The Role of Ocean Circulation in Southern Ocean Heat Uptake, Transport, and Storage Response to Quadrupled CO₂

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ABSTRACT: In response to quadrupled CO₂, the Southern Ocean primarily uptakes excess heat around 60ºS, which is then redistributed by the northward ocean heat transport (OHT) and mostly stored in the ocean or released back to the atmosphere around 45ºS. However, the relative roles of mean ocean circulation and ocean circulation change in the uptake and redistribution of heat in the Southern Ocean remain controversial. Here, a set of climate model experiments embedded with a novel partial coupling technique are used to separate the roles of mean ocean circulation (passive component) and ocean circulation change (active component). For the ocean heat uptake (OHU) response, the mean ocean circulation and ocean circulation change are of equal importance. The OHT response south of 50ºS is mainly determined by mean ocean circulation, while the ocean circulation change generates an anomalous southward OHT north of 50ºS. A heat budget analysis finds that the divergence of passive OHT acts to balance the passive surface heat gain to the south of ~50ºS, while the convergence of active OHT acts to balance the active surface heat loss to the north of ~50ºS. Intriguingly, all the increase in ocean heat storage (OHS) is attributable to the passive component, with the ocean circulation change playing almost no role. In the Southern Ocean, both the active and the passive ocean heat transports are overcompensated by the reverse atmospheric heat transport via the Bjerknes compensation.

KEYWORDS: Southern Ocean; Ocean circulation; Heat budgets/fluxes

1. Introduction

Given its vast capacity to store heat, the ocean can largely regulate Earth’s climate. As the climate warms, the ocean has absorbed more than 90% of the excess heat in the climate system since the 1970s (Levitus et al. 2012; IPCC 2021), especially over the Southern Ocean, which has been recognized as the dominant region for ocean heat uptake (Sen Gupta et al. 2009; Durack et al. 2014; Roemmich et al. 2015; Frölicher et al. 2015; Shi et al. 2018). Previous studies have suggested that the dominance of the Southern Ocean in heat uptake could be caused by wind-driven cold deep-water upwelling (Bryan et al. 1988; Manabe et al. 1990; Morrison and Hogg 2013; Frölicher et al. 2015; Czaja and Marshall 2015). The uptake of heat by the ocean plays a role in changing the pattern of ocean warming as well as the rate of climate warming due to increasing greenhouse gases. Corresponding to enhanced ocean heat uptake, the Southern Ocean subsurface warming has occurred at a faster rate than the global average during the recent decades (Gille 2002; Cai et al. 2010; Purkey and Johnson 2010; Durack et al. 2014; Roemmich et al. 2015; W. Liu et al. 2016), accompanied by a large increase in ocean heat content (Levitus et al. 2012; Durack et al. 2014; Llovel and Terray 2016; Cheng et al. 2017; Zanna et al. 2019). Sallée (2018) showed that 67%–98% of the increase in global ocean heat since 2006 is accounted for by the Southern Ocean, which is consistent with the conclusions of Roemmich et al. (2015). Therefore, the Southern Ocean is a key region in regulating the response of the global climate system to external forcing, and understanding its change in the heat uptake and storage is directly relevant to the diagnosis and prediction of global warming (Gregory et al. 2001; Winton et al. 2010; Rose et al. 2014; Rintoul 2018; Swart et al. 2018).

The processes related to Southern Ocean heat uptake and storage under global warming have been extensively explored (Manabe et al. 1991; Frölicher et al. 2015; Morrison et al. 2016; Armour et al. 2016; Liu et al. 2018; Hu et al. 2020; Dias et al. 2020). Climate models show that most of the anthropogenic heat enters the Southern Ocean at the southern flank of the Atlantic Circumpolar Current (ACC) but is stored north of it, and this mismatch between the locations of heat uptake and storage is attributed to northward ocean heat transport. However, the relative roles of mean ocean circulation and ocean circulation change in the uptake and storage of heat in the Southern Ocean remain controversial. For example, using an ocean-only model forced with a spatially uniform surface flux, Marshall et al. (2015) and Armour et al. (2016) found that the climatological meridional overturning circulation of the Southern Ocean controls the response patterns of ocean heat uptake and storage, while the changes in ocean circulation play a secondary role. Several other studies have come to the same conclusion that heat is absorbed and stored primarily as a passive tracer in the Southern Ocean (Gregory et al. 2016; Swart et al. 2018; He et al. 2019; Zanna et al. 2019; Dias et al. 2020). For example, Dias et al. (2020) showed that...
redistribution of heat by circulation changes only accounts for 25% of heat storage increase in response to doubled CO₂ in the Southern Ocean midlatitudes (35°–50°S). On the other hand, other studies concluded that the changes in ocean circulation have a significant impact on the heat absorption and redistribution in the Southern Ocean, which cannot be solely considered as a passive process (Banks and Gregory 2006; Xie and Vallis 2012). In addition, the ocean circulation changes have been found to play a dominant role in the heat storage pattern change over the Southern Ocean (Winton et al. 2013; Garuba and Klinger 2016; Chen et al. 2019; Hu et al. 2020).

However, these previous attempts to isolate the contributions of mean ocean circulation and circulation change were mostly done using passive tracers within an ocean-only framework and thus have limitations (Xie and Vallis 2012; Marshall et al. 2015; Armour et al. 2016; Garuba and Klinger 2016; Dias et al. 2020). For example, Xie and Vallis (2012) employed a primitive equation ocean model to explore the influence of ocean circulation change on heat uptake, with an ocean surface boundary that is restored to prescribed conditions. The restoring boundary conditions are derived from a fully coupled system with a two-way interaction involving active ocean dynamics. Specifically, the SST anomaly used to calculate the surface heat flux anomaly has already had the imprint of the ocean circulation change in it. Therefore, this ocean model-only framework cannot cleanly separate the passive effect due to the advection by the mean ocean circulation from the active effect due to the ocean dynamics adjustment.

Recently, Liu et al. (2018) applied an overriding technique in climate model to decompose the total Southern Ocean response to quadrupled CO₂ into components caused by wind stress change and the direct CO₂ effect, respectively. They showed that the wind-driven meridional overturning circulation change contributes about 20% to the maximum ocean heat storage change in the Southern Ocean, while the direct CO₂-induced warming contributes about 80%. However, their experiments were not able to separate the role of the mean ocean circulation from that of the ocean circulation change either, since their wind-driven circulation change only captures partially the effect of the ocean dynamical adjustments in response to CO₂ forcing, with the buoyancy-driven component of the ocean circulation change unidentified.

To quantify more properly the relative contributions of mean ocean circulation and ocean circulation change to Southern Ocean heat uptake and storage under anthropogenic CO₂ forcing, this study uses a set of purposely designed experiments in Garuba et al. (2018a). Specifically, within the architecture of the Community Earth System Model (CESM), a partially coupled experiment is configured to isolate the contribution of the mean ocean circulation (referred to as the passive component) by disabling effect of the ocean circulation change on the temperature response and its feedback to the air–sea interaction, while the full contribution of the ocean circulation change (referred to as the active component) is then obtained by subtracting the passive component from the response in a fully coupled experiment. Unlike previous studies within the framework of ocean-only model, this partial coupling capability can cleanly isolate and quantify the passive and active components as well as the associated feedbacks within an atmosphere–ocean coupled system (Garuba and Rasch 2020). The partial coupling technique helps to reveal the important role of ocean circulation changes in regulating climate sensitivity (Garuba et al. 2018a) and in driving Atlantic multidecadal variability (Garuba et al. 2018b). Here, the same advantage of the partial coupling approach is taken to decompose the passive and active components in the Southern Ocean heat response to global warming. Compared with the study of Liu et al. (2018), in which ocean circulation change is wind-driven only, the active component here contains the total contribution of ocean circulation change, including both the wind-driven and the buoyancy-induced parts.

