**The impacts of cloud-radiative changes on poleward atmospheric and oceanic
energy transport in a warmer climate**

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Abstract

Based on theory and climate model experiments, previous studies suggest most of the uncertainties in projected future changes in meridional energy transport and zonal mean surface temperature can be attributed to cloud feedback. To investigate how radiative and dynamical adjustments modify the influence of cloud-radiative changes on energy transport, this study applies a cloud-locking technique in a fully-coupled climate model, CESM. Under global warming, the impacts of cloud-radiative changes on the meridional energy transport are asymmetric in the two hemispheres. In the Northern Hemisphere, the cloud-radiative changes have little impact on energy transport, because 89% of the cloud-induced heating is balanced locally by increasing outgoing longwave radiation. In the Southern Hemisphere, on the other hand, cloud-induced dynamical changes in the atmosphere and the ocean cause enhanced poleward energy transport, accounting for most of the increase in energy transport under warming. Our experiments highlight that the local longwave radiation adjustment induced by temperature variation can partially offset the impacts of cloud-radiative changes on energy transport, making the estimated impacts smaller than those obtained from directly integrating cloud-radiative changes in previous studies. It is also demonstrated that the cloud-radiative impacts on temperature and energy transport can be significantly modulated by the oceanic circulation, suggesting the necessity of considering atmospheric-oceanic coupling when estimating the impacts of cloud-radiative changes on the climate system.

1. Introduction

Meridional energy transport plays an important role in modulating the response of the climate system to external forcings. Despite its important influence on regional climate, the response of meridional energy transport to increasing CO₂ concentration exhibits large uncertainty across climate models (Huang and Zhang 2014; Zelinka and Hartmann 2012). A large portion of the inter-model diversity is attributed to the uncertainty in cloud-radiative changes, based on diagnostics of global climate model (GCM) outputs (Huang and Zhang 2014; Zelinka and Hartmann 2012) or highly simplified models (Armour et al. 2019; Bonan et al. 2018; Frierson and Hwang 2012; Hwang and Frierson 2010; Roe et al. 2015) that do not fully account for radiative and dynamical adjustments. In this study, we investigate these adjustment processes, with the goal of understanding the impacts of the cloud-radiative changes under warming on energy transport and meridional temperature structure.

1.1. Meridional energy transport under unperturbed and warmer climate

From an energetic perspective, meridional energy transport is tightly linked to the radiation imbalance at the top-of-atmosphere (TOA) by the zonal-mean energy budget of an atmosphere-ocean column,

$$\frac{dH}{dt} = R^{TOA} - \nabla \cdot MET, \quad (Eq. 1)$$

where $\frac{dH}{dt}$ represents the heat content tendency of the atmosphere and ocean. R^{TOA} represents the downward net TOA fluxes, and MET represents the meridional energy transport. Ignoring the

tendency term, the connection between meridional energy transport and net TOA fluxes can be obtained by integrating Eq. 1 from the South Pole to a latitude ϕ ,

$$MET(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} \hat{R}^{TOA} a^2 \cos\phi d\lambda d\phi, \quad (Eq. 2a)$$

where a represents the radius of the earth, λ represents longitude, and the hat indicates anomalies from the global mean.

Under global warming, the meridional structure of net TOA fluxes changes due to the greenhouse gas (GHG) forcing and a variety of local feedbacks, implying changes in meridional energy transport. Global climate models predict a robust increase in poleward atmospheric energy transport under global warming (Huang and Zhang 2014; Zelinka and Hartmann 2012). In the tropics, the increase of poleward atmospheric energy transport is attributed to the increase in dry static energy transport in the upper branch of the Hadley cell, which is associated with the increased depth of the Hadley cell and the reduced lapse rate under warming; in the mid-latitudes, the increase of poleward atmospheric energy transport is associated with the latent energy transported by eddy mixing (Donohoe et al. 2020; He et al. 2019; Held and Soden 2006; Hwang and Frierson 2010; Wu et al. 2011). GCMs also predict a robust decrease in poleward oceanic energy transport that partially offsets the anomalous atmospheric energy transport (Huang and Zhang 2014). In the tropics, the decrease in poleward oceanic energy transport is attributed to the slowdown of the subtropical cell in the Indo-Pacific basin (He et al. 2019; Huang and Zhang 2014); in the Northern Hemisphere extra-tropics, the anomalous southward energy transport is suggested to be associated with the slowdown of the Atlantic Meridional Overturning Circulation (Hazeleger, 2005; Stouffer et al., 2006), while in the Southern

Hemisphere extra-tropics, the anomalous northward energy transport may be attributable to the advection of temperature anomalies carried by the climatological oceanic circulation (Armour et al. 2016). The total meridional energy transport, however, has a large inter-model spread, with the multi-model mean values showing an increased poleward meridional energy transport (Huang and Zhang 2014).

1.2. Attributing the anomalous meridional energy transport to different feedback processes

To better understand the source of uncertainty in the anomalous meridional energy transport under warming, previous studies (e.g. Huang and Zhang 2014; Zelinka and Hartmann 2012) have attributed the anomalous meridional energy transport to different feedback processes through Eq. 2b below.

$$\Delta MET(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} (\Sigma_i \widehat{\Delta R}^i + \widehat{F}_g) a^2 \cos \phi d\lambda d\phi$$

$$\Delta MET^i(\phi) \equiv \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} \widehat{\Delta R}^i a^2 \cos \phi d\lambda d\phi, \quad (Eq. 2b)$$

where F_g represents the forcing, and ΔMET^i represents the implied changes in energy transport corresponding to a certain feedback process $\widehat{\Delta R}^i$. Among all the feedback processes, the cloud feedback, which represents the cloud-radiative changes associated with a 1K warming of global mean temperature, has been suggested to strongly contribute to the inter-model spread in energy transport anomalies (Huang and Zhang 2014; Hwang and Frierson 2010; Zelinka and Hartmann 2012). The multi-model mean values suggest cloud feedback increases the poleward energy

transport in the mid-latitudes of both hemispheres and induces a southward cross-equator energy transport in CMIP3 (Zelinka and Hartmann 2012) and CMIP5 models (Huang and Zhang 2014), although the inter-model spread is large. The multi-model mean enhancement of poleward energy transport associated with cloud feedback may be explained by the robust LW cloud feedback in the tropics and negative SW cloud feedback in mid-to-high latitudes (Ceppi et al. 2016a, 2016b; Gettelman and Sherwood 2016; Hartmann and Larson 2002; Zelinka et al. 2012a, 2012b), which enhances the equator-to-pole energy gradient and implies anomalous poleward energy transport.

Decomposing the total changes in energy transport into components due to different feedbacks improves our understanding of the physical mechanisms that affect the energy budget. This method has the advantage that the partitioning is always exact: as the summation of feedbacks and forcing will always equal the net TOA fluxes, so will their implied energy transport. However, the approach does not consider the interactions among feedback processes. In other words, for a certain feedback process, the anomalous heating relative to its global mean value must be balanced by the relative cooling in other regions regardless of the distance. This is not necessarily true, however, as the heating induced by a certain feedback process can also interact locally with other feedbacks. For example, when there is heating due to cloud-radiative changes within a latitude band, the temperature will rise, increasing outgoing longwave radiation (OLR) cooling to offset the original cloud-induced heating. Based on this implied energy transport calculation, the increased OLR cooling would be attributed to the temperature feedback. However, as the OLR cooling is caused by the cloud-induced heating, it could also be considered as a response to cloud feedback. In the situation that a large portion of the cloud-radiative

heating is offset by the OLR cooling, the responses of energy transport would be much smaller than predicted by the implied energy transport.

