and contours are not drawn for values smaller than \(-500\).

via the North Pole. Large net sea ice production takes place along the coasts where there is divergent sea ice flow. Because there are no observed data for the distribution of net sea ice production, we compare our results with the results of the numerical simulation by Hilmer et al. (1998). Both results show that net production takes place in the Kara Sea and the Laptev Sea, and net melting occurs in the GIN Seas. However, in our result, there is large net melting along the east shore of Greenland in the Irminger basin. The sea ice transport through the Denmark Strait is 0.093 Sv in EXP-NP, which is about 10 times larger than 0.011 Sv in Hilmer et al. (1998). Aagaard and Carmack (1989) estimate it to be 0.018 Sv in reality, so our model seems to overestimate it. The rough representation of topography there in our model, where the resolution of the model topography is 2\(^{\circ}\), may cause this. The melting of large amounts of sea ice transported through the Denmark Strait is considered to cause the low salinity in the Irminger basin mentioned above (Fig. 5). On the other hand, the simulated sea ice transport at the Fram Strait is 0.117 Sv. This is close to 0.08 Sv of observed estimate by Aagaard and Carmack (1989).

b. Experiments forced by OMIP flux

In this section, we conduct the experiment forced by the OMIP freshwater flux (EXP-OMIP). The simulated Atlantic deep circulation in EXP-OMIP is shown in Fig. 8. The volume transport at the equator is 6.26 Sv, which is considerably weak compared with that of EXP-NP. Figure 9 shows the SSS difference between EXP-OMIP and EXP-NP. The SSS in EXP-OMIP is decreased in the Northern Hemisphere compared with EXP-NP. In the Southern Ocean, on the contrary, the SSS in EXP-OMIP is higher than that in EXP-NP. The salinity increase in EXP-OMIP also occurs in the center of the Arctic Ocean. The difference between the two freshwater flux datasets, NP and OMIP, given as the boundary condition in EXP-NP and EXP-OMIP, is shown in Fig. 10. The difference is significant, especially in the Tropics, where the precipitation in the ITCZ is stronger in OMIP than in NP. In the Southern Ocean, the freshwater loss of OMIP is larger than that of NP, excluding just around Antarctica. The evaporation of OMIP is larger than that of the NCEP–NCAR reanalysis in the Southern Ocean (see Fig. 2), and on the other hand, the OMIP river runoff from Antarctica is larger than that of Perry. There is a significant difference in the river runoff at northern high latitudes, although this is hard to be seen in Fig. 10.

In the following subsections, we discuss how the difference in freshwater flux shown in Fig. 10 causes the difference in the SSS (Fig. 9) and affects the Atlantic deep circulation. In these experiments, we focus on the difference in the freshwater flux at northern high and midlatitudes (60°–90°N and 30°–60°N), at low latitudes (30°S–30°N), and in the Southern Ocean (30°–90°S), separately.

c. Difference at northern high and midlatitudes

The total amount of freshwater flux at northern high latitudes is shown in Table 3. The largest freshwater
flux is the precipitation both in NP and OMIP. If we compare between the river runoff and the evaporation minus the precipitation \((E - P)\), the river runoff contributes more than \(E - P\) both in NP and OMIP. As for the difference between NP and OMIP, the 0.065 Sv difference in \(E - P - R\) is composed of 0.058 Sv in the river runoff, 0.021 Sv in the evaporation, and \(-0.014\) Sv in the precipitation. This means that the difference between NP and OMIP mainly comes from the difference of the river runoff, although the largest freshwater flux is from the precipitation rather than the river runoff. In order to evaluate the effects of difference in river runoff data, we conduct the experiment where the runoff data of Perry in EXP-NP are replaced globally by that of OMIP (R-GLB). As a result, the volume transport of the Atlantic deep circulation at the equator is decreased to 7.19 Sv (figure not shown), which is almost the same as that in EXP-OMIP. The SSS difference from EXP-NP is shown in Fig. 11. The salinity decrease takes place in the northern North Atlantic Ocean and there is a salinity increase in the center of the Arctic Ocean. Such a change is also seen in Fig. 9.