In this study, we examine the relative contributions of mean ocean circulation and circulation change using a partially coupled method within an atmosphere–ocean coupled system, in contrast to the passive tracer method used in the previous studies within ocean-only framework, with new findings as follows: (i) the mean ocean circulation dominates the heat gain south of ~50°S, while the ocean circulation change dominates the heat loss south of ~50°S in the Southern Ocean; (ii) all the increase in ocean heat storage is attributable to the mean ocean circulation, and the ocean circulation change has no contribution, due partly to the almost perfect Bjerknes compensation; and (iii) the buoyancy-induced ocean circulation change plays a negative role in the heat gain and heat storage in the Southern Ocean, in contrast to the positive role of wind-driven ocean circulation change. This is the first time that the role of buoyancy-induced ocean circulation change in the Southern Ocean heat uptake and storage has been identified.

The rest of the paper is organized as follows. Section 2 describes the model experiments and metrics used in this study, with an emphasis on the isolation of the passive and active components. Section 3 presents the response of Southern Ocean heat uptake, transport, and storage to greenhouse forcing, as well as the Bjerknes compensation in the meridional energy transport, due to both the passive and active components. Section 4 verifies role of buoyancy-induced ocean circulation change with the FAFMIP experiments. Section 5 summarizes the main conclusions of this study.

2. Model experiments and analysis methods

a. CESM experiments

We use the output of the CESM experiments implemented by Garuba et al. (2018a), which consist of three simulations (Table 1). The control simulation (CTRL) is integrated with no external forcing in the coupled atmosphere–ocean system, and the fully and partially coupled simulations are both forced by abrupt CO₂ quadrupling. In the fully coupled simulation (FULL), the standard atmosphere–ocean coupling is used. In the partially coupled experiment (PARTatl), the ocean circulation change and its impact on the atmosphere–ocean coupling are suppressed. All three simulations are integrated for 150 years. A mean of the model years of 101–150 is taken for our analysis in this study. Given the novelty of the experimental
approach, we here present a detailed review of the design and configuration of the partially coupled simulations.

1) SURFACE-FORCED AND DYNAMICALLY INDUCED COMPONENT DECOMPOSITION

First, we review the decomposition method for ocean temperature response under an external climate forcing commonly used in previous coupled and ocean-alone tracer experiments and point out its caveats (Banks and Gregory 2006; Bouttes et al. 2014; Gregory et al. 2016; Garuba and Klinger 2016; Marshall et al. 2015; Xie and Vallis 2012).

According to the evolution equation for the ocean temperature response to external forcing,

$$\frac{dT_F'}{Dt} = Q' - \nu_F' \cdot \nabla T_F,$$  \hspace{1cm} (1)

where $Q'$ is the heat flux anomaly at the ocean surface due to the CO2 forcing; $T_F'$ and $\nu_F'$ are three-dimensional ocean temperature and circulation anomalies in response to the forcing, respectively; subscript $F$ denotes the variables in the fully coupled context; the overbar indicates the control value, and $D/Dt$ means the total derivative, consisting of a time derivative and advection due to ocean circulation $[D/Dt = (\partial/\partial t) + \nu_F \cdot \nabla]$. The total response of ocean temperature ($T_F'$) can be decomposed into components due to the surface heat flux anomaly (i.e., surface-forced component $T_{FS}'$) and due to the ocean circulation change that advects the control ocean temperature field (i.e., dynamically induced component $T_{FD}'$). This decomposition can be expressed as

$$T_F' = T_{FS}' + T_{FD}'$$  \hspace{1cm} (2)

$$\frac{dT_{FS}'}{Dt} = Q',$$  \hspace{1cm} (3)

$$\frac{dT_{FD}'}{Dt} = -\nu_F' \cdot \nabla T_F.$$  \hspace{1cm} (4)

To achieve this decomposition in a fully coupled simulation, a set of additional temperature-like tracers are added to the experiment (Banks and Gregory 2006; Xie and Vallis 2012; Marshall et al. 2015; Garuba and Klinger 2016; Gregory et al. 2016). Specifically, a passive tracer $T_{FS}'$, which is designed to present the surface-forced ocean temperature anomaly, is built into the ocean component and integrated forward from zero at initialization under the surface heat flux anomalies $Q'$.

However, the above decomposition cannot cleanly isolate the role of ocean circulation change from mean ocean circulation. This can be seen from a restoration expression of the surface heat flux anomaly $Q'$:

$$Q' = \alpha (T'_a - T_F') = \alpha (T'_a - T_{FS}' - T_{FD}' ||),$$  \hspace{1cm} (5)

where $T'_a$ and $T_F'$ are the atmospheric and oceanic temperature anomalies, respectively; the notation “||” represents surface values of the variables [i.e., surface air temperature (SAT) and sea surface temperature (SST)], and $\alpha$ represents the strength of the coupling between the atmosphere and ocean. Equation (5) shows that the surface heat flux anomaly $Q'$ itself is also dependent on the dynamically induced surface temperature anomaly $T_{FD}' ||$ (i.e., the portion of SST change induced by ocean circulation change). Therefore, in this decomposition framework, the $T_{FS}'$ component is affected by the $T_{FD}'$-induced surface flux perturbation. In other words, the so-called surface-forced component is not entirely surface-originated. Given this caveat, there is a need to design a cleaner decomposition method to separate the active, ocean circulation change-induced temperature response from the passive, pure surface-forced response.

2) PASSIVE AND ACTIVE COMPONENT DECOMPOSITION

To this end, an alternative decomposition was developed by Garuba et al. (2018a) and Garuba and Rasch (2020) leveraging both partial coupling and a passive temperature tracer. To isolate the passive component in the ocean response to CO2 forcing in practice, a passive tracer $T_{PS}'$ that only responds to the passive component of the surface flux is introduced. To ensure that the surface heat flux is consistent with the passive component of the ocean temperature response, the surface heat flux is only allowed to couple with the passive component of the ocean temperature response at the surface; that is,

$$Q'_{\text{passive}} = \alpha (T'_a - T_{PS}') ||.$$  \hspace{1cm} (6)

Through this partial coupling, $T_{PS}'$ is a self-contained component of the ocean temperature response as it is forced by its self-induced surface flux:

<table>
<thead>
<tr>
<th>Name</th>
<th>Run (years)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>150</td>
<td>Control fully coupled simulation (preindustrial CO2)</td>
</tr>
<tr>
<td>FULL</td>
<td>150</td>
<td>Perturbed fully coupled simulation (4 × CO2)</td>
</tr>
<tr>
<td>PARTatm</td>
<td>150</td>
<td>Perturbed partially coupled simulation (4 × CO2)</td>
</tr>
<tr>
<td>faf-water</td>
<td>70</td>
<td>Impose a perturbation in surface freshwater flux</td>
</tr>
<tr>
<td>faf-heat</td>
<td>70</td>
<td>Impose a perturbation in surface heat</td>
</tr>
<tr>
<td>faf-passiveheat</td>
<td>70</td>
<td>As in faf-heat, but the heat flux added as a passive tracer</td>
</tr>
</tbody>
</table>
the total temperature anomaly: $T_v$ change surface flux anomaly here denotes the partially coupled variables) being forced by the fully coupled case comprises of a passive component (induced by the solid box in (c)), $T_p$ is the oceanic temperature anomaly in the partially coupled simulation [represented by the dashed box in (c)], $T_{PS}$ and $T_{FD}$ are the surface forced component of the oceanic temperature anomalies in the fully and partially coupled simulation, respectively [represented by green color in (c)]; $T_{PS}$ and $T_{PD}$ are the dynamically induced component of the oceanic temperature anomalies in the fully and partially coupled simulation, respectively [represented by pink color in (c)]. The stippled portion in (c) represents the passive ocean temperature component $T_{active}$, while the space outside of the stippled portion represents the active ocean temperature component $T_{active}$ of the climate change response. The notation $\sigma_l$ denotes surface values of the variables.

$$\frac{dT_{PS}}{dt} = Q_{passive} = a(T_a - T_{PS})\sigma_l,$$  \hspace{1cm} (7)

This component is referred to as $T_{passive}$ and illustrated as the stippled portion of the full response in Fig. 1c. One can see that the $T_{passive}$ component is due entirely to the passive surface flux $Q_{passive}$. Consequently, the total active response $T_{active}$ is derived by subtracting the passive component from the total temperature anomaly:

$$T_{active} = T_p - T_{PS}.$$  \hspace{1cm} (8)

Note that $T_{active}$ is the result of both the active surface heat flux anomaly $Q_{active}$ and the full circulation change $v_p$, governed by

$$\frac{dT_{active}}{dt} = Q_{active} - v_p \cdot \nabla T.$$  \hspace{1cm} (9)

The expression above is derived from subtracting the passive component [Eq. (7)] from the full temperature response [Eq. (1)] and by acknowledgment that the surface heat flux anomaly ($Q'$) in the fully coupled case comprises of a passive component ($Q_{passive}$) and an active component ($Q_{active}$) induced by ocean circulation change $v_p$ (i.e., $Q' = Q_{passive} + Q_{active}$). As a result, $T_{active}$ contains all the effects of the ocean circulation change.

We note that, even in the partially coupled experiment, the temperature anomaly also consists of a surface-forced and dynamically induced components, with the former identified to be the passive tracer response $T_{PS}$ and the latter $T_{PD}$ (subscript $P$ here denotes the partially coupled variables) being forced by

$$\frac{dT_{PD}}{dt} = -v_p \cdot \nabla T,$$  \hspace{1cm} (10)

where $v_p$ is the partially coupled ocean circulation anomaly. Note that $v_p$ is different from the fully coupled ocean circulation change $v_F$, and hence $T_{PD}$ also differs from $T_{FD}$ [comparing with Eq. (4)]. The relationships among all the temperature variables introduced above, including $T_{PS}$, $T_{PD}$, $T_{FD}$, $T_{passive}$, and $T_{active}$ are illustrated in Fig. 1c.

In summary, through the extraction of the passive surface flux $Q_{passive}$ and the passive ocean temperature $T_{passive}$ using the partial coupling technique, one can cleanly separate the surface forced component (or the passive component $T_{passive}$) from the ocean circulation change induced, active component (i.e., $T_{active}$) under greenhouse gas forcing. Garuba et al. (2018a) and Garuba and Rasch (2020) have demonstrated that the fully coupled surface heat flux and temperature response to greenhouse forcing can be recovered from the sum of the two decomposed components, validating the experimental design. Therefore, $Q_{active}$ and $T_{active}$ can be obtained from the differences between the fully coupled simulation and the partially coupled simulation for the purpose of understanding Southern Ocean heat uptake here.

b. FAFMIP experiments

In addition to the CESM experiments, this study also uses the simulations made by the Flux-Anomaly-Forced Model Intercomparison Project (FAFMIP), which is part of phase 6 of the Coupled Model Intercomparison Project (CMIP6) and designed to isolate the effect of the thermally and buoyancy-forced component of the ocean circulation changes. The FAFMIP experiments branch out from preindustrial conditions (piControl) and are forced by the surface perturbation flux computed from a 13-member CMIP5 ensemble at years 61–80 of the 1pctCO2 scenario, which centers around the time of CO2 doubling. Two types of single-forcing experiments are included, one with the prescribed surface heat flux and another with freshwater flux. The
perturbation heat flux contains both a passive and an active component (in the terminology of the previous subsection) under the CO2 forcing. A passive tracer is introduced to isolate the redistributive component of the ocean temperature and the induced surface heat flux therefrom in the surface heat flux–forced experiment, while a tracer is not necessary for the freshwater flux–forced experiment since the induced ocean temperature redistribution does not feed back immediately to the freshwater flux. In addition, the same heat flux perturbation and passive tracer are applied to the heat tracer experiment and surface heat flux–forced experiment, with the ocean circulation being fixed at the control run in the former experiment but adjusted by heat flux perturbation in the latter experiment. Therefore, a comparison of the passive temperature tracer between the two experiments is able to assess the redistributive effect of heat flux–induced ocean circulation change. Taken together, the two sets of single-forcing experiments plus the heat tracer experiment help isolate the thermohaline component of the ocean circulation change and the resultant changes in the ocean heat uptake and the heat distribution in the ocean. Table 1 summarizes the two types of single-forcing experiments and the passive heat tracer experiment. Interested readers are referred to Bouttes and Gregory (2014) and Gregory et al. (2016) for the details of the FAFMIP experimental design.

As such, the FAFMIP experiments fill the niche in bridging the gap between the full active ocean dynamics feedback detailed in the previous subsection and the wind stress–driven component of the ocean dynamics feedback in Liu et al. (2018). The FAFMIP experiments with two climate models, MPI-ESM1.2-HR (Gutjahr et al. 2019) and MRI-ESM2.0 (Yukimoto et al. 2019), will be examined when we discuss the discrepancy between the full active ocean dynamical feedback and wind-driven ocean dynamical feedback onto the ocean heat uptake (OHU) and ocean heat storage (OHS) in section 4.

d. Ocean heat budget

To connect the heat uptake, transport, and storage, we analyze the heat budget within the Southern Ocean following Liu et al. (2018). Specifically, the zonally integrated full-depth ocean heat budget is

\[ OHS = \frac{\partial}{\partial t} OHC = \frac{\partial}{\partial t} \int_{-H}^{0} \rho_0 C_p \theta \, dz \, dx. \]  (13)

The second term in Eq. (12) is due to the meridional ocean heat transport (OHT), which can be decomposed into contributions from Eulerian-mean flow, eddies, and diffusion (Yang et al. 2015):

\[ OHT = \int_{-H}^{0} \rho_0 C_p (\nabla \cdot (\mathbf{V} + \mathbf{u}^* \theta + D)) \, dz \, dx = \text{OHT}_{\text{Eul}} + \text{OHT}_{\text{Ed}} + \text{OHT}_{\text{Diff}}. \]  (14)

In addition, the OHT_{Eul} is further decomposed into components associated with meridional overturning cells and horizontal gyres (Bryan 1962; Yang and Saenko 2012; He et al. 2019; Williams et al. 2021):

\[ \text{OHT}_{\text{Eul}} = \int_{-H}^{0} \rho_0 C_p (\bar{\mathbf{V}} \cdot \mathbf{u}^* \theta) \, dz \, dx = \text{OHT}_{\text{Cell}} + \text{OHT}_{\text{Gyre}}. \]  (15)

where \( \bar{\cdot} \) represents the zonal average, and the wavy overbar represents the deviation from the zonal average. The term OHT_{Cell} is the heat transport by the overturning cells, while OHT_{Gyre} is the heat transport by the horizontal gyres.