To account for the local interactions among feedback processes, an alternative approach for estimating the influence of anomalous TOA fluxes on meridional energy transport is by using a diffusive energy balance model (EBM, e.g. Roe et al. 2015; Rose et al. 2014), which is a one-dimensional diffusive atmospheric model that predicts the responses of surface temperature and atmospheric energy transport to external forcing. At each grid box, the energy imposed into the model is partially offset by the anomalous TOA flux that is parameterized by anomalous surface temperature. In its simplest version, the EBM parameterizes surface temperature (Planck) feedback by assuming a globally uniform dependence of anomalous OLR on temperature anomalies, and the anomalous TOA fluxes caused by other feedback processes are treated as a prescribed forcing in the EBM. With this EBM setting, which considers the local interaction between Planck feedback and other feedback processes when attributing the changes in energy transport, the cloud feedback is found to increase the poleward energy transport in the mid-latitudes, albeit with a large inter-model spread (Frierson and Hwang 2012; Hwang and Frierson 2010). Some recent studies modify the EBM to consider a latitude-dependent feedback parameter, which is determined by the ratio between anomalous TOA fluxes and local surface temperature anomalies, representing the joint impact of all feedback processes (Armour et al., 2019; Bonan et al., 2018; Roe et al., 2015; see section 2.3 for details). In response to CO₂ forcing, the influences of a certain feedback are examined by varying the feedback parameter, to mimic the changes in that specific feedback's magnitude and meridional structure. It is found that the inter-model spread of feedback strength in the lower latitudes strongly influences the inter-model spread of energy transport, which results in uncertainty in temperature changes

globally; at higher latitudes, on the other hand, the inter-model spread of feedback strength causes uncertainty in local temperature anomaly (Armour et al., 2019; Bonan et al., 2018; Roe et al., 2015). In particular, cloud feedback is suggested to be the most important contributor to the uncertainty in the meridional pattern of warming (Bonan et al. 2018).

Despite offering great insights, the assumption of a globally uniform diffusivity may lead to biases in its estimation of energy transport (Peterson and Boos 2019). In addition, although the radiative responses associated with each feedback are allowed to interact through their impacts on surface temperature, the magnitudes and structures of the feedbacks themselves are pre-determined in the EBM and are not interactive. For example, the surface temperature anomalies caused by varying cloud feedback may be amplified or damped by water vapor feedback, but the magnitude and meridional structure of water vapor feedback would not change because it is prescribed. Moreover, the EBM inhibits the interactions among feedbacks and oceanic processes. Many recent studies point to the important role of ocean dynamics or ocean heat content in modulating the atmospheric energy transport (Chiang and Friedman 2012; Green and Marshall 2017; Green et al. 2019; Hawcroft et al. 2017; He et al. 2019; Kang et al. 2017, 2018; Kay et al. 2016; Schneider 2017; Yu and Pritchard 2019). To fully account for the dynamical and thermodynamic impacts of cloud feedback, a “cloud-locking” method, in which the cloud-radiative feedback is disabled, is used in some recent studies (more details in Section 2.1; Albern et al. 2019, 2020; Ceppi and Hartmann 2016; Ceppi and Shepherd 2017; Middlemas et al. 2019, 2020; Trossman et al. 2016; Voigt et al. 2019; Voigt and Shaw 2016). It is found that the cloud-radiative changes under warming lead to the poleward shift of mid-latitude jets (Albern et al. 2019, 2020; Ceppi and Hartmann 2016; Voigt et al. 2019; Voigt and Shaw 2016), enhanced poleward atmospheric energy transport (Trossman et al. 2016), and can lead to amplified

warming in the Arctic region via remote impacts (Middlemas et al. 2020). Trossman et al. (2016) demonstrated that the oceanic processes can interact with cloud feedback: while the oceanic processes act to slow down the pace of warming, they also stabilize the lower troposphere and help to maintain the low cloud amount in the mid-latitudes of the Northern Hemisphere, resulting in less-positive cloud feedback. Building upon previous studies investigating the impacts of cloud-radiative changes on the climate system, our study documents the impacts of cloud-radiative changes on energy transport in a fully-coupled GCM, and provides a linkage to the conventional estimation of implied energy transport.

2. Methodology

2.1 Experimental design

We use CESM1 (Hurrell et al., 2013) with the f19_g16 grid of approximately 2° horizontal resolution in the atmospheric model (CAM5) and approximately 1° in the ocean model (POP2). The cloud-locking method is applied (Ceppi and Hartmann 2016; Mauritsen et al. 2013; Voigt and Shaw 2016), in which the radiative properties of clouds are prescribed rather than being interactive in the radiation calculation. By doing so, the cloud-radiative properties are always identical to those in either the unperturbed simulations or the warming simulations, which allows us to isolate the cloud-driven changes and the non-cloud-driven changes under warming.

The cloud-locking method consists of two steps. First, a 50-year-long control simulation (CTL) is performed with pre-industrial conditions and interactive clouds, along with a 30-year-long warming experiment ($4\times\text{CO}_2$) in which the CO_2 concentration is quadrupled abruptly. The time scale of 30 years of the warming experiments allows one to investigate the “fast response” of global warming, during which the atmosphere and shallow wind-driven oceanic circulation reach their quasi-equilibrium state while the deep ocean is still slowly adjusting. In these simulations, at every radiative time step, the instantaneous cloud properties that are necessary for the radiation calculation are saved. These variables include liquid and ice cloud fraction, in-cloud liquid/ice/snow water path, effective diameter for ice and snow, and size distribution parameters of liquid drops. The initial conditions of these simulations are obtained from another unforced simulation with pre-industrial conditions, in which the radiation and dynamical fields are coupled every hour (the default CESM setting). To reduce the data size of cloud properties needed for the cloud-locking method, in all simulations used in this study, the radiation module

is called every two hours instead of every hour. The changes in radiative coupling time result in a climate drift of -0.05 K/decade of global mean temperature in the first 30 years of the CTL simulation. As the signals of climate drift should exist in all experiments, they are expected to cancel out, at least in part, when two simulations are differenced.

Next, a set of pre-industrial and warming simulations with several ensemble members are performed, summarized in **Table 1**. In these simulations, the cloud properties from the CTL or 4xCO₂ simulations saved beforehand are used for the radiation calculations, instead of the interactive cloud properties. Therefore, in these simulations, the clouds are decoupled from the background climate. The impacts of cloud-radiative changes on the climate system can then be obtained by

$$\Delta X_{cloud} = \frac{1}{2} [(X_{F1C2} - X_{F1C1}) + (X_{F2C2} - X_{F2C1})]. \quad (Eq. 3a)$$

where X represents the field of a certain variable. The subscripts of F and C denote forcing and cloud properties, respectively: $F1$ and $C1$ correspond to the CO₂ concentration and cloud properties in the CTL simulation, while $F2$ and $C2$ correspond to those in the 4xCO₂ simulations. The two terms on the right-hand side of Eq. 3a represent the differences between the two experiments under the same CO₂ scenarios, while the cloud-radiative properties are fixed to two different climate evolutions. We define the “cloud impacts” on X as ΔX_{cloud} , which can be thought of as the responses of X to the cloud-radiative changes under warming in a coupled climate system in which the atmospheric and oceanic processes are free to adjust. The cloud-radiative changes are estimated as the kernel-adjusted cloud-radiative effects (Appendix A1), demonstrated in Fig. 1.

Similarly, the “non-cloud impacts” on X are defined as

$$\Delta X_{non_cloud} = \frac{1}{2}[(X_{F2C1} - X_{F1C1}) + (X_{F2C2} - X_{F1C2})], \quad (Eq. 3b)$$

which represent the responses of X to increased CO_2 concentration without cloud-radiative changes. The total warming response under the prescribed-cloud framework can be expressed as

$$\Delta X_{total} = \Delta X_{cloud} + \Delta X_{non_cloud} = X_{F2C2} - X_{F1C1}, \quad (Eq. 3c)$$

where ΔX_{total} represents the responses of X under warming, with the nonlinear interaction between cloud-radiative changes and the background environment ignored. When comparing the total responses to quadrupled CO_2 concentration under the prescribed-cloud framework (Eq. 3c) to those in the standard interactive-cloud simulations ($4xCO_2 - CTL$), the prescribed-cloud framework captures the zonal-mean responses of surface temperature and meridional energy transport reasonably well (Figs. 2a, 2b), suggesting the nonlinear interaction between cloud-radiative changes and background environment is weak.