The total river runoff in the northern North Atlantic Ocean, the GIN Seas, and the Arctic Ocean is shown in Table 4. This shows that the OMIP runoff is larger than Perry in all of the basins at northern high latitudes. The difference in the northern North Atlantic Ocean is largely explained by the river runoff from Greenland, which is 0.016 Sv in OMIP but is 0 Sv in Perry. In OMIP, 0.013 Sv of this runoff discharges into the northern North Atlantic Ocean, 0.001 Sv into GIN Seas, and 0.003 Sv into the Arctic Ocean. Although evaluation of the runoff from Greenland seems difficult, Ohmura and Reeh (1991) estimate it from the freshwater budget over
Greenland. The net freshwater input \((P - E)\) averaged over the Greenland is observed to be 317 mm yr\(^{-1}\), which must be balanced by 0.018 Sv of the iceberg calving or the runoff from Greenland if there are no mass changes in the Greenland ice sheet. The value of 0.018 Sv seems to agree well with 0.016 Sv of OMIP. However, the value is affected by the uncertainty in observations over Greenland and there is no validation that the mass of the Greenland ice sheet is not changed. Moreover, the horizontal distribution of the river runoff from Greenland in OMIP is difficult to validate. On the other hand, the difference in the GIN Seas and the Arctic Ocean is mainly caused by the river runoff from Eurasia.

To evaluate the effects of the river runoff difference shown in Table 4 on the strength of the Atlantic deep circulation and the SSS distribution, we conduct additional experiments where the runoff data of Perry in EXP-NP is replaced by that of OMIP in the northern North Atlantic Ocean (R-ATL), the GIN Seas (R-GIN), and the Arctic Ocean (R-ARC), separately. The volume transport of the Atlantic deep circulation in R-ATL, R-GIN, and R-ARC is 11.69, 11.40, and 11.47 Sv, respectively (figures not shown). This indicates that the river runoff in the northern North Atlantic Ocean, the GIN Seas, and the Arctic Ocean evenly contributes to the weakening of the Atlantic deep circulation. Figure 12 shows the SSS difference from EXP-NP in each experiment. The freshening takes place in the North Atlantic Ocean in all of the experiments due to the reduction in the Atlantic deep circulation. The salinity increase near the coast in the Arctic Ocean is caused by changes in the distribution of the sea ice net production, which is induced by cooling associated with the weakening of the Atlantic deep circulation. In R-ARC, the difference in the river runoff is along the coasts of Greenland and Eurasia, which can be seen in the SSS difference in Fig. 12d. The runoff from Greenland introduces freshwater along the shore of Greenland, which is advected by the East Greenland Current into the GIN Seas. On the other hand, the water freshened by the runoff from Eurasia flows into the GIN Seas at the subsurface and middepths. It is considered that this freshwater advection reduces the convection in the Greenland Sea and weakens the strength of the Atlantic deep circulation. In R-GIN, the runoff difference is also seen along the coasts of Eurasia and Greenland. However, the large amount of the runoff from Greenland discharges into the south of the Greenland Sea, hence, it does not affect the convection in the Greenland Sea as much because the East Greenland Current carries it out of the GIN Seas. On the other hand, the river runoff from Eurasia flows into the GIN Seas at the subsurface and middle depths, which affects the Atlantic deep circulation. Contrary to R-ARC and R-GIN, there is not significant salinity decrease in the GIN Seas in R-ATL. Unrealistic convection takes place in the Irminger basin in the model (Fig. 6). In R-ATL, it is interpreted that the runoff from Greenland influences the Atlantic deep circulation by affecting the convection in the Irminger basin.

