Transient Influence of the Reduction of Deepwater Formation on Ocean Heat Uptake and Heat Budgets in the Global Climate System

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Abstract The formation and spreading of dense deepwater in the polar regions play a key role in one of the most important climate systems, namely ocean meridional overturning circulation, and the deepwater formation is projected to decrease under the global warming. However, the impact of the reduced deepwater formation on the climate system has not been explored in detail. Here, we performed a series of numerical experiments with a climate model where the downward water mass transport through the bottom boundary layer is artificially reduced to quantitatively evaluate its impacts on the transient ocean and climate responses. It is demonstrated that changes in deepwater formation have non-negligible impacts on not only ocean heat content but also the Earth's radiation budget at the top of the atmosphere: reduction in deepwater formation in high-latitude oceans causes warming of bottom water, cooling of the ocean surface, and a subsequent decrease in outgoing longwave radiation.

Plain Language Summary The sinking and spreading of cold, dense water into the ocean deep layers at high latitudes plays a crucial role in large-scale ocean circulation, closely linked to the climate system. In this study, we use a climate model to investigate the impact of the reduced dense deepwater formation on the heat budget of the climate system, aiming at a comprehensive understanding of deepwater formation in a warming climate transition. The results show that changes in the dense water formation have a non-negligible effect on the heat budget of the atmosphere as well as on the heat content of the ocean. A decrease in deepwater formation in the high-latitude oceans leads to warming of the bottom waters, cooling the ocean surface, and a concomitant decrease in outgoing longwave radiation.

1. Introduction

The ocean, which plays a dominant role in the thermal inertia of Earth's climate system, has been absorbing excess heat as the globe has warmed, suppressing rapid increases in the surface temperature (e.g., Gregory & Forster, 2008, Raper et al., 2002). More than 90% of the heat energy accumulated in the Earth system in recent decades due to global warming is stored in the ocean (e.g., Church et al., 2013; Levitus et al., 2012). While most of the heat gain is retained in the surface layers of the ocean, a significant amount of heat transport to deeper layers is also predicted. Using a data assimilation approach, Kouketsu et al. (2011) estimated the change in heat content below 3,000 m to be $0.810^{+22}_{-16}$ J/decade (25.4 TW) during the 1990s–2000s, which is approximately 0.05 W/m² for the global average. This value is 5%–10% of the excess heat input (0.5–1 W/m²) at the top of the atmosphere (TOA; Trenberth et al., 2014). Also, Kawano et al. (2010) evaluated the heat input below 3,000 m to be 5% of the total in the Pacific Ocean during 1999–2007 based on observational data. Purkey and Johnson (2013) estimated that heat transport to the deep ocean during the recent warming of the Antarctic Bottom Water (AABW) south of 30°S is $34 \pm 14$ TW below 2,000 m and $5 \pm 2$ TW below 4,000 m. AABW origin water covers 58% of the global ocean floor (Johnson, 2008). Here, the deepwater produced in the Southern Ocean is referred to as AABW.

The heat transport from the ocean surface to the deeper layers affects the ocean heat uptake efficiency (i.e., transient climate response) during global warming (Gregory et al., 2015). Changes in large-scale ocean circulation not only can redistribute ocean storage temperatures but also increase the ocean's total heat content (Garuba & Klinger, 2016). On the global vertical heat transport in the ocean, the heat transport by advection is downward in the upper ocean, balancing upward diffusion, but the direction changes in the deep ocean (e.g., Gregory, 2000; Saenko et al., 2015). Gregory (2000) also states that the deepwater is formed at high latitudes but upwells in low latitudes, where its cooling effect is balanced by the downward diffusion. Reduction of deepwater formation at high latitudes results in reduced upwelling of cold water at low latitudes, giving a net heat uptake. In the CMIP5...
global warming experiments, many climate models have found that the formation of deepwater, such as NADW and AABW, is weakened with a consequent reduction in the flow to the deep layers (Heuzé et al., 2015). These deepwater formations connect to the lower branch of the Pacific and Atlantic meridional overturning circulation (PMOC and AMOC, respectively). Hu et al. (2020) have shown that the mean strength of AMOC and the associated ocean stratification significantly impact the heat uptake in the transient climate response. However, it is still unclear how the weakening of deepwater formation at high latitudes affects the climate sensitivity and ocean heat uptake and feeds back to the radiative forcing during global warming. This study aims to clarify the transient response of the global heat budget to the decrease in deepwater formation in a climate model. There are two critical processes for deepwater formation: the first step is the dense water formation by cooling and mixing around the continental shelf and marginal seas. The second step is the sinking process of dense water into deeper layers along the bottom slope. This study focused on the second process, which links the surface climate system to the deep ocean layers.

