Control of transient climate response and associated sea level rise by deep-ocean mixing

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Keywords: transient climate response, ocean heat uptake, sea level rise, turbulent mixing, vertical diffusivity

Abstract
To evaluate uncertainty in the transient climate response (TCR) associated with microscale deep-ocean mixing processes induced by internal tidal wave breaking, a set of idealized climate model experiments with two different implementations of deep-ocean mixing is conducted under increasing atmospheric CO$_2$ concentration 1% per year. The difference in TCR between the two experiments is 0.16 $^\circ$C, which is about half as large as the multimodel spread of TCR in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. The TCR difference can be attributed to the difference in the preindustrial climatological state. In the case where deep-ocean mixing works to enhance ocean stratification in the Pacific intermediate-to-deep layers, because the Pacific water mass is transported to the Southern Ocean by the Pacific meridional overturning circulation, the subsurface stratification in the Southern Ocean is also enhanced and deep wintertime convection there is suppressed. Our study shows that in this case during CO$_2$ increase, ocean heat uptake from the atmosphere to deeper layers is suppressed and TCR is estimated to be higher than the other case. Diminished accumulation of oceanic heat in the deep layer also leads to the sea level depression of ~0.4 m in the Southern Ocean when atmospheric CO$_2$ concentration has quadrupled. Together with convective and cloud-radiative processes in the atmosphere and oceanic mesoscale processes, microscale deep-ocean mixing can be one of the major candidates in explaining uncertainty in future climate projections.

1. Introduction
Climate models have shown that anthropogenic emissions of greenhouse gases will lead to future global warming during the present century (Cubasch 2001, IPCC 2013). However, future climate projections have significant uncertainties in various aspects, which can be attributed to numerous deficiencies in climate model parameterizations for unresolved subgrid-scale processes (e.g. those of clouds, radiation, oceanic mesoscale eddies, and turbulence) and their mutual interactions (e.g. Cess et al 1996, Murphy et al 2004, Stainforth et al 2005, Williams and Webb 2008, Roberts et al 2009, Collins et al 2011, Sakamoto et al 2012, Shiogama et al 2012). Conventionally, the magnitude of human-induced global warming is determined by the change of the globally averaged surface air temperature (SAT). One of the major metrics commonly used to estimate model sensitivity to increasing greenhouse gases is the transient climate response (TCR), which is defined as the globally averaged SAT increment at the time of CO$_2$ concentration doubling in an experiment with CO$_2$ concentration increasing at 1% per year (Cubasch 2001; hereafter, 1pctCO$_2$ experiment). The mean TCR of the 17 climate models participating in the fifth phase of the Coupled Model Intercomparison Project (CMIP5) is 1.83 $^\circ$C with a standard deviation of 0.40 $^\circ$C (Kuhlbrodt and Gregory 2012). It is necessary to narrow the range of uncertainty in current estimates of TCR by identifying the subgrid-scale processes responsible and improving their model parameterizations.

When atmospheric CO$_2$ concentration is increased, TCR is determined by the strength of both ocean heat uptake (OHU) and climate sensitivity.
(e.g. Wigley and Schlesinger 1985, Knutti and Hegerl 2008, Goodwin et al. 2013). This can be expressed approximately in terms of the energy balance (e.g. Raper et al. 2002, Dufresne and Bony 2008, Winton et al. 2010):

\[
\text{TCR} = \frac{F_R - \text{OHU}}{\lambda},
\]

where \(F_R\) is radiative forcing due to the increase in CO₂ concentration, and \(\lambda\) is an equilibrium feedback parameter. Previous studies attempted to estimate the intermodel spread in \(F_R\), \(\lambda\), and OHU in equation (1) (Dufresne and Bony 2008, Geoffroy et al. 2012). For example, Geoffroy et al. (2012) estimated that the strength of radiative feedback constitutes the primary source (56%) of intermodel spread of TCR. They also evaluated that the second most important factor in TCR spread is the adjusted radiative forcing associated with the surface temperature response (which cannot be considered feedback), e.g. stratospheric and tropospheric adjustments (Gregory and Webb 2008). Although OHU is affected considerably by subgrid-scale processes in the ocean, previous studies have not discussed the relationship between subgrid-scale processes and OHU.