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2.2 Energy transport

2.2.1 Atmospheric energy transport

The total atmospheric energy transport can be estimated indirectly by the atmospheric energy budget,

$$AET(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} (\hat{R}^{TOA} - \hat{Q}^{sfc}) a^2 \cos\phi d\lambda d\phi, \quad (Eq. 4) \quad (\text{Hartmann 1994})$$

where Q^{sf_c} represents downward surface flux. By assuming that the local energy gain of the atmosphere is balanced by the energy transport divergence, this method provides an accurate estimation of the atmospheric energy transport, as the atmospheric heat content tendency is usually negligible over sufficiently long timescales.

Another approach to estimating the atmospheric energy transport, which can be used to separate the contributions from mean flow and eddies, is by zonally and vertically integrating the moist static energy (MSE) transport,

$$AET_{MSE}(\phi) = \int_0^{P_s} \int_0^{2\pi} v MSE a \cos\phi d\lambda \frac{dp}{g}, \quad (Eq. 5a)$$

where p represents pressure, P_s represents the surface pressure, v is the meridional velocity, and MSE is defined by

$$MSE = C_p T + gz + L_v Q,$$

where the three terms on the right-hand side represent the internal energy, potential energy, and latent energy, respectively. The MSE transport contributed by the mean-flow circulation can be written as

$$AET_{meanflow}(\phi) = \int_0^{PS} \int_0^{2\pi} \overline{v} \overline{MSE} a \cos\phi d\lambda \frac{dp}{g}, \quad (Eq. 5b)$$

where the overbar indicates the mean seasonal cycle of monthly data. The mean-flow transport with seasonal variation is calculated using monthly outputs, and the annual average is taken

afterward. To eliminate the spurious vertically-integrated mass transport (v_s) arising from the procedure of interpolating the CESM outputs from a sigma-pressure vertical coordinate to a pressure coordinate, we remove the non-conserved mass transport in $\bar{v}$ following Yang et al. (2015),

$$\bar{v}_s = \int_0^{P_s} \int_0^{2\pi} \bar{v} a \cos \phi \, d\lambda \frac{dp}{g} / \int_0^{P_s} \int_0^{2\pi} a \cos \phi \, d\lambda \frac{dp}{g},$$

$$AET_{meanflow}^*(\phi) = \int_0^{PS} \int_0^{2\pi} (\bar{v} - \bar{v}_s) \overline{MSE} a \cos \phi \, d\lambda \frac{dp}{g}. \quad (Eq. 5c)$$

After removing the spurious mass transport, Yang et al. (2015) showed that the climatological atmospheric energy transport obtained by the integrated MSE transport (Eq. 5c) is close to that obtained by the energy budget method (Eq. 4) in CESM. In this study, we estimate the total atmospheric energy transport by Eq. 4, and estimate the contribution of mean-flow transport by Eq. 5c. The eddy-induced atmospheric energy transport is estimated as the difference between the total atmospheric energy transport and mean-flow transport,

$$AET_{eddy} = AET - AET_{meanflow}^*. \quad (Eq. 5d)$$

To further understand the changes in eddy-induced atmospheric energy transport, we assume that, based on a diffusive perspective, the eddy-induced atmospheric energy transport is proportional to the near-surface MSE gradient (Mooring and Shaw 2020),

$$AET_{eddy} = a \cos \phi D^* MSE_y, \quad (Eq. 6a)$$

where MSE_y is the zonal-mean meridional MSE gradient at 900 hPa, and D^* is the diffusivity (kg/s), which represents the rate of mass exchange.

We calculate the diffusivity D^* by the ratio of eddy-induced energy transport to the zonal-mean meridional MSE gradient, divided by the circumference of the latitude. We can further relate the cloud impacts on anomalous eddy-induced energy transport to the changes in MSE gradients and diffusivity:

$$\Delta(AET_{eddy})_{cloud} =$$

$$a \cos \phi [D^* \Delta(MSE_y)_{cloud} + \Delta D^*_{cloud} MSE_y + \Delta D^*_{cloud} \Delta(MSE_y)_{cloud}]. \quad (\text{Eq. 6b})$$

2.2.2 Oceanic energy transport

CESM resolves oceanic energy transport caused by Eulerian-mean transport and parameterizes the energy transport due to mesoscale eddies, sub-mesoscale eddies, and diffusion. The total oceanic energy transport is the summation of the energy transport caused by the above processes. In this study, all the terms other than the Eulerian-mean transport are combined and defined as sub-grid processes.

The Eulerian-mean transport, calculated as the product of velocity (V) and potential temperature (T), can be decomposed to

$$\overline{VT} = \overline{V} \overline{T} + \overline{V'T'}, \quad (\text{Eq. 7a})$$

where the prime indicates a departure from the mean seasonal cycle. The first term on the right-hand side of Eq. 7a is defined as mean-flow transport. The second term, which is associated with large-scale eddies, is defined as eddy-induced transport.

The cloud impacts on the anomalous mean-flow transport can be further decomposed to

$$\Delta(\overline{V} \overline{T})_{cloud} = \overline{V} \Delta \overline{T}_{cloud} + \overline{T} \Delta \overline{V}_{cloud} + \Delta \overline{V}_{cloud} \Delta \overline{T}_{cloud}. \quad (Eq. 7b)$$

The three terms on the right-hand side of Eq. 7b represent the contributions of anomalous oceanic energy transport due to changes in temperature, changes in circulation, and the nonlinear term, respectively.

2.3 Descriptions of the EBM

Similar to that used in earlier works (Armour et al., 2019; Bonan et al., 2018; Roe et al., 2015), the EBM used here is a one-dimensional diffusive model, predicting the response of temperature and energy transport to a given prescribed forcing F_{EBM} ,

$$F_{EBM} + \lambda \delta T_s = -D^* \frac{\partial^2 \delta MSE_y}{\partial y^2}. \quad (Eq. 8)$$

In response to the net forcing F_{EBM} , the EBM predicts the anomalous TOA fluxes caused by the local surface temperature anomalies ($\lambda \delta T_s$) and the divergence of atmospheric energy transport (right-hand side of Eq. 8a), both of which are parameterized to be functions of surface

temperature anomalies δT_s . The latitude-dependent radiative feedback λ represents the
 anomalous TOA fluxes corresponding to a unit change of surface temperature, which is
 estimated by decomposing the anomalous radiative fluxes under warming with radiative kernels
 (Appendix A1). A key difference between the EBM setting in this study and previous studies
 (Armour et al. 2019; Bonan et al. 2018; Roe et al. 2015) is in the calculation of feedback
 parameter λ . To be consistent with our GCM simulations in which the cloud feedback is
 deactivated, we estimate λ by the differences in radiative fluxes between F2C1 and F1C1
 simulations (e.g. warming without cloud feedback; **Table 1**), in contrast to the conventional way
 to estimate λ by the anomalies in a 4xCO₂ simulation. Therefore, λ includes feedbacks by
 temperature, water vapor, surface albedo, and the residual term in the kernel decomposition. The
 specific calculation method for λ is not critical to our conclusions. Diagnosing λ from the
 standard interactive-cloud simulations (4xCO₂) and excluding cloud feedback from the net
 feedback λ would yield similar EBM responses. The EBM assumes a diffusive atmospheric
 energy transport that is proportional to the anomalous meridional MSE gradient at the surface
 (δMSE_y) and the diffusivity D^* . A fixed relative humidity of 80% is assumed when calculating
 MSE. A uniform diffusivity D^* of 9.6×10^9 kg/s is used, following Armour et al. (2019) and
 Bonan et al. (2018). The results are not sensitive to the value of D^* . Note that the diffusivity D^*
 defined here differs from what in Eq. 6b: the former is pre-determined and is assumed to be
 uniform globally, which makes our results more comparable with previous studies; the latter is
 diagnosed from the GCM simulations to understand the changes in eddy-induced energy
 transport in the atmosphere.