The vertical and horizontal resolution of the global ocean component for the general climate models is insufficient to reproduce the down-slope flow of dense water associated with deepwater formation (e.g., Riemenschneider & Legg, 2007). Therefore, many climate models introduce parameterization for a bottom boundary layer (BBL; e.g., Beckmann & Döscher, 1997) to reproduce the down-slope flow (e.g., Boucher et al., 2020; Griffies et al., 2012; Jungclaus et al., 2013; Tatebe et al., 2019). The BBL model controls deepwater formation by transporting dense water produced around the northern North Atlantic and Antarctica from the surface to deeper layers. We used a climate model to evaluate the transient impact of reduced deepwater formation on the heat budget of the climate system by adjusting the thickness of the BBL to reduce the amount of dense water transported to the deep ocean. The response to changing the BBL thickness can be seen as a drift toward a steady-state, but it can also be considered as a part of the transient response to global warming. The novelty of this study is to focus separately only on the effects of the reduction in deepwater formation among the changes in the global heat budget associated with global warming.

2. Methods

This study used the Model for Interdisciplinary Research on Climate version 5.2 (MIROC5.2; Tatebe et al., 2018). This model corresponds to the physical core of MIROC Earth System version 2 for Long-term Simulations (Hajima et al., 2020), which is participating in CMIP6. The atmospheric component has a horizontal resolution of a T42 spectral truncation (approximately 2.8° intervals) with 40 vertical layers up to 3 hPa. The ocean component has a tri-polar horizontal coordinate system with a nominal horizontal resolution of 1°. The vertical levels number 62 with partial step representation in the z-co-ordinate.

To reproduce the down-slope flow, a BBL model was introduced (Text S1 in Supporting Information S1; Nakano & Suginochara, 2002). The model assumes a “slab” layer with constant thickness at the bottom of each water column, and the horizontal advection of tracers is realized between the BBL grids. Vertically, the BBL grid interacts with the upper OGCM grid through vertical diffusion and vertical advection in the normal way. The conservation of the tracer quantities and water volumes is guaranteed even after the BBL parameterization is introduced. The flow in the BBL model is driven by the pressure gradient in a sloping direction as follows:

\[
\begin{align*}
\frac{\partial}{\partial x} \left[ u \right]_{z=H} &= -\left(\frac{1}{\rho_0} \frac{\partial P}{\partial x} + \rho g \frac{\partial H}{\partial x} \right) + V_u = a u, \\
\frac{\partial}{\partial y} \left[ v \right]_{z=H} &= -\left(\frac{1}{\rho_0} \frac{\partial P}{\partial y} + \rho g \frac{\partial H}{\partial y} \right) + V_v = a v,
\end{align*}
\]  

(1)

where \( u \) and \( v \) are the horizontal velocity in the BBL, \( P \) is the pressure, \( H \) is the depth of the seafloor, \( \rho_0 \) is the reference density, \( \rho_g \) is the density in the BBL, \( V \) is the harmonic viscosity term, \( f \) is the Coriolis term, \( g \) is the acceleration due to gravity, and \( au \) and \( av \) are the Newtonian drag terms, that represent the effect of baroclinic eddies which allows flows on a slope to deviate from geostrophic contours (lines of constant \( f/H \)). When the value of \( \alpha \) is \( f \), dense water sinks into the deep layer at an angle of about 45° downslope by the pressure gradient in the downslope direction. Meanwhile, if the value of \( \alpha \) is 0, the currents flow along the isobath line as an approximation. To adjust the depth of the deepwater distribution to observations, we set \( \alpha \) to \( f \) in regions shallower than...
2,000 m in the Northern Hemisphere (NH) and 4,000 m in the Southern Hemisphere (SH), and zero below these depths. The conservation of the fluid volume in the BBL is represented by the continuity equation:

$$\nabla \cdot u = 0,$$

The tracer equations in the BBL are as follows:

$$T_i = -\nabla \cdot (Tu) + D(T),$$

$$S_i = -\nabla \cdot (Su) + D(S),$$

where, $T$ is the potential temperature and $S$ is the salinity in the BBL, and $D$ indicates diffusion term. The vertical diffusion at the upper boundary of the BBL grids represents the effect of entrainment taking place at the top of the turbulent boundary. The exchange of water masses between the BBL and the ocean interior layer takes place through a vertical velocity across the upper boundary of the BBL ($w^*$), which is defined as the horizontal convergence as follows:

$$w^* = -h \times \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right),$$

where, $h$ is the thickness of the BBL and is set to 100 m following Nakano and Suginohara (2002). Here, positive $w^*$ indicates the flow out of the BBL to the ocean interior, and negative $w^*$ indicates the flow into the BBL from the ocean interior (Text S2 in Supporting Information S1). The downward water mass transport through the BBL (DMTB) at depth $Z$ is the area integration of $w^*$ below $Z$ as follows:

$$\text{DMTB}(Z) = \int \{w^* | -H < z < Z\} ds,$$

where, $s$ is the area of BBL grids, and DMTB is proportional to the thickness of the BBL and the velocity in the BBL driven by the pressure gradient. Here, the DMTB is treated as a deepwater formation. The BBL model is configured in an ad-hoc manner to reproduce the observed ocean stratification structure. For example, the BBL grids are added under the lowest layer north of 49°N and south of 54°S (Text S1 in Supporting Information S1). However, the BBL scheme is helpful to investigate the impact of changes in the water mass transport to deeper layers.

In this study, after 1000 years of spin-up under preindustrial conditions, we extend the integration for 70 years with a BBL 100 m thick (control experiment) and also integrate for 70 years each with BBLs 10, 20, and 50 m thick (b10, b20, and b50 m experiments, respectively) to investigate the influence of deepwater formation on the climate system. The initial dynamical shock on thinning the BBL was adjusted on the time scale of the barotropic response. To see the impact of the reduction of only AABW formation, we conducted an experiment in which the thickness of the BBL was set to 20 m in the SH and 100 m in the NH (SO_b20 m experiment).

### 3. Results

Focusing on the water mass exchanges between the BBL and the ocean interior, water masses are taken into the BBL at the top of the slope, and then, the masses are discharged from the BBL into the ocean interior at the lower part of the slope (Text S2 in Supporting Information S1). In the SH, water is taken into the BBL at a depth of about 1000 m (i.e., on the continental shelf), transported through the BBL, and discharged at depths below 4,000 m (Figure 1a). In the NH, in contrast, water is discharged around depths of 2,000 m or more (Figure 1b). This difference is due to the above-mentioned ad-hoc tunings of $\alpha$. The amount of DMTB across 2,500 m in the SH and the NH for the control case is about 4.3 and 4.6 Sv, accounting for about 53% and 61% of the estimated transport of AABW and NADW, respectively, from the observed dissolved chlorofluorocarbons (Orsi et al., 2001). For the control experiment, the intensities of PMOC and AMOC, quantified by northward transport of AABW in the Pacific across 10°S and the maximum southward transport of NADW, are 8.34 and 16.49 Sv, respectively. These amounts are within the uncertainty of observational estimates of 11 ± 5.1 Sv and 17.6 ± 3.1 Sv, respectively (Lumpkin & Speer, 2007).
Under the assumption that the initial state of the ocean remains substantially unchanged, reducing the thickness of the BBL decreases the amount of deepwater formation. When the BBL thickness is changed to 50 m, 20 m, and 10 m, the DMTB is respectively 72%, 47%, and 26% of the control DMTB at 2,500 m in the SH and 58%, 30%, and 17% at 2,000 m in the NH, on average for years 1–70 (Figures 1a and 1b). Because the change in the density structure of the interior region modifies the pressure gradient along the BBL, the rate of change of the DMTB is not strictly proportional to the change in BBL thickness. As the upstream DMTB decreases, the corresponding downstream PMOC and AMOC are weakened relative to the control case (contours in Figure 2).

To investigate the impact of deepwater formation on vertical ocean heat transport, we calculated the global ocean downward heat transport (DHT) as in previous studies (e.g., Gregory, 2000) as follows:

\[
DHT(Z) = \int_{-H}^{H} \int_{-A}^{A} \rho_0 C_p \nabla \theta \, dA \, dz = \int_{-H}^{H} \int_{-A}^{A} \rho_0 C_p (Adv(\theta) + Diff(\theta)) \, dA \, dz
\]