Uncertainty in OHU stems from uncertainty in the internal heat redistribution mechanism of the ocean. This redistribution mechanism is driven by the meridional overturning circulation (MOC), and the strength of the MOC is closely tied to the distribution and intensity of subgrid-scale mixing process (Munk and Wunsch 1998). In ocean general circulation models and in climate models, oceanic mixing intensity is parameterized as vertical diffusivity. Many modeling studies have shown that the distribution of vertical diffusivity is important in reproducing the Pacific MOC (e.g. Bryan 1987, Hasumi and Suginohara 1999, Furue and Endo 2005). This is because downward mixing of heat at the thermocline reduces the density of cold deeper waters, facilitating their upwelling into the upper ocean. The energy required for the mixing process in the ocean interior is mainly supplied by processes on the scale of tide–topography interactions to internal waves (Munk and Wunsch 1998, Egbert and Ray 2000, Niwa and Hibiya 2011). Then, the supplied energy cascades down to dissipation scales of just a few meters (Munk and Wunsch 1998). However, owing to limited observational data and computational resources, the distribution of the mixing intensity in the real ocean remains unclear. For this reason, climate models generally use the one-dimensional (Bryan and Lewis 1979, Tsujino et al. 2000) or three-dimensional (Jayne and St. Laurent 2001, Jayne 2009) prescribed vertical diffusivity distribution, which is ‘tuned-up’ to reproduce the global MOC.

The heat absorbed by OHU, which is accumulated within the ocean interior, leads to increase in the ocean heat content and to thermal (steric) expansion of the ocean. Recent studies based on satellite altimeter data collected since 1993 have shown that the rate of global sea level rise (SLR) is \(3 \pm 0.4 \text{ mm yr}^{-1}\) (Ablain et al. 2017, Nerem et al. 2018). Church et al. (2013) reported that the most important contribution (30%–50%) to 20th and 21st century SLR is likely to be the thermal expansion of seawater. Other contributions to SLR include the melting of glaciers, changes in the mass of the Antarctic and Greenland ice sheets, and changes in terrestrial storage of water (Church et al. 2013). Church et al. (2004) also clarified the regional distribution of SLR relative to the global mean using satellite altimeter data and tide gauge data. Societal demands for information on the global and regional effects of climate change have increased significantly worldwide. Thus, to support political decision-making processes related to strategies for the mitigation of and adaptation to the effects of global warming, the serious implications of regional SLR associated with the heat redistribution as well as global SLR must be considered. However, Pardaens et al. (2011) and Melet and Meyssignac (2015) reported considerable intermodel spread in the regional patterns of SLR projected by climate models. Consequently, it is necessary to study the uncertainty in global and regional SLR arising from model parameterizations for unresolved subgrid-scale processes.

This study conducted a set of climate model experiments to evaluate the TCR and SLR differences associated with the distribution of mixing intensity within the ocean. These experiments employed two vertical diffusivity distributions: an empirical one-dimensional distribution (Tsujino et al. 2000) and a three-dimensional distribution derived from a simulated conversion rate of the barotropic tide to internal tide energy (Niwa and Hibiya 2014). An idealized 1pctCO₂ experiment was conducted to discuss the effects of mixing processes on OHU, TCR, and SLR in an easily comprehensible manner. Given both the importance of steric SLR and the uncertainty associated with the melting of glaciers, changes in the mass of the Antarctic and Greenland ice sheets, and changes in the terrestrial storage of water, only steric SLR was considered. In section 2, we describe the experimental method, and the results of the simulations are presented in section 3. Finally, a summary and discussion follow in section 4.

2. Model description

This study used the global climate model MIROC5.2 (Tatebe et al. 2018; hereafter T18), which is a version of MIROC5 with a minor upgrade (Watanabe et al. 2010). The horizontal resolution of the atmospheric component has T42 spectral truncation (approximately 300 km) and there are 40 vertical levels up to 3 hPa. The oceanic component is a tri-polar grid. In the spherical coordinate portion to the south of 63°S, the longitudinal grid spacing is 1°, while the meridional
grid spacing varies from approximately 0.5° near the equator to 1° in mid-latitude regions. There are 63 vertical levels, the lowermost of which is located at the depth of 6300 m.