3. Results

First, we examine the responses of the climate system to increased CO₂ concentration in CESM. In response to quadrupled CO₂ concentration, the surface temperature increases everywhere and features amplified warming in the polar regions, which is mostly contributed by non-cloud impacts at all latitudes (Fig. 2). The cloud impacts enhance the hemispheric asymmetry of warming, contributing to polar amplification in the Northern Hemisphere and having little impact in the Southern Hemisphere. This decomposition of surface temperature changes is consistent with the results in Middlemas et al. (2020).

The responses of energy transport under warming are characterized by the enhanced poleward atmospheric energy transport and decreased poleward oceanic energy transport that partially offset each other (Fig.2b), consistent with previous studies (Donohoe et al. 2020; He et al. 2019; Held and Soden 2006; Huang and Zhang 2014; Hwang and Frierson 2010; Wu et al. 2011; Zelinka and Hartmann 2012). The anomalous net energy transport is southward at most latitudes. This anomalous meridional energy transport (Fig. 2b) is close to the multi-model mean values of CMIP5 models in the Southern Hemisphere; in the Northern Hemisphere, the anomalous meridional energy transport in CESM1 is more southward than in most of the CMIP5 models (Fig. 2d).

When decomposed into the contributions of non-cloud and cloud impacts, the non-cloud impacts again show compensating changes in atmospheric and oceanic energy transport, similar to the total warming responses (Fig. 2c). The cloud impacts, on the other hand, are associated with an anomalous southward atmospheric and oceanic energy transport that reinforce each other in the Southern Hemisphere, while having little impact on energy transport in the Northern Hemisphere

northward of 30°N. The anomalous net energy transport under warming is contributed by the non-cloud impacts in the Northern Hemisphere, while the cloud impacts dominate the anomalous energy transport under warming in the Southern Hemisphere.

As shown in Fig. 2c, the anomalous net energy transport in response to global warming would be very different if there were no cloud-radiative changes (i.e. only considering the non-cloud impacts), especially in the Southern Hemisphere, highlighting the importance of the cloud-radiative changes to the meridional structure of anomalous energy transport under global warming. In the following sections, we will investigate the cloud impacts on energy transports in more detail.

3.1 Energetic perspective on the cloud impacts on meridional energy transport

In response to the cloud-radiative changes, there is a significant net southward energy transport in the Southern Hemisphere and negligible changes in transport in the Northern Hemisphere extra-tropics. In this section, we perform an energy budget analysis to understand this latitudinal dependency.

3.1.1 Local energy balance

Recalling the TOA energy budget written in Eq. 1, the cloud impact on the energy budget can be expressed as

$$\Delta \frac{dH}{dt}_{cloud} = \Delta R^{cld}_{cloud} + \Delta R^{clr}_{cloud} - \Delta(\nabla \cdot MET)_{cloud}, \quad (Eq. 9a)$$

where R^{cld} represents the cloud-radiative changes, and R^{clr} is defined as the difference between net TOA fluxes and R^{cld} (Appendix A1). Note the difference between the cloud subscript and the cld/clr superscript: the former refers to the impacts of clouds on coupled climate estimated by the cloud-locking experiments (Eq. 3a), while the latter only describes the nature of the radiative anomalies that are estimated by radiative kernels. For example, ΔR^{clr}_{cloud} denotes the clear-sky radiative change caused by the climate response to cloud feedback.

We approximate $\frac{dH}{dt}$ by the tendency of ocean heat content, which is calculated as the difference between zonal-mean surface flux and divergence of oceanic energy transport. Rearranging Eq. 9a yields

$$\Delta R^{cld}_{cloud} + \Delta Adj_{cloud} - \Delta(\nabla \cdot MET)_{cloud} = 0, \quad (Eq. 9b)$$

$$\Delta Adj_{cloud} \equiv \Delta R^{clr}_{cloud} - \Delta \frac{dH}{dt}_{cloud}.$$

In Eq. 9b, the changes in R^{clr} and local heat content are combined as a local adjustment term (ΔAdj_{cloud}), and the new equation represents the balance among the cloud-radiative changes, energy transport divergence, and local adjustments. Since the cloud properties are prescribed in the simulations, the cloud-radiative changes, ΔR^{cld}_{cloud} , can be viewed as a forcing imposed on the climate system, which is shown in Fig. 3a. The zonal structure of the net cloud-radiative changes features positive TOA fluxes that warm the climate system between 40°S-60°N and peak in the Northern Hemisphere mid-latitudes around 40°N. In the Southern Hemisphere high

latitudes, the net cloud-radiative changes result in negative TOA fluxes that cool the climate system. In response to the local cloud-radiative change — for example, a positive cloud-radiative change — the local energy balance can be achieved by increasing heat content and OLR locally (anomalously negative ΔAdj_{cloud}) and by transporting energy away from the region (anomalously positive $\Delta(\nabla \cdot MET)_{cloud}$). As shown in Fig. 3a, the relative importance of local adjustments and energy transport convergence in balancing the net cloud-radiative changes depends on latitude. In the Northern Hemisphere, the strong negative adjustment term compensates the energy transport divergence that would otherwise occur, while the anomalous energy transport plays a dominating role in balancing the cloud-radiative changes in the Southern Hemisphere.

3.1.2 Implied energy transport

In this section, we link the latitudinal structures of local energy flux changes discussed in Section 3.1.1 to the anomalous energy transport by Eq. 10 below, and further compare the implied energy transport induced by cloud-radiative changes with the cloud impacts on energy transport diagnosed from our cloud-locking simulations. Integrating Eq. 9b from the South Pole to a certain latitude yields

$$\Delta MET_{cloud}(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} \Delta \widehat{R}_{cloud}^{cld} a^2 \cos \phi d\lambda d\phi + \int_{-\frac{\pi}{2}}^{\phi} \int_0^{2\pi} \Delta \widehat{A}d_{cloud} a^2 \cos \phi d\lambda d\phi. \quad (Eq. 10)$$

The first term on the right-hand side represents the implied energy transport due to the cloud-radiative changes, which is directly determined by the latitudinal distribution of the cloud-radiative changes that are prescribed to the simulations. The second term is the implied energy

transport due to the anomalous TOA fluxes induced by changes in temperature, water vapor, albedo, and heat content, representing the modulations caused by those local adjustment processes. In previous studies diagnosing the outputs of CMIP3 and CMIP5 models (Huang and Zhang 2014; Zelinka and Hartmann 2012), the impacts of the cloud-radiative changes on energy transport are estimated by their implied energy transport, which should equal the first term on the right-hand side of Eq. 10. The second term, however, cannot be diagnosed via standard CMIP outputs. Through performing cloud locking experiments in a fully coupled model, we allow all radiative and dynamical processes to operate when evaluating how energy transport and temperature would respond to the cloud-radiative changes.

As shown by the green line in Fig. 3b, the meridional structure of the cloud-radiative changes implies a southward energy transport at most latitudes to transport energy from the heating regions to the cooling regions. This implied energy transport is close to the actual changes in energy transport in subtropics and mid-latitudes of the Southern Hemisphere (the solid black line in Fig. 3b), while in the Northern Hemisphere and in the tropics the implied transport is larger than the actual changes in energy transport. This can be explained by the adjustment term, which shows a peak of cooling in the Northern Hemisphere mid-latitudes (Fig. 3a), offsetting the cloud-induced heating and reducing the required southward energy transport.

As demonstrated in Fig. 4b, the overall meridional structure of the adjustment term is dominated by the changes in the longwave (LW) component of R^{clr} (Fig. 4b), and further decomposition shows that it is mostly contributed by the Planck response (Fig. 4d). In the region of 20°N-60°N where the cloud-induced heating peaks, the surface warming results in a strong LW cooling, which offsets 89% of the cloud-induced heating. This LW cooling associated with surface warming implies a northward energy transport in the Northern Hemisphere, dominating the

implied energy transport due to the adjustment term. The changes in water vapor and lapse rate can also significantly contribute to the adjustment term locally, while their modulations of energy transport are much smaller compared to that of the Planck response.

