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Key Points:
- Modified extratropical radiative feedbacks increase equilibrium climate sensitivity by 1.5 K, but hardly impact 21st century warming
- Energy input by extratropical shortwave cloud feedbacks is taken up by the ocean and moved to depth, delaying transient surface warming
- Extratropical cloud biases may not be as important to transient warming as biases in other regions due to extratropical ocean heat uptake

Supporting Information:
- Supporting Information S1

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Do Southern Ocean Cloud Feedbacks Matter for 21st Century Warming?

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Abstract Cloud phase improvements in a state-of-the-art climate model produce a large 1.5 K increase in equilibrium climate sensitivity (ECS, the surface warming in response to instantaneously doubled CO_2) via extratropical shortwave cloud feedbacks. Here we show that the same model improvements produce only a small surface warming increase in a realistic 21st century emissions scenario. The small 21st century warming increase is attributed to extratropical ocean heat uptake. Southern Ocean mean-state circulation takes up heat while a slowdown in North Atlantic circulation acts as a feedback to slow surface warming. Persistent heat uptake by extratropical oceans implies that extratropical cloud biases may not be as important to 21st century warming as biases in other regions. Observational constraints on cloud phase and shortwave radiation that produce a large ECS increase do not imply large changes in 21st century warming.

1. Introduction

Equilibrium climate sensitivity (ECS), the global surface warming resulting from an instantaneous doubling of CO_2, is widely used for climate model intercomparison (i.e., Andrews et al., 2012; Charney et al., 1979). The range in ECS predicted by state-of-the-art climate models has remained consistent, evolving from 1.5–4.5 K (Charney et al., 1979) to 2.1–4.7 K in the Coupled Model Intercomparison Project Phase 5 (CMIP5) ensemble (Flato et al., 2013). Many processes impact ECS. In particular, observationally constrained cloud phase improvements in the Community Earth System Model (CESM) aimed at rectifying a shortwave radiation bias common to many models (Trenberth & Fasullo, 2010) increase ECS by up to 1.5 K via more positive extratropical shortwave cloud feedback (Frey & Kay, 2017; Tan et al., 2016).

While ECS is important for model intercomparison and used in integrated assessment models, which estimate climate change impacts (Calel & Stainforth, 2017), the real-world significance of changes in ECS is unclear (Allen & Frame, 2007). Some have argued that intermodel ECS spread has a limited impact on climate change policy (Rogelj et al., 2014), while others argue that an ECS change of 0.5 K has important policy implications (Kaya et al., 2016). There is not a consistent relationship between ECS and transient warming quantified by transient climate response (TCR), an idealized transient warming metric (Cubasch et al., 2001), especially for high ECS values (Flato et al., 2013; Knutti et al., 2005; Meehl et al., 2007; Millar et al., 2015; Tsutsui, 2017). Many have argued that transient warming should not be inferred from ECS or vice versa (Allen et al., 2006; Andrews et al., 2015; Armour et al., 2013; Gregory et al., 2015; Gregory & Andrews, 2016; Rose et al., 2014; Senior & Mitchell, 2000; Wigley & Schlesinger, 1985; Zhou et al., 2016).

In addition, to idealized experiments used to estimate ECS and TCR, realistic 21st century forcing scenarios are applied to climate models to produce climate change projections (e.g., Representative Concentration Pathway (RCP) scenarios (Meinshausen et al., 2011)). Here we consider the relationship between ECS and 21st century warming. Among CMIP5 models (Forster et al., 2013) these quantities are positively correlated (R^2 = 0.72) though meaningful scatter exists (Figure S1 in the supporting information). Some models separated by more than 1 K in ECS predict very similar transient warming. In short, based on existing literature, it remains unclear whether the large ECS increase caused by improved model cloud phase over the Southern Ocean (Frey & Kay, 2017; Tan et al., 2016) is meaningful to projected transient 21st century warming.

Ocean heat uptake (OHU) is one of many factors influencing transient climate change (Hoffert et al., 1980; Manabe & Stouffer, 2007; Raper et al., 2002) in part by moving heat vertically (Gregory, 2000) to delay surface warming (Flato et al., 2013; Winton et al., 2010). OHU is not spatially uniform. Many models predict polar-amplified OHU as the climate warms (Marshall et al., 2015). This extratropical (defined as poleward of 30°

latitude) OHU is particularly effective at slowing surface warming because of its impact on radiative feedbacks (Armour et al., 2013; Bitz et al., 2012; Rose & Rayborn, 2016; Rose et al., 2014; Rugenstein, Caldeira, et al., 2016; Trossman et al., 2016; Winton et al., 2010).