When we force the OGCM by the freshwater flux where the runoff data of NP are replaced by that of OMIP in the northern North Atlantic Ocean, the GIN Seas, and the Arctic Ocean all together, the volume transport of the modeled Atlantic deep circulation at the

### Table 3

<table>
<thead>
<tr>
<th></th>
<th>(E)</th>
<th>(P)</th>
<th>(E - P)</th>
<th>(R)</th>
</tr>
</thead>
<tbody>
<tr>
<td>NP</td>
<td>0.143</td>
<td>0.214</td>
<td>-0.071</td>
<td>0.110</td>
</tr>
<tr>
<td>OMIP</td>
<td>0.122</td>
<td>0.200</td>
<td>-0.078</td>
<td>0.168</td>
</tr>
</tbody>
</table>
equator is 9.29 Sv (figure not shown). This value is a little larger than that in R-GLB. It indicates that the runoff difference in regions except for the northern high latitudes also has an influence on the strength of the Atlantic deep circulation. For example, in the Atlantic Ocean, the OMIP runoff is larger than Perry by 0.026 Sv in 30°–60°N, and by 0.173 Sv in 30°S–30°N. Such a difference is thought to affect the Atlantic deep circulation.

We have seen that the contribution of the difference in $E - P$ is small at the northern high latitudes compared with that of river runoff (Table 3). On the other hand, between 30° and 60°N, $E - P$ is also important. For example, in the Atlantic Ocean, the area-integrated $E - P$ between 30° and 60°N is 0.109 Sv in NP and 0.084 Sv in OMIP. The difference between NP and OMIP is 0.025 Sv, which is comparable with the difference in river runoff there. To evaluate the effects of this $E - P$ difference on the Atlantic deep circulation, we conduct the experiment where $E - P$ of the NCEP–NCAR reanalysis in EXP-NP is replaced by that of OMIP between 30° and 60°N (EP-NM). As a result, the volume transport of the Atlantic deep circulation at the equator is 11.54 Sv in EP-NM (figure not shown). Compared with EXP-NP, the reduction in EP-NM is 0.72 Sv. This result confirms that the $E - P$ difference at mid-latitudes affects the strength of the Atlantic deep circulation to some degree. In EP-NM, the weakening of the Atlantic deep circulation is considered to be caused by the salinity decrease in the Mediterranean Sea. Figure 13 shows the salinity difference between EP-NM and EXP-NP at the depth of 680 m. The salinity decrease in the Mediterranean Sea spreads over the Atlantic Ocean. The freshening in the Mediterranean Sea is explained by the $E - P$ difference between OMIP and the NCEP–NCAR reanalysis there. The evaporation of OMIP is smaller than that of the NCEP–NCAR reanalysis (see Fig. 10), which causes 0.012 Sv of the $E - P$ difference in the Mediterranean Sea.

d. Difference in the contrast between Atlantic and Pacific Oceans at low latitudes

The total amount of $E - P$ at low latitudes in the Atlantic Ocean and the Pacific Ocean is shown in Table 5. The contrast between the Atlantic Ocean and the Pacific Ocean is more significant in OMIP than in the NCEP–NCAR reanalysis. If we want to investigate effects of this difference on the strength of the Atlantic deep circulation, it seems straightforward to conduct experiments where $E - P$ of the NCEP–NCAR reanalysis in EXP-NP is replaced by that of OMIP between 30°S and 30°N. However, such an experiment is not appropriate for evaluating effects at low latitudes because the replacement of $E - P$ also affects the freshwater flux at high latitudes. The total amount of $E - P$ between 30°S and 30°N in the Atlantic and Pacific Oceans is 1.303 Sv in the NCEP–NCAR reanalysis and 0.814 Sv in OMIP. The replacement of $E - P$ of the NCEP–NCAR reanalysis by that of OMIP leads to the excess freshwater gain of 0.489 Sv, which must be adjusted by subtracting a constant value globally. Such a procedure brings about the freshwater loss of 0.012 Sv in the Arctic Ocean, which is not negligible because the river runoff difference between Perry and OMIP is 0.019 Sv (see Table 4).