The last term in Eq. (12) is defined as ocean heat uptake (OHU):

\[ \text{OHU} = \int \langle \text{SHF} \rangle \, dx. \]  (16)

Therefore, the ocean heat budget can be written as

\[ \text{OHS} = \text{OHU} - \frac{\partial}{\partial y} \text{OHT}, \]  (17)
which indicates that the OHS is determined by the OHU and heat redistribution due to the convergence of the OHT.

3. Results

a. Climate response to quadrupled CO₂ forcing

We first examine the changes of OHU, OHT, and OHS over the Southern Ocean in response to quadrupled CO₂ in the fully coupled simulation of CESM (Fig. 2). The excess heat resulting from increased atmospheric CO₂ enters the Southern Ocean mostly at the southern flank of the ACC. In contrast, there is a significant heat loss between 50° and 60°S in the southwest Atlantic and Indian Oceans. The irregularly distributed heat gain is also observed north of 40°S (Fig. 2a). The zonally integrated OHU response shows a dipole-like pattern within 40°–70°S, with a negative anomaly to the north of 50°S and a positive anomaly to the south of 50°S (Fig. 2b). The peak of the positive OHU anomaly appears to be around 60°S, but the change in OHS reaches the maximum around 45°S (Fig. 2b), with the warming being concentrated in the upper 1000 m (Fig. 2d). This suggests that the excess heat entering the ocean from the surface is transported northward rather than stored locally, which is confirmed by the northward OHT anomaly in response to quadrupled CO₂ (Fig. 2c). These above characteristics are broadly consistent with previous studies (Frölicher et al. 2015; Marshall et al. 2015; Morrison et al. 2016; Armour et al. 2016; Liu et al. 2018; Hu et al. 2020). In addition to redistributing heat in the Southern Ocean, the northward OHT anomaly also helps keep SST cold around 60°S, thereby allowing additional heat uptake at the surface. Regarding the physical processes that are responsible for the northward OHT anomaly, Fig. 2c shows the total change of OHT and contributions from three processes. It is found that the northward OHT anomaly is dominated by the contributions from the Eulerian-mean flow, while the eddy-induced OHT acts to compensate partially for the Eulerian-mean OHT. The contribution of diffusion can be ignored except around 42°S, where it has a positive contribution to the OHT change.

Previous studies suggested that the Eulerian-mean OHT anomaly is mainly caused by advection of temperature anomalies due to the mean MOC in the Southern Ocean. Specifically, the stronger warming at the ocean surface relative to the deeper ocean increases the vertical stratification (Fig. 2d), which is advected by the mean Southern Ocean MOC (contours in Fig. 2d) and results in a large northward OHT anomaly. However, this conclusion is derived from passive-tracer simulations with an ocean-only model (Armour et al. 2016) or diagnostic analysis (He et al. 2019; Li et al. 2021) rather than a coupled model. Recently, Hu et al. (2020) found that the ocean circulation change plays a more important role than the passive advection process in the transient response of OHT in the Southern Ocean to greenhouse forcing. We will examine the relative roles of the mean ocean circulation and circulation change in the northward OHT change in the Southern Ocean within a coupled model framework in section 3b.

Besides altering ocean temperature, anthropogenic CO₂ forcing also induces a change in the MOC. As introduced in section 2c, the total MOC in the Southern Ocean is maintained by a balance between the Eulerian-mean and eddy-induced MOC. Figure 3 shows the total MOC and its two components as well as their response to greenhouse forcing. The Eulerian-mean MOC mainly represents the wind-driven Deacon cell, which is clockwise and reaches the maximum of ~35 Sv (1 Sv ≡ 10⁶ m³ s⁻¹) at 50°S (contours in Fig. 3b). On the other hand, the eddy-induced MOC due to the mean buoyancy gradient (Marshall and Radko 2003) is counterclockwise and has the maximum volume transport of ~15 Sv at 51°S (contours in Fig. 3c). Their residual is the total MOC with its maximum being ~23 Sv at 50°S (contours in Fig. 3a).
These characteristics of the MOC are overall consistent with previous studies (Marshall and Radko 2003; Marshall et al. 2007; Viebahn and Eden 2010; Yang et al. 2015; Farneti et al. 2015). In response to quadrupled CO2, the upper Eulerian-mean MOC change shows a dipole-like pattern, with negative anomalies to the north of 50°S and positive anomalies to the south of 50°S (Fig. 3b). Since the upper Eulerian-mean MOC is wind driven, we examine the changes of the zonal wind stress. It is found that the Southern Hemisphere westerly winds strengthen and displace poleward (Fig. 4), shifting the Deacon cell in the Southern Ocean poleward by ∼3° and intensifying the cell by about 1 Sv at its maximum (Fig. 3b). Meanwhile, the eddy-induced MOC is strengthened south of ∼50°S and weakened north of ∼50°S (Fig. 4c), which partially offsets the wind-driven Deacon cell. Consequently, the residual MOC generally follows the change in the Deacon cell, featuring a poleward-intensified change in the Southern Ocean (Fig. 3a; Liu et al. 2018).

In summary, the excess heat resulting from increased atmospheric CO2 enters the Southern Ocean around 60°S, transports northward in the upper ocean, and is mostly stored around 45°S. Particularly, the surface heat uptake is largely balanced by anomalous northward heat transport associated with the Eulerian-mean MOC. This above phenomenon has been described in previous studies. However, the relative roles of mean ocean circulation and ocean circulation change in this process remains controversial (Banks and Gregory 2006; Frölicher et al. 2015; Marshall et al. 2015; Armour et al. 2016; Liu et al. 2018; Hu et al. 2020). In the next subsection, we will use the partially coupled experiment with temperature tracer to isolate their contributions to heat uptake, storage, and redistribution in the Southern Ocean.

b. Roles of the passive and active components

1) OCEAN HEAT UPTAKE

Making use of the fully and partially coupled experiments (Table 1), we decompose the total climate response into the surface-forced passive component (driven by mean circulation; i.e., \( T'_{\text{passive}} \)) and ocean-dynamics-induced active component (driven by circulation change; i.e., \( T'_{\text{active}} \)). Figure 5 shows the total OHU response as well as its passive and active components. The decomposition indicates that neither the atmosphere nor the ocean dynamic alone can fully account for the spatial pattern of the OHU anomaly; it is a combination of the passive and active OHU components that sets its overall response in the Southern Ocean, with both components being

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**FIG. 3.** Changes of the MOC (shading; Sv) in response to the quadrupled CO2. (a) The total MOC is the residual of (b) the Eulerian-mean MOC and (c) the eddy-induced MOC. The superimposed gray contours are the MOC of the corresponding components in CTRL (solid lines for clockwise and dashed lines for anticlockwise circulation; contour interval is 5 Sv).