The efficacy of extratropical OHU to delay surface warming has been demonstrated with a variety of idealized models (Armour et al., 2016; Rose et al., 2014; Rugenstein, Caldeira, et al., 2016; Trossman et al., 2016; Winton et al., 2010). Here we investigate the impact of OHU with a realistic model configuration including a full-depth dynamic ocean and a plausible forcing scenario. We show that model improvements that produce a large ECS increase produce only a small, though statistically significant, increase in 21st century warming under the RCP8.5 forcing scenario. Our results suggest that the effectiveness of OHU to slow transient warming in our experiment is linked to the collocation of increased positive shortwave cloud feedback, which drive the ECS increase, with areas of maximum OHU in the extratropics.

2. Model Description

We use the large ensemble (LE) version of CESM (Kay et al., 2015) based on CESM version 1 (Hurrell et al., 2013). Following Kay et al. (2016) and Frey and Kay (2017), we modify the Community Atmosphere Model version 5 to produce more liquid and less ice in extratropical shallow convective clouds. This improvement to cloud phase, along with tuning the threshold relative humidity for low cloud formation to maintain radiative equilibrium, reduces shortwave radiation biases over the Southern Ocean and tropics (Kay et al., 2016). Starting at the end of a 200 year run forced with constant 1850 conditions (which allows upper ocean temperatures to reach a new equilibrium (Kay et al., 2016)), we initialize a transient run with historical (1850–2005) and RCP8.5 (2006–2100) forcings (Riahi et al., 2011). Hereafter, we refer to this modified transient run as the "Experiment."

We compare our Experiment with the CESM LE (Kay et al., 2015). The LE simulates the years 1920–2100 multiple times with small atmospheric initial condition differences using the same historical and RCP8.5 forcing as our Experiment. We use 38 of the 40 LE members, omitting 31 and 33, which were postprocessed in a way that may impact radiative feedback calculations (Baker et al., 2016). The LE allows for separation of internal variability from forced response and provides an ideal data set to identify the impact of our cloud phase improvement and tuning on transient warming.

3. Results

3.1. Surface Warming

We first examine global, annual mean surface warming (Figure 1a). During the historical period (1920–2005) Experiment warming is within the LE range and comparable to observations. In the RCP8.5 period (2006–2100) the Experiment warms more than the LE, but only slightly above the LE range. By the late 21st century (2081–2100) the Experiment has warmed by 4.78 K above the 1850–1899 baseline. For comparison, LE warming over the same period ranges from 4.38 to 4.72 K with a mean of 4.50 K. The 0.3 K difference in warming between the Experiment and LE mean is statistically significant (99% confidence level) but small compared with the 1.5 K ECS increase (Frey & Kay, 2017).

The Experiment spatial pattern of warming (Figure 1b) follows well-established patterns in response to greenhouse forcing: notably polar amplification and North Atlantic cooling. Comparing warming between the Experiment and LE (Figure 1c) reveals interesting extratropical patterns. In the North Atlantic, the Experiment warms less than the LE. In the southern extratropics, there is increased Experiment warming from roughly 30 to 50°S and less warming further poleward.

3.2. Radiative Feedback Analysis

To understand the drivers behind the small 21st century warming increase (Figure 1a) and the geographic differences in warming (Figure 1c) between the Experiment and LE, we analyze radiative feedbacks (Soden & Held, 2006). We find that the Experiment has more positive shortwave cloud feedback than the LE and this increase is not fully compensated by other feedbacks. In the global mean (Figure 2a), both the Experiment and LE exhibit positive shortwave feedback as a result of positive cloud and surface albedo feedbacks. In contrast, both the Experiment and LE exhibit negative longwave feedback dominated by a negative Planck feedback partly compensated by positive water vapor and cloud feedbacks. The difference in shortwave