In this study, we conducted two experiments, in which the distribution of vertical diffusivity \( K_v \) was set as follows. In the tide energy dissipation run (hereafter TED), \( K_v \) was diagnosed based on a global three-dimensional map of the turbulent energy dissipation rate of tide-induced internal waves. The global map was obtained in the same manner as described in Tanaka et al. (2012), except the global three-dimensional model of Niwa and Hibiya (2014) was used as the tide model. It was assumed that 30% (100%) of internal tide energy dissipation for each tidal constituent occurred locally with a vertical decay scale of 500 m from the sea floor if the tidal frequency was superinertial (subinertial) (St. Laurent et al. 2002). The remaining part was considered to radiate away and to contribute to the background vertical diffusivity of \( 10^{-6} \text{ m}^2\text{s}^{-1} \). See figure S1 (available online at stacks.iop.org/ERL/9/094001/mmedia) for an illustration of the depth-integrated tidal energy dissipation rate (\( \epsilon \)). The value of \( \epsilon \) reached \( 10^{-2} \text{ W kg}^{-1} \) just above rough bottom topography, e.g. areas such as the Solomon Islands, Hawaiian Ridge, and Izu–Ogasawara Ridge. Following Osborn (1980), \( \epsilon \) could be converted to \( K_v \) as \( K_v = 0.2 \epsilon N^{-2} \), where \( N \) is the buoyancy frequency. Above the depth of 500 m, \( K_v \) was replaced by vertical diffusivity diagnosed using a turbulent closure model, if the diagnosed value was larger than \( K_v \). Figure S2 displays the Pacific zonal mean \( K_v \) that reaches \( 10^{-2} \text{ m}^2\text{s}^{-1} \) in the deep layers of the Pacific Ocean. In the control run (hereafter CTRL), \( K_v \) was prescribed following an empirical vertical profile proposed by Tsujino et al. (2000):

\[
K_v = \begin{cases} 
0.1 + 0.9 \left(1 + \tanh \frac{z}{1500}\right) & \text{for } z \leq 1500, \\
-1 + 2.0 \left(1 + \tanh \frac{z}{1500}\right) & \text{for } z > 1500,
\end{cases}
\]

where depth \( z \) is measured in meters, and the unit of the resultant \( K_v \) is \( 10^{-4} \text{ m}^2\text{s}^{-1} \). This profile, which considers observed bottom-intensified mixing (e.g. Roemmich et al. 1996, Polzin et al. 1997, Morris et al. 2001), ensures realistic reproducibility of the Pacific MOC.

For model spinup, CTRL was integrated for 2000 years under preindustrial forcing conditions of 1850; TED was integrated for 1500 years using initial conditions from CTRL in the 1200th year. It should be noted that T18 discussed in detail the results obtained under the prescribed preindustrial forcing conditions. Then, we conducted the 150 years long 1pctCO2 experiments. The atmospheric CO2 concentrations in the 71st and 141st years correspond to values double and quadruple that at the beginning of the experiment, respectively.

3. Results

3.1. Transient climate response and surface heat flux

Figure 1(a) shows the annual mean time series of global mean surface air temperature anomaly from preindustrial conditions (Tatebe et al. 2018) over 150 years. The TCR defined as the global mean SAT anomaly averaged over a 20-year period centered at the 71st year, when the prescribed atmospheric CO2 concentration becomes doubled, is 1.76 °C in TED and 1.60 °C in CTRL (table 1). Based on Welch’s t-test, the TCR in TED is significantly larger than in CTRL (p < 0.05). In addition, the TCR4, defined as the global mean SAT change averaged over a 20-year period centered at the 141st year when the prescribed atmospheric CO2 concentration becomes quadrupled, is 4.54 °C in TED, i.e. 0.5 °C larger than in CTRL (4.04 °C). As already mentioned in section 1, the standard deviation of TCR of the 17 climate models participating in CMIP5 is 0.40 °C (Kuhlbrodt and Gregory 2012); thus, the TCR difference associated with the distribution of mixing intensity is half that of the CMIP5 multimodel spread.