Then, we conduct the sensitivity experiment (EP-L)
Fig. 12. (a) Difference in annual mean river runoff (cm yr\(^{-1}\)) between OMIP and Perry. Annual mean SSS difference from EXP-NP in (b) R-ATL, (c) R-GIN, and (d) R-ARC. (b), (c), (d) Contour interval is 0.1 psu.

Fig. 13. Difference in annual mean salinity between EP-NM and EXP-NP at the depth of 680 m (the former minus the latter). Contour interval is 0.02 psu.
where \( E - P \) between 30°S and 30°N in EXP-NP is modified so that this is increased by 0.31 Sv in the Atlantic Ocean and decreased by 0.31 Sv in the Pacific Ocean. Such a procedure corresponds to enhancing the atmospheric freshwater transport from the Atlantic to the Pacific Ocean so that the \( E - P \) contrast between the Atlantic and Pacific Oceans in EP-L becomes the same as that in EXP-OMIP. As a result, the volume transport of the Atlantic deep circulation is larger than EXP-NP by 3.48 Sv at the equator (figure not shown).

Hasumi (2002) conducts sensitivity experiments where the atmospheric freshwater transport from the Atlantic to the Pacific Ocean is changed in the Tropics. The sensitivity of the volume transport of the Atlantic deep circulation in this study is lower than that of Hasumi (2002). The difference in the sensitivity is considered to come from the difference in the modification area of the freshwater flux. The modification area is between 30°S and 30°N in EP-L, but between 2.8°S and 11.2°N in Hasumi (2002). The water at low latitudes is subducted into the subsurface around 30°N by the Ekman downwelling and is advected to high latitudes by the wind-driven gyre. This accounts for a considerable part of the oceanic meridional water transport to the deep-water formation regions (Oka et al. 2001). The reduced sensitivity seems to be explained by the difference in the amount of the freshwater transported to the deep-water formation regions.

It should be noted that the difference in the freshwater flux contrast at low latitudes between the Atlantic and the Pacific Oceans is not entirely accounted for by the difference in the atmospheric zonal freshwater transport. This could also be caused by the difference in the atmospheric meridional freshwater transport. For a more reliable discussion about atmospheric freshwater transport between the Atlantic and Pacific Ocean, the difference in the atmospheric vapor transport between the NCEP–NCAR reanalysis and OMIP must be investigated.

e. Difference in the Southern Ocean

Table 6 shows the total amount of the freshwater flux between 90° and 30°S and between 30°S and 30°N. The difference between the NCEP–NCAR reanalysis and OMIP can be interpreted that the atmospheric meridional freshwater transport from low latitudes (30°S–30°N) to southern high latitudes (90°–30°S) is stronger in the NCEP–NCAR reanalysis than OMIP. In order to evaluate the effect of this difference on the strength of the Atlantic deep circulation, we carry out the experiment where \( E - P \) in EXP-NP is modified so that it is decreased by 0.45 Sv between 30°S and 30°N and increased by 0.45 Sv between 90° and 30°S (EP-S). As a result, the volume transport of the Atlantic deep cir-
Fig. 15. Annual mean SSS difference between EP-S and EXP-NP (the former minus the latter). Contour interval is 0.3 psu.