**FIG. 4.** (a) Changes of surface wind stress (vectors; 10^{-2} \text{ N m}^{-2}) and zonal wind stress (shading; 10^{-2} \text{ N m}^{-2}) in response to quadrupled CO2. (b) Zonal-mean changes of zonal wind stress (blue) and zonal wind stress in CTRL (red) weighted by cosine latitude (10^{-2} \text{ N m}^{-2}).
of equal importance. In particular, the positive OHU band at the southern flank of the ACC is dominated by the passive component (Fig. 5b), while the significant heat loss over both the southwest Atlantic and Indian Oceans, as well as the uneven heat gain north of 40°S, are controlled by the active component (Fig. 5c). This above finding is further verified by the zonally integrated OHU changes; that is, the total response south of 50°S is determined by the passive component, while the total response north of 50°S is determined by the active component (comparing black curves in Figs. 5d–f).

While our above result north of 50°S is similar to that of Liu et al. (2018), there is a clear distinction between our and their conclusions south of 50°S. In particular, Liu et al. showed that the peak of heat gain around 60°S is attributed to a combined effect of the direct CO2-induced warming (about two-thirds) and wind-driven ocean circulation change (about one-third) (Figs. 9d and 9f of Liu et al. 2018). In contrast, we find that the heat gain peak around 60°S (Fig. 5d) is attributed solely to the passive component (Fig. 5e), while the ocean circulation change makes little contribution (Fig. 5f). The distinction is also shown by the spatial pattern of the OHU change; for example, the significant positive anomaly around 60°S in their wind-driven response (Fig. 9c of Liu et al. 2018) is nonexistent in our active component (Fig. 5c). In addition, our passive OHU component is characterized by significant positive band around 60°S but weak at other latitudes (Fig. 5b), while there are significant changes north of 50°S in Liu’s direct CO2-induced response, such as the heat loss in the midlatitude Atlantic and India Ocean and the heat gain in the midlatitude Pacific Ocean (Fig. 9e of Liu et al. 2018). These above differences mainly come from the role of the buoyancy-induced ocean circulation change, which is not considered as a part of the effect of the ocean circulation change in Liu et al.’s experimental design. In other words, the ocean circulation change is wind-driven only in the study of Liu et al. (2018), but it is the total effect by both the surface winds and buoyancy fluxes in our study. This suggests that, although the buoyancy-induced MOC change is secondary (Fig. 5g of Liu et al. 2018), it induces an OHU anomaly with the same amplitude as but opposite direction to the wind-driven ocean circulation change, resulting in a near-zero active OHU change around 60°S (Fig. 5f). Therefore, in comparison with Liu et al.’s modeling result, we can conclude that in the high latitudes of the Southern Ocean around 60°S the ocean circulation change driven by the surface winds acts to increase the ocean heat gain, while the buoyancy-induced ocean circulation change hinders the ocean heat gain.

Figures 5d–f show a decomposition of the OHU changes into contributions from shortwave and longwave radiation, latent and sensible heat fluxes, and the heat flux due to sea ice formation and melt. To the south of the ACC (south of ~50°S) where the total response is dominated by the passive component, it is found that the positive OHU anomaly is attributed primarily to a positive sensible heat flux contribution between 50° and 60°S (yellow curves in Figs. 5d,e) and a positive radiative (SW + LW) contribution south of 60°S (orange curves in Figs. 5d,e). Meanwhile, the positive heat flux anomaly due to sea ice formation and melt also makes a significant positive contribution around 60°S where the OHU anomaly reaches its maximum (green curves in Figs. 5d,e). In contrast, the latent heat flux contributes negatively to the positive OHU anomaly south of ~50°S in both the total response and the passive component (blue curves in Figs. 5d,e). In fact, the latent heat flux appears to have a negative anomaly across all the latitudes in the Southern Ocean.
The positive sensible heat flux is due to the fact that the atmosphere warms more rapidly than the sea surface over the region 50°–70°S where the upwelled subsurface waters maintain a relatively constant SST. In terms of the positive radiative contribution south of 60°S, it is likely resulted from reduced shortwave reflection via ice-albedo feedback, which is associated with decreasing Antarctic sea ice concentration. Figure 6 shows the change in sea ice concentration in response to the CO₂ forcing. The reduction of sea ice cover south of 60°S reflects less incoming shortwave irradiance and leads to a positive anomaly there, which is partially compensated by negative LW anomaly (not shown). In addition, the positive heat flux anomaly due to sea ice formation and melt around 60°S may also be linked to the sea ice retreat there, which gives less chance for the seawater to interact with and melt the overlying sea ice (Fig. 6). Regarding the positive radiative flux anomaly north of 50°S, it results from a positive SW cloud feedback due to the decreases of low cloud fraction in response to increased CO₂ (Zelinka et al. 2012; Ceppi and Hartmann 2016; Ceppi and Shepherd 2017; Hu et al. 2020).

On the other hand, the negative anomalous OHU over the region between 40° and 50°S is dominated by negative latent heat anomaly (blue curve in Fig. 5d), while sensible and radiative heat flux anomalies are positive there (yellow and orange curves in Fig. 5d). Since the passive changes in the sensible, latent, and radiative heat fluxes cancel each other out, the passive OHU anomalies appear to be small north of 50°S (black curve in Fig. 5e). As a consequence, the OHU anomalies north of 50°S are dominated by its active component, which comes largely from the latent heat (blue curve in Fig. 5f). A further decomposition finds that the negative latent heat is mainly rooted in Newtonian cooling effect due to the warmer SST (not shown; Xie et al. 2010). Specifically, the negative OHU anomaly within 40°–50°S is caused by a convergence of surface waters, which warms the SST and thus leads to oceanic heat loss via latent heat flux. This is proved by a good match between patterns of active SST anomaly (Fig. 8c) and active OHU anomaly (Fig. 5c).

2) OCEAN HEAT TRANSPORT

In this subsection, we examine how the mean ocean circulation and ocean circulation change modify the OHT and its convergence/divergence in response to quadrupled CO₂ within a coupled model framework. The northward total OHT response (black curve in Fig. 7a) is found to be controlled by its Eulerian-mean component (red curve in Fig. 7a), while its eddy-induced component acts to play a negative role (blue curve in Fig. 7a) and its diffusive component is relatively small and localized around 42°S (green curve in Fig. 7a), a latitude

![Figure 6](image1.png)

**Fig. 6.** Change of sea ice concentration in response to quadrupled CO₂ (shading; %). Superposed solid black and purple lines show the sea ice margin (based on 15% sea ice concentration) in CTRL and FULL simulations, respectively.

![Figure 7](image2.png)

**Fig. 7.** Changes of the meridional OHT (black) and Eulerian-mean OHT (red), eddy-induced OHT (blue), and diffusive OHT (green) in the (a) total response and its (b) passive and (c) active components. Changes of the Eulerian-mean OHT (red) as well as cell-induced OHT (orange) and gyre-induced OHT (yellow) in the (d) total response and its (e) passive and (f) active components.
corresponding to the northern SST front associated with the ACC.