We estimated the equilibrium feedback parameter \( \lambda \) in equation (1) following Gregory and Mitchell (1997). Net downward heat flux at the top of the atmosphere (\( F_N \)) is given by \( F_N = F_R - F_U \), where \( F_U \) is the increase in upward radiation at the top of the atmosphere. In TED (CTRL), \( F_N \) averaged...
over a 20 year period centered at the 71st year is 1.27 (1.32) W m$^{-2}$. Assuming that the heat capacity of the atmosphere is much smaller than that of the ocean, $F_N$ is nearly equal to the OHU in equation (1) (Gregory and Mitchell 1997, Dufresne and Bony 2008). $F_R$ at the time of doubled CO$_2$ in MIROC5.2 is 4.1 W m$^{-2}$ (see figure S3). Thus, we obtained a value for $\lambda$ of 1.61 (1.74) W m$^{-2}$ K$^{-1}$ in TED (CTRL). In addition to $\lambda$, under the 1pctCO$_2$ scenario, we evaluated the ‘ocean heat uptake efficiency’ ($\kappa$) as OHU $\approx \kappa$TCR (Gregory and Mitchell 1997). We obtained $\kappa = 0.72$ W m$^{-2}$ K$^{-1}$ in TED, which is 0.11 W m$^{-2}$ K$^{-1}$ smaller than in CTRL (0.83 W m$^{-2}$ K$^{-1}$). Based on the comparison of the multimodel mean value for $\lambda$ ($\kappa$) of 1.15 (0.64) W m$^{-2}$ K$^{-1}$ with the standard deviation of 0.36 (0.12) W m$^{-2}$ K$^{-1}$ of the CMIP5 climate models (see table 1 in Text S1 in the auxiliary material of Kuhlbrodt and Gregory (2012)), the difference in $\lambda$ ($\kappa$) due to the difference in the distribution of mixing intensity corresponds to about half (the same degree) of $\lambda$ ($\kappa$) uncertainty of the CMIP5 models. The smaller magnitude of $\kappa$ in TED is discussed in section 3.3.

In the 1pctCO$_2$ experiments, the difference between the beginning and the end of the experiments is greatest, so that we analyze the simulated results averaged over years 1–10 and 141–150 in detail. To examine the latitudinal dependence of SAT increase, figure 1(b) shows the zonal mean SAT averaged over years 1–10 and 141–150. The SAT in TED is smaller than in CTRL by approximately 5 $^\circ$C within 80$^\circ$S–50$^\circ$S during years 1–10 (red and black dashed lines). In high-latitude regions of the Northern Hemisphere, owing to ice albedo feedback, SAT increases over the 140 years reach approximately 15 $^\circ$C in both runs; however, it should be noted that only small differences exist between CTRL and TED for years 1–10 and 141–150.

As the atmosphere is warmed primarily by long-wave radiation from the Earth’s surface, the air–sea heat flux in years 1–10 is evaluated. Figure 2 displays the simulated spatial variations of air–sea heat flux and sea ice area to the south of 40$^\circ$S averaged over years 1–10. In TED, the Ross Sea (80$^\circ$S–50$^\circ$S, 150$^\circ$E–150$^\circ$W) is covered with ice during winter, and upward heat flux is significantly smaller than in CTRL. The greater sea ice coverage in the Southern Ocean in TED is discussed in T18. Briefly, given the preindustrial forcing in TED, T18 clarified that the subsurface stratification in the Southern Ocean becomes reinforced and that the deep convection process that mixes relatively cold fresh water of the surface layer with warm saline water of the subsurface layer becomes suppressed. This causes a decrease in sea surface temperature and greater sea ice coverage during winter in the Southern Ocean. The diminished wintertime ocean cooling in TED could lead to colder SAT (figure 1(b)) because of the smaller supply of heat from the ocean to the atmosphere (see Text S1).

Figure 3 shows the time series of downward air–sea heat flux averaged over 80$^\circ$S–50$^\circ$S in both runs. In years 1–10, the air–sea heat flux to the south of 50$^\circ$S is 7.14 and 12.4 W m$^{-2}$ in TED and CTRL, respectively. The upward heat flux over 80$^\circ$S–50$^\circ$S decreases gradually and around the 120th (140th) year, ocean heat release stops in TED (CTRL). We note that our results show almost the same upward heat flux in the Southern Ocean in years 141–150. This leads to the same distribution of zonal mean SAT in both TED and CTRL in years 141–150 (figure 1(b)).