Table 7. Here, ΔFwfx is total amount of modified atmospheric vapor transport from EXP-NP in EP-L and EP-S, and total amount of modified freshwater flux from EXP-NP in other experiments; ΔVtpt indicates the change of the volume transport of the simulated Atlantic deep circulation at the equator; ΔVtpt/ΔFwfx indicates sensitivity of the simulated Atlantic deep circulation on the freshwater flux difference.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Description</th>
<th>ΔFwfx (Sv)</th>
<th>ΔVtpt (Sv)</th>
<th>ΔVtpt/ΔFwfx</th>
</tr>
</thead>
<tbody>
<tr>
<td>EXP-OMIP</td>
<td>Difference between NP and OMIP</td>
<td></td>
<td>-6.00</td>
<td></td>
</tr>
<tr>
<td>R-GLB</td>
<td>Global runoff</td>
<td></td>
<td>-5.07</td>
<td></td>
</tr>
<tr>
<td>R-ARC</td>
<td>Runoff in the Arctic Ocean</td>
<td>+0.019</td>
<td>-0.84</td>
<td>45.16</td>
</tr>
<tr>
<td>R-GIN</td>
<td>Runoff in the GIN Seas</td>
<td>+0.023</td>
<td>-0.86</td>
<td>37.23</td>
</tr>
<tr>
<td>R-ATL</td>
<td>Runoff in the northern North Atlantic Ocean</td>
<td>+0.017</td>
<td>-0.57</td>
<td>34.13</td>
</tr>
<tr>
<td>EP-NM</td>
<td>$E - P$ at northern middle latitudes</td>
<td>-0.065</td>
<td>-0.72</td>
<td>11.01</td>
</tr>
<tr>
<td>EP-L</td>
<td>$E - P$ at low latitudes</td>
<td>0.310</td>
<td>+3.48</td>
<td>11.23</td>
</tr>
<tr>
<td>EP-S</td>
<td>$E - P$ in the Southern Ocean</td>
<td>0.450</td>
<td>-2.08</td>
<td>4.62</td>
</tr>
</tbody>
</table>

5. Discussion and concluding remarks

In this paper, we have conducted sea ice–coupled OGCM experiments forced by two different climatological freshwater fluxes and discussed how the reproducibility of the Atlantic deep circulation is affected by the difference in the freshwater flux datasets.

In EXP-NP, where the freshwater flux consisting of $E - P$ of the NCEP–NCAR reanalysis and $R$ of Perry is used as the boundary condition, the volume transport of the simulated Atlantic deep circulation at the equator is 12.26 Sv. On the other hand, in EXP-OMIP, where the freshwater flux of OMIP is used as the boundary condition, its volume transport is 6.26 Sv. These results indicate that the difference in freshwater flux significantly affects the Atlantic deep circulation. Although the NP forcing yields a more realistic state of the deep ocean, it does not necessarily mean that the NP forcing is more realistic than the OMIP forcing, due to imperfection of the model used. In this study, we just intend to evaluate the sensitivity of the Atlantic deep circulation to the freshwater flux within the range of uncertainty in the observed freshwater flux.

We conduct a series of experiments to discuss the difference in the simulated Atlantic deep circulation by focusing on the difference in the two freshwater flux datasets at northern high and midlatitudes, at low latitudes, and in the Southern Ocean, separately. These results are summarized in Table 7. The river runoff in the Arctic Ocean and the GIN Seas affects the strength of the Atlantic deep circulation by altering the convection in the Greenland Sea, which in the northern North Atlantic Ocean influences the convection in the Irminger basin. The atmospheric vapor transport at low latitudes between the Atlantic and Pacific Oceans influences the
SSS in the Atlantic Ocean, which is advected toward the high latitudes and affects the Atlantic deep circulation. The meridional atmospheric vapor transport influences the sea surface buoyancy flux around 40°S in the Atlantic sector of the Southern Ocean where the upwelling of the Atlantic deep circulation takes place, which affects the strength of the Atlantic deep circulation.