According to the experimental design, the passive OHT change results from the advection of anomalous temperature by mean ocean circulation, and it can explain almost all of the northward total OHT change (comparing Fig. 7b with Fig. 7a). This indicates that the total OHT response in the Southern Ocean is dominated by the passive process, which is consistent with previous studies (Armour et al. 2016; He et al. 2019; Li et al. 2021). This is so because the increasing CO₂ concentration causes greater ocean warming near the surface, and thus the climatological clockwise MOC in the Southern Ocean advects anomalous warm waters northward (red curves in Figs. 7a,b). Similarly, the southward eddy-induced OHT anomaly also results mostly from the advection of temperature anomalies by the climatological eddy-induced MOC (blue curves in Figs. 7a,b). In contrast, the ocean circulation change mainly generates an anomalous southward OHT north of 50°S (black curve in Fig. 7c), and all the Eulerian-mean flow, eddies, and diffusion have contributions to the active component of the total OHT response. In spite of its small magnitude, this active component plays an important role in the convergence of OHT around 45°S, which will be discussed in the next subsection.

Following Williams et al. (2021), we further decompose the Eulerian-mean OHT change into components associated with the meridional overturning cells and the horizontal gyres. Consistent with Williams et al. (2021), the mean MOC induces a northward OHT (orange curve in Fig. 7e) and the mean gyre circulation induces a southward OHT (yellow curve in Fig. 7e). However, the cell-induced northward OHT change overwhelms the gyre-induced southward OHT change, which explains the net northward anomaly of the OHT resulted from the mean ocean circulation (red curve in Fig. 7e). Meanwhile, the change of MOC gives rise to a southward OHT, and the change of gyre circulation gives rise to a northward OHT, with the former being larger than the latter (Fig. 7f).

Compared to the modeling result in Liu et al. (2018), we come to a different conclusion about the role of ocean circulation change in modifying the OHT. While Liu et al. (2018) showed that the wind-driven MOC change generates an anomalous northward OHT to the south of 45°S and an anomalous southward OHT to the north of 45°S, we find that the OHT south of 50°S is hardly altered by the total ocean circulation change. This can be understood in conjunction with the results of OHU in section 3b(1). To be specific, the wind-driven ocean circulation change alone drives a net gain of heat south of 50°S, which needs to be counterbalanced by a northward OHT. However, after taking the surface buoyancy effect into account, the total ocean circulation change plays a minor role in the ocean heat uptake south of 50°S and thus there is no need for a northward OHT.

3) OCEAN TEMPERATURE

Similar to the OHU and OHT, the total response of ocean temperature can also be partitioned into the passive and active components (Fig. 8). As introduced in section 2a(2), the passive temperature component is forced by passive surface heat fluxes and distributed by mean ocean circulation, while the active temperature component contains the total effects of ocean circulation change, including both circulation change-driven temperature anomalies and the circulation change induced, active surface heat fluxes.

We start by examining the response of SST, which is closely related to the surface heat fluxes. It is found that the spatial pattern of the SST response over the Southern Ocean is determined by both its passive and active components (Figs. 8a–c). The passive SST change shows an overall significant warming in the Southern Ocean, except for the region around Antarctica (Fig. 8b). The quite uniformly passive SST change (Fig. 8b) is distinct from the passive surface heat gain, which is centered around 60°S (Fig. 5b). This mismatch between the passive OHU and SST component patterns can be explained by the advective effect of mean ocean circulation, as revealed by previous studies (Armour et al. 2016; Marshall et al. 2015). On the contrary, the active SST component pattern (Fig. 8c) coincides well with the active OHU component pattern (Fig. 5c), with an increase of SST corresponding to the heat loss from the ocean to the atmosphere and vice versa. For example, significant SST warming is found in the southeast of Australia and the southwest Atlantic (Fig. 8c) where the ocean releases a large amount of heat fluxes to the atmosphere (Fig. 5c). This indicates that the active surface heat fluxes have a damping effect on the ocean dynamics–induced SST change.

The zonal mean temperature changes are shown in Figs. 8d–f. It is found that the warming in the total response is mostly stored in the upper ~1000 m of the ocean, with a warming tongue penetrating at a slant from the subsurface into the deeper ocean between 40° and 60°S (Fig. 8d), where the mean downward Ekman pumping is strongest. The passive component (Fig. 8e) appears to have a warming pattern similar to the total response and is able to explain most of the warming in the total response. The passive temperature response can be attributed to the heating from positive surface heat fluxes around 60°S as well as its redistribution by the climatological northward Ekman transport near the surface and subduction into the subsurface ocean at midlatitudes. However, it is the active component of the total warming that contributes mostly to the formation of the warming tongue between 40° and 60°S (Fig. 8f). It is interesting to note that the active temperature component mainly concentrates in the subsurface with a warming south of 40°S and a cooling north of 40°S. To understand the formation processes for the active temperature component, we examine the change of surface wind stress curl in response to increased CO₂ concentration (Fig. 9). It is found that the poleward and intensified westerly winds result in a meridional dipole structure of wind stress curl over the Southern Hemisphere middle latitudes, with a positive wind stress curl anomaly south of 40°S and negative anomaly north of 40°S. The positive wind stress curl generates a downwelling anomaly and leads to subsurface warming, while the negative wind stress curl generates an upwelling anomaly and leads to subsurface cooling (Fig. 8f).

In summary, the mean ocean circulation and ocean circulation change work together to determine the structure of
Southern Ocean warming. Specifically, the climatological westerly winds over the Southern Ocean induce a northward Ekman transport to move excess heat gain at higher latitudes northward and accumulate it in the subduction regions at mid-latitudes, where the heat is converged by climatological ocean circulation. Meanwhile, the warming-induced positive wind stress curl anomalies within 40°–60°S enhance the Ekman convergence and the resultant downwelling there, which transfers more heat downward and increases the subsurface warming. These results are in agreement with previous studies (Fyfe et al. 2007; Cai et al. 2010; Liu et al. 2018; Lyu et al. 2020).

4) OCEAN HEAT BUDGET

The ocean heat budget introduced in section 2d provides a way of linking the OHU, OHT, and OHS. In this subsection we quantify the contributions of mean ocean circulation and ocean circulation change to Southern Ocean heat uptake and redistribution through a heat budget analysis.

Figure 10a shows the total response of the ocean heat budget terms to greenhouse forcing. It is found that the anomalous OHT divergence (blue curve in Fig. 10a) closely follows the OHU change (black curve in Fig. 10a); for example, the surface heat gain around 60°S coincides well with an anomalous OHT divergence, and the surface heat loss around 45°S with an anomalous OHT convergence. The difference between changes of the OHU and OHT divergence is the heat storage (red curve in Fig. 10a). It can be seen that, in response to quadrupled CO₂, a majority of the heat entering the Southern Ocean around 60°S is stored to the north of 60°S as a result of ocean heat redistribution. These above results are also consistent with previous studies (Morrison et al. 2016; Armour et al. 2016; Liu et al. 2018; Hu et al. 2020).