### 3.2. Formation of deep and bottom water

In the Southern Ocean, sea surface cooling results in deep-ocean convection that forms Antarctic Bottom Water (AABW). The AABW in the Southern Ocean spreads northward as AABW in the Atlantic, as Circumpolar Deep Water (CDW) in the Pacific, and as CDW or modified North Atlantic Deep Water in the Indian Ocean (Lumpkin and Speer 2007, Heuzé et al 2013). The CDW volume transport from the Southern Ocean to the Pacific across 30$^\circ$S is 4 and 12 Sv (1 Sv = 10$^6$m$^3$s$^{-1}$) in TED and CTRL, respectively (T18). Observed CDW volume transport is estimated to be 12 Sv (Timplis et al 1998), 8–9 Sv (Wijffels et al 2001), or 13 Sv (Talley et al 2003), indicating that the deep-water formation in TED is smaller than the observational estimates. The smaller CDW volume transport (weaker Pacific MOC) in TED than in CTRL is discussed in T18. In TED, potential density in the Pacific intermediate layer around the 1000 m depth is lighter than in CTRL because of weaker ocean mixing in the intermediate layer in TED. This relatively light intermediate water is transported to the Southern Ocean along the return path of CDW, resulting in reinforcement of the subsurface stratification in the Southern Ocean. Consequently, the deep wintertime convection in the Southern Ocean becomes weaker and the formation rate of AABW becomes smaller in TED than in CTRL. At the same time, the Pacific MOC becomes weaker and so does the northward transport of CDW. In the real ocean, AABW usually forms when cold dense water spills off the continental shelf (Orsi et al 1999, 2002, Williams et al 2010). In most current climate models, however, the major part of AABW is formed by spurious open-ocean deep convection off the Antarctica (Heuzé et al 2013). It is noted that the formation process of AABW

| Table 1. Comparison of climate response between CTRL and TED. |
|------------------------|------------------------|
|                         | CTRL                  | TED          |
| TCR (°C)                | 1.60                  | 1.76         |
| TCR4 (°C)               | 4.04                  | 4.54         |
| $F_N$ (OHU) (W m$^{-2}$) | 1.32                  | 1.27         |
| $F_R$ (W m$^{-2}$)      | 4.1                   | 4.1          |
| $\lambda$ (W m$^{-2}$ K$^{-1}$) | 1.74             | 1.61         |
| $\kappa$ (W m$^{-2}$ K$^{-1}$) | 0.83              | 0.72         |
in the present modeling study is different to that in the real ocean.

When SAT increases, sea surface cooling becomes weak and the temperature of both AABW and CDW increases (i.e. effective OHU) (Purkey and Johnson 2010, Kouketsu et al 2011). Weakening of sea surface cooling (i.e. less negative buoyancy input) also causes less deep-ocean convection (de Lavergne et al 2014) and less AABW and CDW formation (Purkey and Johnson 2012). When convection does not reach the deep layer, accumulation of oceanic heat in the deep layer stops (i.e. less effective OHU). In this section, we examine the deep-ocean convection and ocean temperature change in each run.

Figure 4(a) displays the time series of maximum wintertime mixed layer depth south of 50°S, and the zonally averaged ocean temperature change between years 1–10 and 141–150 is shown in figure 4(b) and (c), respectively. In the Southern Ocean, because of deep convective overturning reaching the depth of ~4000 m (figure 4(a)) for ~130 years, the temperature in the deep layers (>1000 m depth) in CTRL increases by ~0.5 °C (figure 4(b)). In TED, after the 20th year, the wintertime convection reaches a depth of only ~1000 m (figure 4(a)). Owing to weak convective overturning, the increase in temperature in the deep layers in the Southern Ocean is much smaller (<0.1 °C) in TED (figure 4(c)). This leads to less accumulation of heat in the deep layer (less effective OHU), an increase in sea surface temperature, and a larger value of TCR in TED. We note that, in the North Atlantic, which is known
to have significant heat release as in the Southern Ocean, there are only small differences between TED and CTRL. Because not the deep-ocean mixing but the wind-induced Ekman upwelling in the Southern Ocean is a controlling factor for North Atlantic Deep Water (NADW) formation which determines the ocean temperature in the intermediate-to-deep layer in the North Atlantic (Marshall and Speer 2012), there are no significant temperature differences in the North Atlantic.