where, \( \theta \) is the tendency of potential temperature, \( \rho_0 \) is the water reference density, \( C_p \) is the heat capacity, and the surface integral is taken over the total ocean area \( A \) at depth \( z \). DHT is separated into the components advection (\( \text{DHT}_{\text{Adv}} \)) and diffusion (\( \text{DHT}_{\text{Diff}} \)), as shown on the right-hand side of Equation 8. \( \text{DHT}_{\text{Adv}} \) is calculated from the resolved velocity and the parameterized eddy-induced velocity in MIROC5.2. Meanwhile, \( \text{DHT}_{\text{Diff}} \) is a sum of all the diffusive processes, including isopycnal diffusion. The vertical profile of the \( \text{DHT}_{\text{Adv}} \) and \( \text{DHT}_{\text{Diff}} \) for the control experiment (Figure S5c in Supporting Information S1) is consistent with the previous studies (Text S3 in Supporting Information S1). Because the DMTB carries cold seawater to deeper layers in the high latitude (Text S2 in Supporting Information S1) and causes the compensative upwelling in lower latitude, its weakening results in a net DHT anomaly that warms the lower layers as shown in Figure 1c. In this process, \( \text{DHT}_{\text{Adv}} \) contributes to the increase in DHT (Figure 1d), while \( \text{DHT}_{\text{Diff}} \) slightly suppresses the increase (Figure 1e). The DHT anomaly across the 2,000 m depth relative to the control case is 23.7 TW in the b50 m experiment, 43.1 TW in b20 m, and 48.1 TW in b10 m on average for years 1–70. At 4,000 m, the DHT anomaly is 5.4 TW in b50 m, 7.7 TW in b20 m, and 7.8 TW in b10 m. These changes are significant enough to be larger than the standard deviation of the interannual variation of the control case. For BBL thicknesses from 50 to 10 m, the ratio of the decrease in DMTB and the increase in DHT with respect to the control experiment is 5.4–6.0 TW/Sv at 2,000 m and 1.5–3.7 TW/Sv at 4,000 m. As compensation for the decrease in DMTB, the upwelling in the ocean interior is also weakened. The magnitude of the heat flux due to the upwelling depends on the strength of the thermal stratification, which may explain the large DHT around 1000 m. These DHT changes are also comparable with the recent warming of the deep layers, as mentioned in the introduction.

Figure 1. DMTB (Sv) in the (a) Southern and (b) Northern Hemispheres; (c) the global downward heat transport anomalies (TW) relative to the control case; and their (d) advective and (e) diffusive components on average for years 1–70. (The hatch shows the range of the standard deviations for the Control.)
Next, we show the results for the b20 m and SO_b20 m experiments relative to the control case to examine the role of deepwater formation in the NH and SH. For SO_b20 m, the DHT anomaly relative to the control case is 18.9 TW at 2,000 m and 8.4 TW at 4,000 m (orange dotted line in Figure 1c). The difference between the b20 m and SO_b20 m experiments is the contribution of the deepwater formation in the NH. This means that the decrease in DMTB in the NH (in the SH) explains about 60% (40%) of the increase in the DHT for b20 m at 2,000 m, while at 4,000 m, the decrease in DMTB in the SH mostly dominates the decrease in the DHT (orange solid and dotted lines in Figure 1c). We can see the warming of the water masses associated with AABW along the bottom around the Antarctic in both experiments (Figures 2d and 2g). The warming of the deepwater is also seen in the Pacific, with a weakening of the PMOC. The northward flow across 10°S decreases by about 1.1 Sv for b20 m and 0.9 Sv for SO_b20 m (Figures 2e and 2h). In the Atlantic basin, we can see strong cooling around the Greenland, Iceland, and Norwegian (GIN) Seas and warming along the slope of the ocean bottom in the North Atlantic for b20 m (Figure 2f). This contrast in temperature change can be attributed to a reduction in the amount of NADW entering the bottom of the North Atlantic Ocean over the sill from the GIN Seas as the result of a thinner BBL. In addition, the southward transport of AMOC across the equator also decreases by about 1.6 Sv, from 13.9 Sv (control) to 12.3 Sv (b20 m), accompanying the warming along with the southward flow. We also see a cooling of deepwater in the South Atlantic. This cooling may be due to the weakening of the southward AMOC, which could increase the northward intrusion of AABW. These changes in the Atlantic Basin are not seen in the SO_b20 m experiment. Our results suggest that the formation of NADW is responsible for the significant heat absorption in the Atlantic. However, in the climatic field of the Atlantic Ocean in MIROC5.2, NADW origin water appears to be more dominant over AABW origin water than the observational estimate by Lumpkin and

![Figure 2. Zonally averaged temperature (colors) for the control experiment (left), and their anomaly (colors) for the b20 m (middle), and the SO_b20 m (right) cases relative to the control in each basin (Global, Pacific, and Atlantic sections) for years 51–70 and the meridional overturning circulation (black contours) for years 1–70. The contour interval is 2 Sv, a dashed contour indicates a negative value, and the interval between thick contours is 10 Sv.](image-url)
Speer (2007), which may lead to an overestimation of the response to NADW formation. However, the formation of the AABW around Antarctica plays a dominant role in the DHT to a depth below 4,000 m, especially in the Pacific.