Figure 10b shows the heat budget terms from the partially coupled simulation, which represents the surface-driven passive changes of the OHU, OHT, and OHS. South of 50°S, similar to the case of the total response, the passive component also features a synchronous change in the OHT gradient and OHU. However, the passive component contributes little to the total response to the north of 50°S, in which the active component appears to make a dominant contribution (Fig. 10c). In particular, it is the ocean circulation change that brings the maximum anomalous OHT convergence and heat loss at 45°S (Fig. 10c). Since almost all the anomalous active OHU (black curve in Fig. 10c) is due to the active OHT divergence (blue curve in Fig. 10c), the ocean circulation change has no contribution to the OHS change over the Southern Ocean (red curve in Fig. 10c). This may be due to the fact that the active OHU induced by the ocean dynamics takes place at a relatively slow time scale, which continues to evolve even after a century (Garuba et al. 2018a). In another word, the OHU change can always keep up with the change of OHT convergence/divergence at a slow time scale, with OHS being in a quasi-equilibrium.

Note that, although the patterns of the passive and active ocean heat budget terms are overall similar to those in Liu et al. (2018), which did not fully separate the roles of mean
ocean circulation and ocean circulation change, the quantitative results are different. We find that the mean ocean circulation contributes more than 99% to the overall increase of OHS in the Southern Ocean (red curve in Fig. 10b), while it is about 87% in their study (Fig. 10d of Liu et al. 2018). This difference indicates that the contribution of mean ocean circulation to the increase of OHS was underestimated in Liu et al. (2018). Once again, this results from the effect of buoyancy-induced ocean circulation change. The ocean circulation change is wind-driven only in the study of Liu et al. (2018), but it is the total effect of both wind-driven and buoyancy-driven circulation changes in our study. As discussed in section 3b(1), in contrast to the positive contribution of the wind-driven ocean circulation change to the OHU anomaly, the buoyancy-induced ocean circulation change plays a negative role in taking in heat from the atmosphere to the ocean around 60°S. By reducing heat gain to the ocean, the buoyancy-induced ocean circulation change has negative contribution to the increase of OHS in the Southern Ocean. This partly explain why the contribution of mean ocean circulation to the increase of OHS was underestimated in Liu et al. (2018).

c. Bjerknes compensation

Bjerknes (1964) found that, if the net radiation flux at the top of the atmosphere (netTOA) and the OHS do not change much, the total heat transport of the climate system is so constrained that the change of the heat transport in the atmosphere must compensate that of the heat transport in the ocean, which is known as Bjerknes compensation. Since the ocean circulation change is found to make little contribution to the OHS change over the Southern Ocean (red curves in Fig. 10c), we examine whether the Bjerknes compensation is present for the active component.

Figures 11a–c delineate the relationships between the atmosphere heat transport and ocean heat transport changes in the total response as well as its passive and active components. In the Southern Ocean, it is clear that the Bjerknes compensation is present under quadrupled CO2 forcing (Fig. 11a), and both the active and the passive components exhibit compensation (Figs. 11b,c). These findings are consistent with the result of Yang et al. (2018), who found that Bjerknes compensation would be present if the change in OHS were small. Specifically, it is found that the OHS increase (red curves in Figs. 10a and 10b) is only about one-tenth of the heat transport divergence change in the total response (red and blue curves in Fig. 11a) and its passive component (red and blue curves in Fig. 11b), while the active OHS (red curves in Fig. 10c) is even smaller and can be neglected.

To quantify the extent of the compensation, we calculate Bjerknes compensation rate following Zhao et al. (2016):

$$C_R = \frac{r \sigma F_o}{\sigma F_a},$$

where $r$ is the spatial correlation coefficient between the atmosphere and ocean heat transport changes, and $\sigma F_a$ and $\sigma F_o$ are standard deviations of the atmosphere heat transport change and ocean heat transport change along latitude.
respectively. A negative value of the rate ($CR < 0$) indicates a compensation between the atmosphere heat transport and ocean heat transport. Moreover, $|CR| > 1$ implies an overcompensation and $|CR| < 1$ an undercompensation, respectively.

The degree of compensation is determined by the local climate feedback (Z. Liu et al. 2016; Yang et al. 2016). Results show that the Bjerknes compensation rates south of 30°S are $-0.8$, $-1.1$, and $-1.6$ for the total response, the passive, and active components, respectively. As such, the ocean heat transport anomaly south of 30°S is overcompensated by the atmosphere heat transport change in both the passive and active components, as shown in Figs. 11b and 11c. Intriguingly, the active component of ocean heat transport change to the south of 52°S is perfectly compensated by the atmosphere heat transport change with $CR \approx 1$; that is, the change of total meridional heat transport is equal to zero (Fig. 11c).

To understand the responses of the total meridional heat transport, we examine the energy balance in both ocean and atmosphere. In the Southern Ocean, the total excess energy entering the top of the atmosphere south of 30°S is about 0.24 PW (left black bar in Fig. 11d), most of which (0.23 PW) is injected into the ocean to increase the OHS (left red bar in Fig. 11d). This increased OHS is resulted from passive netTOA (0.11 PW; middle black bar in Fig. 11d) and southward passive meridional heat transport (0.12 PW; middle gray bar in Fig. 11d), while the change of active netTOA (0.13 PW; right black bar in Fig. 11d) is totally transported northward by the active meridional heat transport (0.13 PW; right gray bar in Fig. 11d) and thus has no contribution to the OHS change. Moreover, a nearly perfect compensation occurs in the active response south of 52°S, where both the netTOA and OHS have no significant changes (not shown).

4. Verification with the FAFMIP experiments

As discussed above, while Liu et al. (2018) showed that the wind-driven ocean circulation change contributes about one-third to the heat gain around 60°S, our analysis finds that the total ocean circulation change makes little contribution. We thereby infer that the buoyancy-induced ocean circulation change should play a negative role in the heat uptake there. The FAFMIP experiments can be used to verify this inference.

In the FAFMIP experiments, a change of the OHU (surface integral of the downward heat fluxes) due to the buoyancy-induced ocean circulation change can be obtained by adding those from the faf-heat and faf-water experiments together. Figure 12 shows that it is indeed that the buoyancy-induced ocean circulation change leads to a heat loss from the ocean to the atmosphere around 60°S, which is against the heat gain due to the wind-driven ocean circulation change (Fig. 9c of Liu et al. 2018). This confirms that the buoyancy-induced ocean circulation change plays a negative role in taking in the heat from the atmosphere to the ocean around 60°S. By reducing heat gain to the ocean, the buoyancy-induced ocean circulation change has negative contribution to the increase of OHS in the Southern Ocean high latitudes.

In addition to modifying the absorption of heat, the ocean circulation change also plays a role in transporting and storing the heat in the ocean (i.e., the redistributive effect of ocean dynamics). By comparing the faf-passiveheat and faf-heat experiments, one can isolate the ocean heat redistribution due to the thermally forced ocean circulation change (section 2b;
The resultant redistributive effect on ocean temperature by the thermally forced ocean circulation change in two climate models is shown in Fig. 13. It is found that the heat flux–induced ocean circulation change results in a significant subsurface cooling anomaly in the Southern Ocean and a near-surface warming equatorward of 45°S. The subsurface cooling centered around 45°S is in stark contrast to the subsurface warming there.
produced by the wind-driven ocean circulation change (Fig. 7b of Liu et al. 2018). This gives further support to our notion that the buoyancy-induced ocean circulation change acts to decrease the OHS of the Southern Ocean. A corollary is that the contribution of mean ocean circulation to the increase of OHS was underestimated by Liu et al. (2018) wherein the cooling effect of the heat flux–induced ocean circulation change was inadvertently attributed to the mean ocean circulation.