3.3. Increase in ocean heat content and steric sea level rise

The difference in ocean temperature increase between TED and CTRL, discussed in section 3.2, is responsible for the difference in the heat content between the upper (<1000 m depth) and deep (>1000 m depth) layers (figure 5). Owing to deep wintertime convective overturning, the heat content in the deep layer of the Southern Ocean in CTRL increases more than in TED by ~10 GJ m$^{-2}$ (1 GJ m$^{-2} = 10^{9}$ J m$^{-2}$) (figure 5(b)). In TED, owing to weak convective overturning in the Southern Ocean, the increase in heat content in the deep layers in the Southern Ocean is much smaller than in CTRL. The less effective OHU (smaller $\kappa$) in TED is due to this smaller heat transport down to the deep layer. In the upper layer (figure 5(a)), local surface cooling (figure 2) and deep wintertime convection in CTRL decrease the heat content more than in TED by ~10 GJ m$^{-2}$ in the region $70^\circ$S–$50^\circ$S, $170^\circ$W–$120^\circ$W.

The increase in heat content throughout the water column, as well as the salinity change, causes steric SLR. Here, we investigate the regional discrepancies in steric SLR between TED and CTRL. First, SLR due to temperature change (thermosteric SLR, $\Delta \eta_{\text{thermo}}$) can be evaluated as follows:

$$\Delta \eta_{\text{thermo}} = \int_0^H \frac{\Delta T \rho}{\rho_{\text{ref}}} \frac{\partial \rho}{\partial T} dz,$$

(2)

where $\Delta T$ is the change in temperature, $\rho$ is the density of seawater, $\rho_{\text{ref}}$ is the reference density of seawater (with temperature of 0 $^\circ$C and salinity of 35 psu), and $H$ is the ocean depth (Gill and Niiler...
1973, Landerer et al 2007). Figure 6(a) shows the spatial distribution of thermosteric SLR (equation 2) in TED after 140 years relative to CTRL. It should be noted that the increase in heat content throughout the water column causes thermosteric SLR. The steric SLR ($\Delta \eta_{\text{steric}}$), comprised of thermosteric SLR and halosteric SLR (SLR caused by salinity change), can be evaluated as follows:

$$\Delta \eta_{\text{steric}} = \int_{-H}^{0} \Delta \rho \rho_{\text{ref}} dz,$$

where $\Delta \rho$ is the change in density (Gill and Niiler 1973, Landerer et al 2007). Figure 6(b) shows the spatial distribution of steric SLR (equation 3) in TED after 140 years relative to CTRL. It can be seen that steric SLR shows the same distribution pattern as thermosteric SLR (figure 6(a)). This indicates that thermosteric SLR is the main control of steric SLR and that halosteric SLR has a secondary role. The largest difference in steric SLR evident in figure 6(b) is in the Southern Ocean. Within the region 70°S–50°S, 170°W–120°W, where local surface cooling (figure 2) and deep wintertime convection both act to reduce the heat content in the upper layer in CTRL, steric SLR in TED is larger than in CTRL by ~0.4 m. In the Southern Ocean, except this region, steric SLR in TED is smaller than in CTRL by ~0.2 m, which is attributable to the diminished increase in heat content in the deep layer (see figures 5 and 6(b)). Only small differences in steric SLR (<0.1 m) are evident to the north of 30°S in the Pacific, Atlantic, and Indian Oceans.

Figure 6(c) shows the time series of global mean steric height anomaly from preindustrial conditions (T18) after 150 years. Because the simulated sea level is depressed unrealistically in the Mediterranean and Black seas because of insufficient water exchange through the Straits of Gibraltar and too much evaporation in the Mediterranean Sea, steric SLR in these regions is excluded in this analysis. Only small differences between TED and CTRL (<0.01 m) are found throughout the simulation period. According to the CMIP5 model comparison, the multimodel spread in SLR in the Southern Ocean is larger than in other oceans, but the contribution of the Southern Ocean to the global mean SLR is smaller (Yin 2012, IPCC 2013). Similarly, the difference in the distribution of the mixing intensity causes differences in SLR in the Southern Ocean, but the contribution of the Southern Ocean to the global mean SLR is smaller than the other oceans and the difference in global mean SLR is not significant.