To investigate the impact of the weakening of deepwater formation on the climate system, we evaluated the time series of the global-mean surface air temperature (SAT) anomalies (Figure 3a) and the global-mean TOA net downward radiative flux (DRF) anomalies (Figure 3b) relative to the control case. Both seem to reach a steady-state after roughly the 30th year while the deep ocean continues to warm. The global-mean SAT for years 51–70 is lower than the control case by about 0.09 K for the b50 m, 0.14 K for the b20 m, 0.14 K for the b10 m, and 0.05 K for the SO_b20 m experiments. Meanwhile, the global-mean DRF at the TOA is higher than the control by 0.04 W/m² for b50 m, 0.08 W/m² for b20 m, 0.12 W/m² for b10 m, and 0.03 W/m² for SO_b20 m on average for years 1–70. These excess heat inputs can continuously increase the globally averaged ocean temperature (Figures 3c and 3d). As intended in the experiments of this study, the DMTB of each experiment is decreasing compared to the control experiment and is almost fairly stable for 70 years (Figures 3g and 3h). Meanwhile, the meridional overturning circulation (MOC) takes a couple of decades to reach a steady-state (Figures 3e and 3f). The rapid AMOC adjustment might be caused by the dynamical response, such as the Rossby wave propagations. These DMTB anomalies relative to the control seem to be causing the nearly steady DHT anomalies in the deep ocean (Figures 3i and 3j). As a result, the DHT continues to warm the deep ocean layers at a constant rate (Figure 3d) compared with the temperature in the upper layer from the surface to 2,000 m (Figure 3c). Here, the DHT

![Figure 3](image-url)

**Figure 3.** Time series with a 5-year running average of (a) global-mean surface air temperature anomaly, (b) global-mean net downward radiative flux anomaly at the top of the atmosphere, (c) global-mean of the vertically averaged ocean temperature anomaly from the surface to 2,000 m, (d) global-mean of the vertically averaged ocean temperature anomaly from 2000 m to the bottom, (e) southward transport of Atlantic meridional overturning circulation across the equator, (f) northward transport of Pacific meridional overturning circulation across 10°S, (g) DMTB across 2,000 m, (h) DMTB across 4,000 m, (i) global downward heat transport (DHT) anomaly at 2,000 m, and (j) global DHT anomaly at 4,000 m. The hatch shows the range of the standard deviations for the Control. The red dashed lines in (a), (b) show the linear approximations over 16–68 years for b10 m.
anomaly at 2,000 m is roughly equal to the global-mean DRF anomaly at the TOA. The surface climate typically reaches approximately equilibrium on a timescale of a couple of decades, but since the ocean interior is not in equilibrium, there will continue to be slow surface change on the long timescale of deep-ocean adjustment. On average, for both b20 m and b10 m experiments for years 1–70, the global-mean DRF anomaly at the TOA is about 0.1 W/m$^2$ (50 TW), and the DHT anomaly at 2,000 m is 45.6 TW with a DMTB decrease of 8.1 Sv (72%). The DRF anomaly at the TOA consists of downward longwave radiation and downward shortwave radiation. At this time, the downward longwave radiation increases by 0.19 W/m$^2$, while the downward shortwave radiation decreases by 0.09 W/m$^2$. The global average heat budget analysis reveals a series of possible responses as follows:

1. The weakening of deepwater formation due to the reduction of DMTB increases the net DHT anomaly due to the advective component (Figure 1).
2. The increase in the advective DHT causes the surface layer temperature to decrease (Figure 3a).
3. The cooling of the Earth's surface suppresses outgoing longwave radiation from the Earth, which increases the global mean DRF anomaly at the TOA (Figure 3b) and therefore increases the overall heat energy of the Earth system (Figures 3c and 3d).