In summary, from an energetic perspective, we have shown how the meridional energy transport and local adjustments act together to balance the imposed cloud-radiative changes. It is demonstrated that the responses of meridional energy transport to the cloud-radiative changes are asymmetric in the two hemispheres: in the Northern Hemisphere, the strong cloud-induced heating is largely offset by local OLR cooling; in the subtropics and mid-latitudes of the Southern Hemisphere, the cloud-radiative changes are mostly balanced by changes in energy transport, with the local adjustment processes playing a minor role. We will discuss the possible reasons for these hemispherically asymmetric responses in section 4.1. When comparing the anomalous energy transport in cloud-locking simulations with the implied energy transport of cloud-radiative changes, the latter predicts a larger energy transport anomaly in the Northern Hemisphere because the local adjustment processes are not considered; in the Southern Hemisphere, on the other hand, both methods predict a southward energy transport anomaly due to the meridional gradient of cloud-radiative changes. It is worth noting that this meridional gradient of cloud-radiative changes is dominated by its shortwave component, which may be too negative over the Southern Ocean because of insufficient supercooled liquid water there in the baseline climate, as in most of the CMIP5 models (Cesana and Chepfer 2013; Hwang and Frierson 2013; Kay et al. 2016; Trenberth and Fasullo 2010). This bias is largely reduced in the CMIP6 models, resulting in a less negative shortwave cloud feedback in the Southern Hemisphere high latitudes; meanwhile, in the Southern Hemisphere mid-latitudes, however, the

shortwave cloud feedback becomes more positive, so the meridional gradient of shortwave cloud feedback is not substantially reduced in the CMIP6 models (Zelinka et al. 2020).

3.2 Dynamical perspective on the cloud impacts on meridional energy transport

In response to the cloud-radiative changes under global warming, the most significant feature of anomalous energy transport is the strengthened poleward transport in the Southern Hemisphere. In the mid-to-low latitudes, the ocean plays a more important role in carrying additional energy than the atmosphere, while the opposite is true in high latitudes (Fig. 3b). In this section, we investigate the circulation changes underlying the anomalous atmospheric and oceanic energy transport in the Southern Hemisphere.

3.2.1 Oceanic energy transport and circulation changes

First, we examine the mechanisms accomplishing the anomalous oceanic energy transport. In the subtropics and mid-latitudes, the anomalous transport is dominated by the mean-flow advection, with negligible contributions from eddy and sub-grid processes. The mean-flow advection is accomplished via anomalous oceanic circulation (Figs. 5a, 5b, red and orange curves); in particular, the anomalous energy transport is associated with an anomalous overturning cell between 30°S and 50°S, with its upper branch tied to the upper 50 meters of the ocean, and its

lower branch extending vertically to ~800 meters (Fig. 5c¹). The anomalous mass transport in the upper layer is close to the theoretical Ekman transport estimated by the anomalous surface zonal wind stress (Fig. 5d), suggesting the upper branch of the overturning cell is wind-driven. The anomalous surface Ekman transport converges around 50°S, where the water subducts and moves equatorward following the isotherms, forming the lower branch of the overturning cell. The overall energy transport associated with this overturning cell can be thought of as the residual of the energy transport in the upper and lower branches. As the temperature is higher in the near-surface layer in the ocean, the direction of the overall anomalous energy transport induced by the overturning cell is determined by the surface Ekman flow, which is southward. The air-sea coupling through Ekman transport has been proposed by a few recent studies to explain the anomalous oceanic energy transport in response to a variety of hemispherically asymmetric forcings (Green et al. 2019; Kang et al. 2017; Kay et al. 2016; Schneider et al. 2014).

The overturning circulation in the ocean is closely linked to the anomalous atmospheric circulation (Fig. 6). In response to the cloud-radiative changes, the mid-latitude jet shifts poleward, resulting in anomalous easterlies around 30°S (Fig. 6a) that drive the anomalous southward Ekman transport in the ocean surface layer (Fig. 5d). The poleward shift of mid-latitude jet in response to the cloud-radiative changes is also reported in previous studies using GCMs with simplified ocean models, and is explained by the enhanced equator-to-pole gradient of SST (Ceppi and Hartmann, 2016; Ceppi et al. 2014) and the enhanced upper-troposphere

¹ Note that we show the Eulerian-mean circulation here, rather than the residual-mean circulation, because the oceanic energy transport is dominated by mean-flow advection in the regions we focus on.

temperature gradient (Voigt et al. 2019; Voigt and Shaw 2016) induced by the cloud-radiative changes.

3.2.2 Atmospheric energy transport and circulation changes

Next, we examine the anomalous southward atmospheric energy transport, which accounts for 1/3 of the anomalous energy transport around 30°S. Consistent with the poleward shift of the mid-latitude jet in the Southern Hemisphere, there is anomalous eddy momentum flux divergence in the regions around 30°S (Fig. 6b). An anomalous southward motion in the upper troposphere is required as the divergence of eddy momentum flux must be balanced by the Coriolis acceleration. Therefore, a counter-clockwise overturning circulation is driven around 30°S (Fig. 6c). As the MSE is larger in the upper troposphere where air moves southward, this anomalous overturning circulation resulting in the net southward mean-flow energy transport anomalies (Fig. 7a). The anomalous southward energy transport is reinforced by the changes in the vertical MSE profile, as the MSE difference in the upper and lower troposphere is enhanced by the warming in the upper troposphere (Fig. 6d). In the high latitudes of the Southern Hemisphere, the eddy-induced energy transport anomalies outweigh those accomplished by mean-flow advection (Fig. 7a). When eddy-induced energy transport is approximated by the product of diffusivity and near-surface MSE gradient (Eq. 6), the changes in eddy-induced energy transport are primarily caused by the enhanced MSE gradient. Despite the reduced mass exchanging rate of eddy mixing, the efficiency of energy exchange increases due to the enhanced MSE gradient, resulting in an enhanced eddy-induced energy transport (Fig. 7b). The enhanced MSE gradient is associated with the enhanced gradient of moisture. A large portion of the

anomalous moisture gradient can be explained by the enhanced temperature gradient caused by the cloud-radiative changes, which alters saturated water vapor pressure (Fig. 7c).

3.2.3 Diffusive atmospheric energy transport

The EBM is additionally applied to understand the anomalous atmospheric energy transport in the extratropics from a diffusive perspective. Two forcing profiles (i.e. F_{EBM} in Eq. 8) are prescribed to the EBM to mimic the effects of cloud-radiative changes, with and without considering the anomalous surface flux induced by the cloud-radiative changes. The first forcing profile is the zonal-mean change in net cloud-radiative changes shown in Fig. 1, and the second forcing profile is obtained by subtracting the zonal-mean surface flux anomalies over the ocean surface from the first forcing profile, to capture the effects of anomalous oceanic circulation and heat content induced by the cloud-radiative changes. The comparison between the responses of the EBM to the two forcing profiles represents the modulation of oceanic processes to the anomalous temperature and energy transport. When the changes in ocean heat uptake are not considered, the EBM overestimates the warming in the regions northward of 30°S (Fig. 8a), which is accompanied by a more southward atmospheric energy transport compared to the GCM simulations (Fig. 8b). By including the anomalous surface flux, the distribution of surface temperature anomalies simulated by the EBM is closer to that in the GCM simulations, and the anomalous atmospheric energy transport around 40°S is captured. These results suggest important roles for changes in oceanic heat uptake and circulation (discussed in section 3.2.1) in modulating the responses of temperature and energy transport. In contrast to the extratropical atmospheric energy transport which exhibits a more diffusive nature (Armour et al. 2019) that

would tend to smoothen the warming pattern, our experimental setting highlights that the oceanic processes can play an active role in building the temperature anomalies.

In summary, various processes are in charge of accomplishing the increased poleward energy transport caused by cloud-radiative changes in the Southern Hemisphere. In the regions around 30°S, the poleward shift of the mid-latitude jet results in anomalous easterlies in the boundary layer, which drives an overturning cell in the ocean and result in an increased southward oceanic energy transport. Meanwhile, the anomalous eddy momentum flux divergence accompanying the shift of the mid-latitude jet drives an anomalous counter-clockwise overturning circulation in the atmosphere, and results in a net southward atmospheric energy transport which cooperates with the anomalous oceanic transport. In the high latitudes, the eddy-induced atmospheric energy transport increases because the MSE gradient becomes larger, which increases the efficiency of energy exchange.