The sensitivity of the Atlantic deep circulation to the freshwater flux difference is shown in Table 7. The sensitivity to the difference in the river runoff data at the northern high latitudes is considerably larger than that to the difference in $E - P$ in the other regions. It has been pointed out by many studies that the strength of the Atlantic deep circulation is affected by the freshwater flux at northern high latitudes more significantly than that in other regions (e.g., Rahmstorf 1995; Manabe and Stouffer 1997). In the freshwater budget at northern high latitudes (see Table 3), the river runoff contributes more than $E - P$. In addition, the difference in the two freshwater flux datasets dominantly comes from the difference in the river runoff. These results suggest that the Atlantic deep circulation is influenced more significantly by the accuracy of river runoff data than that of evaporation and precipitation data. In this study, we discuss the difference between the NP and OMIP flux. However, it should be noted that the difference between these two and other fluxes is also large. For example, the area-integrated precipitation at northern high latitudes is 0.172 Sv in CMAP, 0.219 Sv in GPCP, 0.198 in Oberhuber, and 0.234 Sv in Legates. The maximum difference between these fluxes is 0.062 Sv. Such a difference will have a great influence on the strength of the Atlantic deep circulation, considering that the difference in the river runoff data between Perry and OMIP is 0.058 Sv there (see Table 3). Although we focused our attention only on the comparison of two reanalysis-based climatological datasets, the difference between other available freshwater flux climatologies implies that the uncertainty of the precipitation data might have the same magnitude of influence on the Atlantic deep circulation as that of the river runoff data.

It should be mentioned that the influence of the procedure to close the global freshwater budget on the strength of the simulated Atlantic deep circulation is not negligible. For simulating the steady state of the ocean, the freshwater flux needs to vanish when it is averaged over time and space, but the freshwater flux used in this study does not satisfy this requirement (Table 8). Therefore, we adjusted the freshwater flux by adding or subtracting a constant value globally so that the global integral of freshwater flux becomes zero. For example, the value of this adjustment is 3.39 cm yr⁻¹ in NP, which causes the freshwater input of 0.010 Sv into the Arctic Ocean. This is not negligible, considering that the total amount of $E - P$ in the Arctic Ocean is −0.042 Sv. Moreover, in OMIP, the freshwater input into the Arctic Ocean by such adjustment is −0.006 Sv. Then, the difference of this adjustment between NP and OMIP becomes 0.016 Sv, which is almost comparable to the runoff data difference between Perry and OMIP in the Arctic Ocean (see Table 4). We should recognize that the imbalance of the freshwater flux data leads to the nonnegligible contribution to the freshwater budget at northern high latitudes.

The two datasets for evaporation and precipitation used in this study, the NCEP–NCAR reanalysis and OMIP (the latter is based on ECMWF reanalysis), are widely used in the ocean modeling community. The difference between such two datasets affects the Atlantic deep circulation considerably. As for the river runoff data, we used Perry and OMIP. In Perry, rivers smaller than the 981 major rivers are not included. On the other hand, OMIP depends on the precipitation and the evaporation over the land, which could reflect the bias of the model used in the reanalysis. Both data seem to contain sources of errors. In this study, the difference of river runoff data considerably affects the Atlantic deep circulation. The importance of river runoff on the water budget in the Arctic Ocean has already been pointed out by some studies (e.g., Aagaard and Carmack 1988). This study confirms this by sea ice–coupled OGCM experiments and demonstrates that the difference of river runoff data has great influence on the Atlantic deep circulation. The uncertainty of river runoff data should be recognized in simulating the Atlantic deep circulation by using OGCMs forced with freshwater flux, and further investigation into the role of river runoff is required for understanding the formation and maintenance mechanism of the present Atlantic deep circulation. In addition, even now, many coupled atmosphere and ocean general circulation models cannot reproduce the realistic Atlantic deep circulation without so-called flux adjustment (Manabe and Stouffer 1988). The results of this study imply that we must pay much attention to simulated river runoff in reproducing the Atlantic deep circulation realistically in coupled models.

Acknowledgments. We are indebted to Prof. Masahiro Endoh and Dr. Nobuo Sugino-hara for their helpful comments and discussions. We also thank Dr. Ryo Furue, Dr. Hideyuki Nakano, and Mr. Yoshihiko Komuro for their fruitful suggestions and discussions. Comments by M. H. England and an anonymous reviewer helped improve our manuscript. The figures are produced with the Dennou Library, developed by the GFD-Dennou Club.
REFERENCES