4. Summary and discussions

In this study, uncertainty in TCR and steric SLR to increasing CO$_2$ concentration, a certain fraction of which can be arisen from microscale deep-ocean mixing processes induced by internal tidal wave breaking, was evaluated based on a set of climate model experiments. The intensity of mixing in the Pacific intermediate layer controls subsurface stratification there, and via the Pacific MOC, it affects subsurface stratification in the Southern Ocean. Our study showed that, with the implementation of deep-ocean mixing enhancing the subsurface stratification in the Pacific as well as in the Southern Ocean (i.e. TED), the deep wintertime convection was suppressed and the OHU in the Southern Ocean became less than with the other implementation (i.e. CTRL), resulting in higher TCR. TCR uncertainty associated with the deep-ocean mixing was 0.16 °C, which was
half as large as that estimated with the CMIP5 models. Less effective OHU also led to significant future sea level depression around the Southern Ocean.

In addition to the larger TCR than CTRL, TED showed a larger sea ice area in the Southern Ocean due to the reduced deep convection that mixes relatively cold fresh water of the surface layer with warm saline water of the subsurface layer. The relationship between larger area of sea ice in the Southern Ocean and larger TCR can also be found in the comparison among the CMIP5 models. Figure S4 shows a scatter plot of the TCR of the CMIP5 models evaluated by MacDougall et al. (2017) and sea ice area evaluated by Shu et al. (2015). The correlation coefficient between TCR and Antarctic sea ice area is estimated at 0.39 (p = 0.07). The positive correlation coefficient shown in the CMIP5 models is consistent with the relationship between TED and CTRL shown in this study.

In TED, a vertical diffusivity distribution that is more realistic than CTRL was adopted in consideration of the generation and breaking of internal waves. However, there is still room for improvement in the method of parameterizing the microscale deep-ocean mixing. In TED, constant local energy dissipation efficiency (30% for superinertial waves) and constant vertical decay scale from the sea floor (500 m) are assumed. Iwamae and Hibiya (2012) highlighted that although the local energy dissipation efficiency and the vertical decay scale of abyssal mixing are associated with topographic features and have a mutual trade-off relationship, this is not considered in model parameterizations. Using an ocean general circulation model, Oka and Niwa (2013) examined the relationship between the local energy dissipation efficiency and the strength of the Pacific MOC. It was shown that the Pacific MOC is strong when the local energy dissipation efficiency is large. The results of our study suggest that, with the larger local energy dissipation efficiency, heat transport to the deep layer becomes larger, which leads to larger OHU and smaller TCR. In our study, it was also assumed that superinertial tidal energy radiating away contributes to the background vertical diffusivity of $10^{-6} \text{m}^2 \text{s}^{-1}$. However, if we assume that 70% of internal tide energy is transferred far away, the value of the background vertical diffusivity is too small. Oka and Niwa (2013) demonstrated that, when the 70% of superinertial tidal energy is transferred far away and converted into a vertical diffusivity, the globally averaged vertical diffusivity can be more than twice when than when only local dissipation is considered (i.e., TED), and the Pacific MOC becomes twice as strong. The dissipation process of internal waves radiating away from the bottom topography and the resulting distribution of mixing intensity remain open to discussion (Hibiya et al. 2002, MacKinnon and Winters 2005). In addition, boundary mixing (Scott and Marotzke 2002) and wind-induced mixing (Inoue et al. 2017) were not considered in this study. For narrowing uncertainties in OHU and its associated TCR and SLR in future climate projections, it is indispensable to further study the generation, propagation and breaking of internal waves and clarify the distribution of mixing intensity.

**Acknowledgments**

The authors thank Dr. M. Kawamiya for helpful discussions. This work was supported by the Integrated Research Program for Advanced Climate Models (TOUGOU) Grant Nos JPMXD017935457 and JPMXD017935715 from the Ministry of Education, Culture, Sports, Science and Technology, MEXT, Japan, and JSPS KAKENHI Grant Number JP18H04923. We thank Mr. James Buxton from Edanz Group (www.edanzediting.com/ac) for editing a draft of this manuscript.

**Data availability**

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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