The energetic and dynamical perspectives discussed in this study provide insights on how cloud feedback shapes the meridional structure of anomalous energy transport. Consistent with the arguments of Armour et al. (2019), the two perspectives have their advantages and limitations. The energetic perspective provides a constraint on energy transport by the energy fluxes at the TOA and surface, which may be easier to understand, without the need for detailed information about the dynamical response of the atmosphere and ocean, which is associated with complex physical processes that compensate each other (Donohoe et al. 2020). However, these dynamical responses can influence the TOA and surface energy fluxes by redistributing energy and modulating the surface temperature (Fig. 8). The two perspectives are complementary for understanding the forced response in the coupled climate system.

4. Discussion

The cloud feedback has been suggested to make a major contribution to the uncertainty in temperature and energy transport responses to CO₂ forcing in previous studies, which attributed the total warming responses to each feedback process based on their implied energy transport (Huang and Zhang 2014; Zelinka and Hartmann 2012) or based on EBM estimates (Armour et al. 2019; Bonan et al. 2018; Frierson and Hwang 2012; Hwang and Frierson 2010; Roe et al. 2015). The attributions in previous studies highlight the direct linkage between radiative feedbacks and energy transport; however, the interactions among the feedback processes are highly simplified and not well understood.

In the present study, the cloud-locking method allows one to investigate how the cloud-radiative changes influence other radiative feedbacks and dynamical processes, and how these adjustments modulate the anomalous energy transport induced by the cloud-radiative changes. By prescribing the cloud-radiative properties to the model, the cloud-locking method provides a causal linkage between the cloud-radiative changes and the responses of the climate system, which is unavailable in a standard warming experiment with interactive cloud-radiative properties. A few issues should be kept in mind when interpreting the results. First, the cloud-locking method cuts off the interactions between cloud-radiative effects and other cloud-related phenomena: for instance, a certain location can experience latent heat release while being cloud-free to the radiative fluxes, and vice versa. This decoupling may alter the baseline climate state, and may influence the responses of the climate system to external forcings through their non-linear interactions. Nevertheless, we have verified that the non-linear interactions between cloud-radiative decoupling and CO₂ forcing have much smaller impacts on energy transport and

surface temperature compared to the cloud impacts (not shown). Second, the cloud-radiative properties prescribed to the simulations are obtained from the interactive-cloud 4xCO₂ simulations. As a result, the cloud-radiative changes themselves would contain information about the changes in dynamical and thermodynamical factors under warming, such as temperature, humidity, and circulation. In particular, previous studies suggest the changes in oceanic processes and surface albedo can have significant impacts on the meridional structure of cloud feedback (Feldl et al. 2017; Trossman et al. 2016). While our experiments demonstrate the differences in the climate state in response to the different cloud states, one should keep in mind that the causality in an interactive-cloud simulation is more complex as the clouds and the cloud-controlling factors can influence each other. Moreover, as the cloud properties are obtained from time-evolving simulations (CTL and 4xCO₂), prescribing these cloud properties to another simulation would also introduce the signal associated with the exact temporal evolution in the CTL or 4xCO₂ simulations. In some previous studies, this additional information of temporal evolutions arising from the experimental setup is eliminated either by randomly rearranging the timeline of the cloud properties fed into the simulations (Radel et al. 2016), or by obtaining cloud properties from a single or several selected years with neutral-ENSO conditions in their control simulation (Benedict et al. 2020; Grise et al. 2019; Middlemas et al. 2019, 2020). Nevertheless, after time- and ensemble-mean is taken, the signals associated with the temporal evolutions of cloud properties are expected to be small, and are unlikely to significantly influence our conclusions.

Despite these caveats, this study reveals a few key mechanisms related to cloud-circulation interactions. We place them in the context of existing literature in the following paragraphs. In

particular, we highlight aspects that were not explicitly addressed in previous studies using the similar cloud-locking technique in a model with simplified ocean.

4.1 Hemispherically asymmetric responses of the cloud impacts on energy transport

It is demonstrated in our experiments that the cloud-induced heating is balanced locally in the Northern Hemisphere, while in the Southern Hemisphere the cloud-radiative changes are mostly balanced by the energy transport divergence. Accordingly, cloud-induced warming in the Northern Hemisphere mid-latitudes is stronger than in the Southern Hemisphere (blue line in Fig. 2a), despite similar magnitudes of cloud-radiative changes at 40°N and 60°S (Fig. 1).

The different behaviors in the Northern and Southern Hemispheres may arise from two aspects:

1. The intrinsic nature of the climate system, such as the land-sea distribution and climatological circulations in atmosphere and ocean: For instance, the land fraction is larger in the mid-to-high latitudes of the Northern Hemisphere. As the air is drier over land, LW cooling may be more efficient in the Northern Hemisphere because the greenhouse effect caused by water vapor is weaker. Meanwhile, the larger ocean basin in the Southern Hemisphere allows for a larger impact of wind torque and may make the energy transport more efficient. Moreover, the climatological ocean circulation around 60°S may suppress the local SST anomalies, where the surface temperature is tied to the upwelling seawater below (Armour et al. 2016), which makes local LW cooling inefficient and anomalous energy transport is required to balance the local cloud-radiative changes. The difference in the

efficiency of local LW cooling versus energy transport in the two hemispheres may lead to hemispherically asymmetric responses to imposed forcing. Using nine fully-coupled GCMs, Kang et al. (2019) showed that the changes in local temperature are larger in the Northern Hemisphere than in the Southern Hemisphere in response to an idealized forcing of reduced insolation in the extra-tropics. This is consistent with the hypothesis here that the hemispherically asymmetric responses may result from the intrinsic nature of the climate system.

2. The meridional structure of the cloud-radiative changes: In the Northern Hemisphere, the cloud-radiative changes exhibit more uniform heating meridionally, while in the Southern Hemisphere, the cloud-radiative changes exhibit a strong gradient of TOA imbalance. The larger latitudinal gradient in the Southern Hemisphere may be able to drive stronger atmospheric and oceanic circulation anomalies and therefore a larger anomalous energy transport. Further experiments would be needed to verify whether the structure of cloud-radiative changes is an essential cause for the hemispherically asymmetric responses in energy transport.

4.2 Inter-model spread in anomalous energy transport and temperature under global warming

Based on the estimations of implied energy transport, previous studies suggest that cloud-radiative changes dominate the uncertainty in energy transport under global warming (Huang and Zhang 2014; Zelinka and Hartmann 2012). Consistent with these results, our experiments show that the cloud-radiative changes under warming strongly modulate the anomalous energy transport, especially in the extra-tropics of the Southern Hemisphere. Accordingly, under the

scenario of quadrupled CO₂ concentration, CMIP5 models exhibit a high correlation between the anomalous energy transport at 40°S and the gradient of cloud-radiative changes in the Southern Hemisphere (Fig. 9a). These results suggest a dominant role of clouds on the inter-model spread of the anomalous energy transport under warming in the Southern Hemisphere. In contrast, in the Northern Hemisphere, the cloud impacts on the anomalous energy transport are small due to the local adjustment of anomalous TOA fluxes caused by increasing surface temperature. In previous studies diagnosing the implied energy transport of each feedback process (Eq. 2b), these cloud-induced changes in surface temperature and their associated TOA flux anomalies would have been attributed to temperature feedback, because they are indistinguishable from the CO₂-forced temperature responses. With the cloud-locking method that considers these cloud-induced changes in dynamical and thermodynamical fields, our finding implies that the impacts of uncertainty in cloud feedback on the anomalous energy transport in the Northern Hemisphere may be smaller than estimated by previous studies.

The contribution of cloud feedback to the uncertainty in the meridional structure of surface temperature has also been highlighted in previous studies using the EBM (Bonan et al., 2018; Roe et al., 2015). Through cloud-locking experiments, our results highlight the linkage between surface fluxes and cloud-radiative changes. A significant portion of the surface flux contribution in previous studies should be considered as being part of the oceanic adjustments to cloud-radiative changes, modulating the clouds' influence on surface temperature.