4. Discussion

In this study, we are not focusing on the dense-water formation due to the open ocean convection itself but assessing the influence on the global heat budget when dense water over the land shelf is transported to the deep layers. However, the formation of dense water and the uptake of heat into the ocean are closely related. To clarify where the ocean heat uptake is dominant, we show the ocean surface heat flux anomalies averaged over 1–70 years relative to the control case for the b20 m experiment in Figure S7a in Supporting Information S1. Only a few specific areas have significant heat absorption, and these areas have substantial reductions in sea surface temperature (SST; Figure S7b in Supporting Information S1). We can see significant heat absorption in the North Atlantic Current, which is the upper branch of the AMOC. The reduction in northward heat transport in the surface layer due to the AMOC weakening (Figure 3e) would cause a decrease in SST, increasing the heat input to the ocean. This process is similar to the “redistribution feedback” in Garuba and Klinger (2016) and may enhance the response to deepwater formation changes in the North Atlantic. Meanwhile, significant heat absorption is found along the Antarctic coastal margins, especially around the Ross and Weddell Seas, accompanying a decrease in SST. This decrease in temperature is likely due to a decrease in coastal deepwater formation, which reduces the transport of warm water masses from off Antarctica. The decrease in SST here is smaller than that in the North Atlantic because the SST has almost reached the freezing point. In the North Pacific, we can also see a decrease in SST that could be related to changes in the Kuroshio Current. A decrease in the deepwater formation may have some indirect effect on the Kuroshio, but the water masses in this region are not directly connected to the deepwater formation, so it is outside the scope of our study.

We show the change in the mixed layer depth (MLD) in Figure S7c in Supporting Information S1 to see the effect of vertical convection carrying heat to the deeper layers. In the mean-field of the control case, an MLD over 2,000 m is found in the Ross, Weddell, and GIN Seas. In the b20 m experiment, the area of the deep mixed layer around Antarctica shifts farther offshore, but the depth itself does not change significantly, likely due to the increase in sea ice extent (for the b20 m experiment, the sea ice extent increases by 500,000 km$^2$ in September). There is also no significant change in the GIN Seas. Therefore, the surface water mass transformation in GIN seas does not affect the formation of deepwater significantly.

The response shown in this study could occur even during global warming (Text S3 in Supporting Information S1). However, it is hard to distinguish the effects of deepwater formation from global warming. Thus, we make this a subject of future research. Our results possibly depend on the model configurations. The ideal experiments suggest that the DMTB and the compensative upwelling play an important role in the DHT in the ocean. To reproduce these processes, water mass transport through the BBL needs to be represented explicitly, while ensuring water mass conservation for the whole model. A model that adapts only the diffusion process to the BBL is unlikely to reproduce the present results.
5. Conclusions

In this study, we evaluated the transient impact of the decrease of deepwater formation on the ocean heat uptake and the heat budgets in the climate system by reducing the thickness of the BBL. The amount of downward water mass transport related to deepwater formation decreases with the change in thickness. The weakening of deepwater formation at high latitudes and associated weakening of the compensative upwelling at lower latitudes increases the DHT. NADW contributes 60% of the heat transported to depths below 2,000 m, and AABW contributes the remaining 40%. AABW is dominant in the DHT to depths below 4,000 m. The DHT is mainly driven by changes in the advection term, while the diffusion term acts to suppress the DHT. We estimate that the DHT increases by a rate of 5.4–6.0 TW/Sv at 2,000 m as the DMTB decreases.

Our experiments demonstrated that changes in deepwater formation have non-negligible impacts on not only ocean heat content but also the atmospheric heat balance. The reduction of deepwater formation leads to a DHT increase of 45.6 TW at 2,000 m due to the advective component. The DHT to deeper layers is roughly equivalent to the change in the global radiative budgets at the TOA (50 TW; the global-mean DRF at TOA is about 0.1 W/m²). The increase in DHT related to deepwater formation warms the deep ocean below the permanent thermocline around 2,000 m and reduces the global average SAT (~0.14 K). Due to the cooling of the Earth’s surface, the outgoing longwave radiation at the TOA is reduced. As a result, the global-mean DRF at the TOA increases to compensate for the heat transported to the deeper layers. Under global warming, the heat transport to the deep layer due to the weakening of the DMTB suppresses the increase in SAT while strengthening the radiative heat input at the TOA and contributing to the increase in the overall heat content of the climate system. The present study also implies that the heat transport to the deep layers through the BBL has a non-negligible impact on estimating climate sensitivity uncertainty.

Data Availability Statement

Data are available at Suzuki, Tatsuo (2022), “DATA for ‘Transient Influence of the Reduction of Deepwater Formation on Ocean Heat Uptake and Heat Budgets in the Global Climate System’”, Mendeley Data, V1, https://doi.org/10.17632/rz3v3mgw hm.1.

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