5. Conclusions

The focus of this study is to isolate and quantify the relative contributions of mean circulation and ocean circulation change to Southern Ocean heat uptake, transport, and storage within an atmosphere–ocean coupled system. To this end, we analyze a pair of fully and partially coupled quadrupled-CO2-forced simulations designed and performed by Garuba et al. (2018a) and Garuba and Rasch (2020). In the partially coupled experiment, since the effect of ocean circulation change and its feedback to air–sea interaction are disabled, the oceanic response to greenhouse forcing can only be attributed to the passive advective process through the heat flux at the air–sea interface and the mean ocean circulation, and it is thus referred to as the atmosphere-driven passive component. In contrast, the contribution of the ocean circulation change, inferred as the residual of the full response minus the passive component, is the ocean-dynamic-induced active component.

In response to quadrupled CO2, the Southern Ocean primarily uptakes excess heat around 60°S, which is redistributed by the northward OHT and then released back to the atmosphere around 45°S, with the residual heat being stored in the ocean. The structures of the OHT and OHU are consistent with previous studies (Frölicher et al. 2015; Marshall et al. 2015; Morrison et al. 2016; Armour et al. 2016; Liu et al. 2018; Hu et al. 2020). The most important advance in our work is that we can cleanly separate the roles of mean ocean circulation and circulation change through the above processes within an atmosphere–ocean coupled system. The main new findings are as follows:

(i) For the OHU response, the mean ocean circulation and ocean circulation change are of equal importance. It is clear that neither the passive component alone (i.e., mean ocean circulation) nor the active component alone (i.e., ocean circulation change) can account for the overall pattern of the OHU change. The mean ocean circulation dominates the OHU response south of ~50°S and the ocean circulation change dominates the OHU response north of ~50°S in the Southern Ocean. Further, the positive passive OHU anomaly around ~60°S can be attributed to the positive sensible heat flux induced by a combination of the faster warming in the atmosphere than the ocean, the positive radiative flux to the reduced sea ice reflection, and positive melt heat flux to the retreating sea ice margin. The negative active OHU anomaly around ~45°S arises mainly from the negative latent heat flux driven by the positive SST anomaly through the Newtonian cooling effect.

(ii) The OHT response south of 50°S is mainly determined by advection of the mean ocean circulation, while the ocean circulation change generates an anomalous southward OHT north of 50°S. Although the magnitude of the active OHT change is small, it plays an important role in the convergence of OHT around 45°S, which balances the heat loss to the atmosphere there. All processes—Eulerian-mean flow, eddies, and diffusion—contribute to the active OHT change.

(iii) Regarding the temperature response, while both the mean ocean circulation and ocean circulation change are involved in generating the overall magnitude of Southern Ocean warming, it is the ocean circulation change that produces the subsurface warm tongue between 40° and 50°S. The passive temperature change can be attributed to the heating from positive surface heat fluxes around 60°S and the distribution by climatological ocean circulation (i.e., northward Ekman transport near the surface and subduction into the subsurface ocean at midlatitudes). In addition, the active temperature component features a subsurface warming south of 40°S, which results from a wind-driven downwelling anomaly. Similarly, the subsurface cooling of the active temperature component north of 40°S results from a wind-driven upwelling anomaly.

(iv) A heat budget analysis finds that the mean ocean circulation and ocean circulation change dominate the heat balance to the south and north of ~50°S, respectively. In particular, the divergence of passive OHT acts to balance the passive surface heat gain to the south of ~50°S, while the convergence of active OHT acts to balance the active surface heat loss to the north of ~50°S. In addition, almost all the increase in the zonal mean OHS arises from the passive component, and the contribution of the active component is negligible.

(v) Both the active and passive ocean heat transports are overcompensated by the reverse atmospheric heat transport via the Bjerknes compensation south of 30°S. Intriguingly, there is a perfect compensation for the active component of ocean heat transport south of 52°S.

Some passive and active responses identified by our partial coupling method are consistent with earlier studies using the ocean-only framework, increasing our confidence in the decomposition approach utilized here. For example, our study confirms previous findings: (i) the heat is absorbed and stored primarily as a passive tracer in the Southern Ocean (Marshall et al. 2015; Armour et al. 2016; Gregory et al. 2016); (ii) the total OHT response in the Southern Ocean is determined by its passive component (i.e., through the advection by the mean ocean circulation; Armour et al. 2016; He et al. 2019; Li et al. 2021); and (iii) it is the ocean circulation change that generates the subsurface warm tongue between 40° and 50°S (Fyfe et al. 2007; Cai et al. 2010; Liu et al. 2018; Lyu et al. 2020).

The novel decomposition in this study does add some new insights into the relative roles of the mean ocean circulation and the circulation change in the Southern Ocean heat
uptake, transport, and storage. First and foremost, all the heat storage increase in the Southern Ocean is due to the passive component and the contribution of the ocean circulation change is negligible. This finding is in stark contrast to what is found in the previous studies (Winton et al. 2013; Garuba and Klinger 2016; Chen et al. 2019; Hu et al. 2020), which have attributed the pattern of the Southern Ocean heat storage change predominantly to the ocean circulation changes. A previous study of Liu et al. (2018) has attempted to isolate the roles of mean ocean circulation and ocean circulation change in the Southern Ocean heat uptake and storage through an overriding technique in the fully coupled climate model. However, since the buoyancy-induced ocean circulation change cannot be isolated from the mean ocean circulation in their experimental design, some of the conclusions they reached are quite different from ours in this study. One of the main differences is about the attribution of the peak heat gain around 60°S. It is attributed to the combined effect of the direct CO2-induced warming and wind-driven ocean circulation change in Liu et al. (2018), while we find that the ocean circulation change makes little contribution to the heat uptake there. This indicates that the buoyancy-induced ocean circulation change plays a negative role in taking in the heat from the atmosphere to the ocean around 60°S. By isolating the buoyancy-induced ocean circulation changes to the OHU and OHS through the analysis of the FAFMIP experiments from CMIP6, our study for the first time identifies the negative role of the buoyancy-induced ocean circulation changes in the OHU around 60°S, countering the positive OHU there through the mean ocean circulation and the wind-induced ocean circulation changes.

Finally, we would like to point out that the CESM employed for the experiments does not resolve mesoscale eddies in the ocean, which is anticipated to bring about at least quantitative difference in regard to the Southern Ocean heat uptake and heat storage, especially in view of the fact that eddies in the Southern Ocean are responsible for a large portion of the total energy transport (Abernathey et al. 2011; Volkov et al. 2008, 2010; Sun et al. 2019). This calls for the implementation of the similar partial coupling technique to the eddy-resolving climate models as the immediate next step.

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Data availability statement. The CESM data used in this study are available from the corresponding authors upon request. The CMIP6 FAFMIP data are publicly available through the Earth System Grid Federation (ESGF) archive (https://esgf-node.llnl.gov/projects/esgf-llnl/).

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