4.3 Cross-equatorial energy transport

Cross-equatorial energy transport plays an important role in determining the location of the tropical rain belt and in balancing the asymmetry of the energy budgets between the two hemispheres (Frierson and Hwang 2012; Kang et al. 2008, 2009). In response to a variety of hemispherically asymmetric forcings, recent studies suggest that the ocean tends to accomplish most of the change in cross-equatorial energy transport, compensating the anomalous cross-equatorial energy flux carried by the atmosphere and inhibiting the tropical rain belt shift (Chiang and Friedman 2012; Hawcroft et al. 2017; Kay et al. 2016; Mechoso et al. 2016; Tomas et al. 2016). This compensation between the anomalous atmospheric and oceanic energy transport is explained by the trade wind-coupling in previous studies (Green et al. 2019; Held 2001; Schneider 2017). Whenever there is an anomalous cross-equatorial overturning circulation in the atmosphere, driving anomalous atmospheric energy transport, the wind stress in the surface layer would drive an anomalous overturning cell in the ocean. To the extent that the gross stabilities in the atmosphere and ocean do not change significantly, the trade wind-coupling between the atmosphere and ocean would always result in an anomalous oceanic energy transport which has the same direction as the anomalous atmospheric transport, and therefore compensates the anomalous atmospheric energy transport. Changes in the Atlantic Meridional Overturning Circulation have also been suggested to contribute to the anomalous oceanic energy transport which compensates the atmospheric responses (Yu and Pritchard 2019).

In our experiments, the cloud-radiative changes under warming lead to the Northern Hemisphere receiving more energy (0.46 PW, Table 2) than the Southern Hemisphere, implying a southward cross-equator transport of 0.23 PW. Approximately 52% of the asymmetry in cloud-radiative changes is offset by the local adjustment term, leaving an actual change in southward cross-equator transport of 0.11 PW (Table 2, Fig. 3b). Consistent with previous studies, this cross-

equatorial energy transport is carried by the ocean. However, the mechanism driving the anomalous oceanic cross-equator energy transport differs from the trade wind-coupling mechanism proposed by previous studies (Green et al. 2019; Schneider et al. 2014). The anomalous oceanic cross-equatorial cell found in these previous works does not appear in our simulations (Fig. 5c). The anomalous oceanic energy transport across the equator in our simulations is mostly due to changes in mean-flow advection, and the contributions of changes in temperature and oceanic circulation are comparable (Fig. 10). The decompositions vary dramatically with latitude, suggesting that the oceanic circulation changes are more complex than those in idealized forcing experiments in previous studies.

The responses of atmospheric circulation and energy transport also differ from those in idealized forcing experiments in previous studies (Green et al. 2019; Schneider et al. 2014). Unlike the situations in these idealized studies, in which the atmosphere and ocean collaborate to transport energy from the hemisphere which receives more energy to the other one, in our experiments the anomalous atmospheric energy transport is northward at the equator, transporting energy into the hemisphere that receives more energy. The unexpected change in atmospheric cross-equatorial energy transport could be explained by the distribution of the cloud-radiative changes. The Northern Hemisphere, as a whole, receives more energy than the Southern Hemisphere, while in the tropics the situation is reversed (Table 2), which suggests that the direction of atmospheric cross-equator energy transport may be determined by the asymmetry of cloud-radiative changes within the tropics, rather than the whole globe. This is consistent with the results in Xiang et al. (2018), suggesting that tropical forcing is more efficient in driving atmospheric cross-equatorial energy transport than extra-tropical forcing.

5. Summary and Outlook

The impacts of cloud-radiative changes on meridional energy transport are investigated under a prescribed-cloud framework. In addition to confirming the significant influence of cloud-radiative changes on meridional energy transport suggested by previous studies using theoretical estimations, our study highlights a few radiative and dynamical adjustments that are important for understanding the influence of clouds on energy transport and temperature distributions. As summarized in Fig. 11, various processes are found to be important in shaping the latitudinal distribution of the anomalous meridional energy transport in response to the cloud-radiative changes:

- i.* The anomalous LW radiation resulting from the changes in temperature acts to locally balance the anomalous cloud-induced heating, leading to a small anomalous meridional energy transport in the Northern Hemisphere mid-latitudes.
- ii.* The cloud-radiative changes cause a poleward shift of the mid-latitude westerlies. The resulting anomalous Ekman transport induces an anomalous overturning circulation in the ocean that accomplishes most of the anomalous energy transport around 30°S.
- iii.* In the high latitudes of the Southern Hemisphere, the cloud-radiative changes enhance the near-surface meridional temperature gradient, which increases the moisture gradient and leads to anomalous poleward transport of energy performed by atmospheric eddy mixing.

Understanding the interaction between radiative fluxes and circulations is important in predicting the pattern of surface temperature anomalies, which has important impacts on regional climate. This study highlights the role of energy transport in modulating the responses of surface temperature to the cloud-radiative changes: the surface temperature anomalies are small in the

Southern hemisphere due to the anomalous energy fluxes carried by the ocean (process *ii* listed above) and the atmosphere (process *iii*). The wind-driven oceanic process *ii* plays an active role in redistributing meridional radiative gradients, which modulates regional temperature anomalies and anomalous atmospheric energy transport. In contrast, there are larger surface temperature anomalies in the Northern Hemisphere (process *i*). The efficiency of these adjustments may depend on model physics and it remains to be seen how sensitive these findings are to the particular climate model employed. In addition to the wind-driven oceanic process *ii* demonstrated in this study, the evolution of deep ocean circulations may also play an important role in modulating the warming pattern on centennial timescales. How the deep ocean circulations respond to cloud-radiative changes could be explored in longer simulations. The cloud-radiative changes in high latitudes discussed in process *iii* may also vary when considering longer timescales.

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Appendix

A1. Radiative kernels

In this study, radiative kernels (Pendergrass et al. 2017) obtained from a set of off-line radiation calculations are used to adjust cloud-radiative effects (CRE), defined as the differences between all-sky (R^{TOA}) and clear-sky (CLR) TOA fluxes, to remove the masking effects of clouds and to decompose the non-cloud TOA fluxes (R^{clr}).

A1.1 Cloud-radiative changes

The changes in TOA radiative flux associated with clouds (i.e. ΔR^{cld}) are estimated by

$$\Delta R^{cld} = \Delta CRE - \sum_f \left(\frac{\delta R^{TOA} - \delta CLR}{\delta f} \right) \Delta f, \quad (Eq. A1a) \text{ (Soden et al., 2008)}$$

Where Δ represents the differences between an experiment and the CTL simulation, and δ represents the anomalies in the off-line radiation calculations provided by the radiative kernels. f represents the non-cloud radiative controlling factors, including temperature, water vapor, surface albedo, and GHG concentration. The differences between δR^{TOA} and δCLR represent the masking effects of clouds on the anomalous TOA fluxes in response to the perturbation of δf . After normalizing by δf and rescaling by the changes in f in our experiments (Δf), the differences between δR^{all} and δCLR are used to estimate the masking effects in our experiments, which are then subtracted from ΔCRE to estimate the cloud-radiative changes. For water vapor and GHG concentration, a logarithmic rescaling is used rather than a linear one.

760 *A1.2 Decomposition of non-cloud TOA fluxes*

761 The change in non-cloud TOA fluxes, ΔR^{clr} , can be further decomposed to the anomalous fluxes
762 associated with the changes in each non-cloud radiation controlling factors f and a residual term:

763
$$\Delta R^{clr} = \sum_f \Delta R_f + residual, \quad (Eq. A1b)$$

764 where $\Delta R_f = \frac{\delta R^{TOA}}{\delta f} \Delta f$.

765 This method assumes that the changes in TOA fluxes are linearly or logarithmically related to
766 the changes in f . It also assumes that the interactions among the impacts of non-cloud radiative
767 controlling factors on the TOA fluxes are negligible. The residual term arises from these two
768 assumptions, but it is negligible in our diagnostics.

769

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Tables

Table 1. List of the experiments.

Experiment name	CO ₂ level	Averaging period (years)	Cloud properties for radiation calculation	Number of ensemble members
CTL	1x	1-50	Interactive	1
4xCO ₂	4x	21-30	Interactive	3
F1C1	1x	21-50	Years 1-25 in CTL	1
F1C2	1x	21-30	Years 1-30 in 4xCO₂	3
F2C1	4x	21-30	Years 1-25 in CTL	3
F2C2	4x	21-30	Years 1-30 in 4xCO₂	3

Table 2. Regional summation of the cloud-radiative changes and the cross-equatorial energy transport. Positive values represent downward TOA fluxes (cloud-radiative changes) or northward energy transport.

	90S-30S	30S-Eq	Eq-30N	30N-90N
Regionally-integrated cloud-radiative changes	-0.06 PW	0.32 PW	0.27 PW	0.45 PW
	0.26 PW		0.72 PW	
Implied cross-equator energy transport	-0.23 PW			
Cross-equator energy transport	Total: -0.11 PW			
	Atm:	0.02 PW	Ocn:	-0.13 PW

Figures

Fig. 1. Cloud-radiative changes at TOA in response to quadrupled CO₂ concentration in interactive-cloud simulations (4xCO₂-CTL, *solid lines*), and those corresponding to cloud impacts (ΔR^{cld}_{cloud} , *dashed lines*). The cloud-radiative changes are estimated by radiative kernels (see **Appendix A1** for details).

Fig. 2. (a) Changes in zonal-mean surface temperature (TS) in response to quadrupled CO₂ concentration in interactive-cloud (4xCO₂-CTL) and prescribed-cloud simulations (ΔTS_{total}), and the contributions of cloud impacts (ΔTS_{cloud}) and non-cloud impacts ($\Delta TS_{non-cloud}$). (b) Changes in energy transport in response to quadrupled CO₂ concentration in interactive-cloud (4xCO₂-CTL, *dashed lines*) and prescribed-cloud ($\Delta MET/AET/OET_{total}$, *solid lines*) simulations, and (c) the contributions of cloud impacts ($\Delta MET/AET/OET_{cloud}$, *blue lines*) and non-cloud impacts ($\Delta MET/AET/OET_{non-cloud}$, *red lines*). See Eqs. 3a-3c for the definitions of prescribed-cloud warming response, cloud impacts, and non-cloud impacts. (d) Anomalous

999 energy transport under warming (4xCO₂-CTL, years 21-30) in CMIP5 models (*solid lines*, with
1000 shadings representing the inter-model spread) and CESM1 (*dashed lines*).

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Fig. 3. (a) Zonal-mean structure of net cloud-radiative changes (ΔR^{cld}_{cloud} , green line, same as the black-dashed line in Fig. 1), and the cloud impacts on meridional energy transport convergence ($\Delta(\nabla \cdot MET)_{cloud}$, black-solid) and adjustment term (ΔAdj_{cloud} , black-dashed) (Eq. 9b). A positive value indicates the earth receives more energy in that latitude. (b) Implied northward energy transport (Eq. 10) due to cloud-radiative changes (ΔR^{cld}_{cloud}) and adjustment term (ΔAdj_{cloud}), and the actual changes in energy transport (ΔMET_{cloud}). The atmospheric and oceanic components of energy transport are shown by the lighter and darker gray lines, respectively. See Eq. 3a for the definition of cloud impacts.

Fig. 4. (a, b) Cloud impacts on adjustment term ($\Delta(Adj)_{cloud}$), and the components contributed by changes in non-cloud TOA fluxes ($\Delta(R^{clr})_{cloud}$) and ocean heat content ($\Delta(dH/dt)_{cloud}$). (c, d) Decomposition of the LW component of non-cloud TOA fluxes (ΔR^{clr}_{cloud}) (see **Appendix A1** for details). (a, c) Zonal-mean structures, and (b, d) implied northward energy transport. See Eq. 3a for the definition of cloud impacts.

Fig. 5. Cloud impacts on (a) oceanic energy transport ($\Delta(OET)_{cloud}$), and the contributions of mean-flow ($\bar{V}\bar{T}$), eddy ($\bar{V}'\bar{T}'$), sub-grid processes, and the interpolation errors. (b) Attribution of mean-flow transport ($\Delta(\bar{V}\bar{T})_{cloud}$) to changes in velocity ($\bar{T}\Delta\bar{V}$), temperature ($\bar{V}\Delta\bar{T}$), and nonlinear effects ($\Delta\bar{V}\Delta\bar{T}$) (Eq. 7b). (c) Meridional overturning circulation ($\Delta(MOC)_{cloud}$) of global ocean; black and blue contours represent positive and negative values of the climatology in the CTL simulation, respectively. The contour interval is 10Sv, with zero contours omitted. Shading represents the anomalies due to cloud-radiative changes. Positive values indicate clockwise circulation. (d) Oceanic mass transport in the near-surface layer, and the theoretical Ekman transport. See Eq. 3a for the definition of cloud impacts.

Fig. 6. Cloud impacts on (a) zonal-mean zonal wind, (b) eddy momentum flux convergence, (c) mass stream function of the atmosphere (positive values indicate clockwise circulation), and (d) air temperature. Black and blue contours represent positive and negative values of the climatology in the CTL simulation, respectively. The contour intervals are (a) 5m/s, (b) $2 \times 10^7 \text{ m}^2/\text{s}^2$, (c) $2 \times 10^{10} \text{ kg/s}$, and (d) 5K, with zero contours omitted. Shading represents the anomalies due to the cloud-radiative changes. Dots indicate significance at the 95% confidence level. See Eq. 3a for the definition of cloud impacts.

Fig. 7. Cloud impacts on (a) atmospheric energy transport ($\Delta(AET)_{cloud}$), along with the mean-flow ($\Delta(AET^*_{meanflow})_{cloud}$, Eq. 5c) and eddy-induced ($\Delta(AET_{eddy})_{cloud}$, Eq. 5d) components. (b) Attribution of the anomalous eddy-induced energy transport ($\Delta(AET_{eddy})_{cloud}$) to the changes in diffusivity (D^*), near-surface MSE gradient (MSE_y), and the non-linear term (Eq. 6b). (c) Attribution of the MSE gradient component (MSE_y) in (b) to the changes in dry static energy (DSE), saturated water vapor pressure (Q), and relative humidity (RH). See Eq. 3a for the definition of cloud impacts.

Fig. 8. Responses of (a) surface temperature and (b) atmospheric energy transport to the cloud-radiative changes (ΔR_{cloud}^{cl}), simulated by the GCM (black lines) and those predicted by the EBM, with (red) and without (green) considering the changes in oceanic energy transport and ocean heat content induced by the cloud-radiative changes.

Fig. 9. Anomalous northward energy transport at 40° versus the gradient of cloud-radiative changes under warming in CMIP5 models (*red dots*) and our CESM1 simulations (*black dots*). The gradient of cloud-radiative changes is defined as the difference between area-weighted mean cloud-radiative changes in the subtropics (10°-40°) and higher latitudes (40°-70°). The anomalies under warming are calculated as the averages of years 21-30 in abrupt4xCO2 simulations minus the corresponding 10-year averages in the piControl simulation. These periods are chosen for consistency with our diagnostics.

Fig. 10. Cloud impacts on the anomalous oceanic energy transport ($\Delta(OET)_{cloud}$) across 5°S/N and at the equator. The decompositions of total changes in oceanic energy transport ($\Delta(OET)_{cloud}$) are on the left-hand side of the vertical lines, while the right-hand side of the vertical lines shows the decompositions of mean-flow transport ($\Delta(\bar{V} \bar{T})_{cloud}$). The decompositions are identical to those in Figs. 5a, 5b. See Eq. 3a for the definition of cloud impacts.

Fig. 11. A schematic of the responses of thermodynamical and dynamical fields to the cloud-radiative changes, and the associated anomalous meridional energy transport in the atmosphere and the ocean.