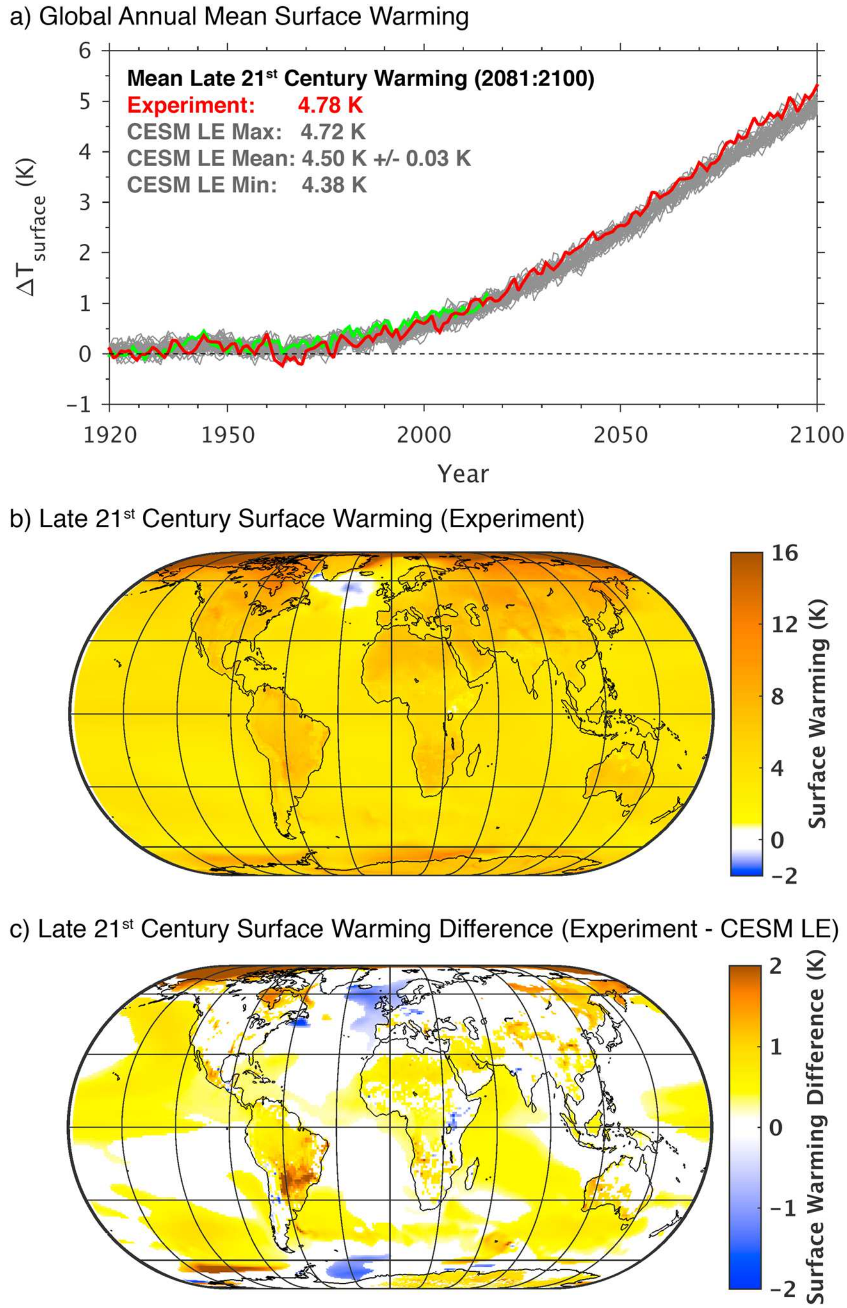

Figure 1. Surface warming above preindustrial (1850–1899): (a) global annual mean surface warming: Experiment (red), CESM LE (gray), and observations (green) (Hansen et al., 2010). (b) Late 21st century (2081–2100) annual mean surface warming (Experiment). (c) Late 21st century annual mean surface warming difference (Experiment-CESM LE mean). Text in Figure 1a shows mean warming over the late 21st century with a 99% confidence interval calculated about the LE mean using the *T* distribution. Differences colored in Figure 1c are statistically significant (99% confidence) (Wilks, 2016). Differences not statistically significant are white.

feedback between the Experiment and LE is more than double the difference in longwave and dominated by more positive cloud feedback (Figure 2b).

The increase in shortwave cloud feedback occurs entirely in the extratropics, primarily over the Southern Ocean (Figure 2c). In the tropics, the Experiment shortwave cloud feedback is within the LE range. In the extratropics, the Experiment shortwave cloud feedback is either at the top of (Northern Hemisphere) or well above (Southern Hemisphere) the LE range. This pattern is caused by the cloud improvements in the

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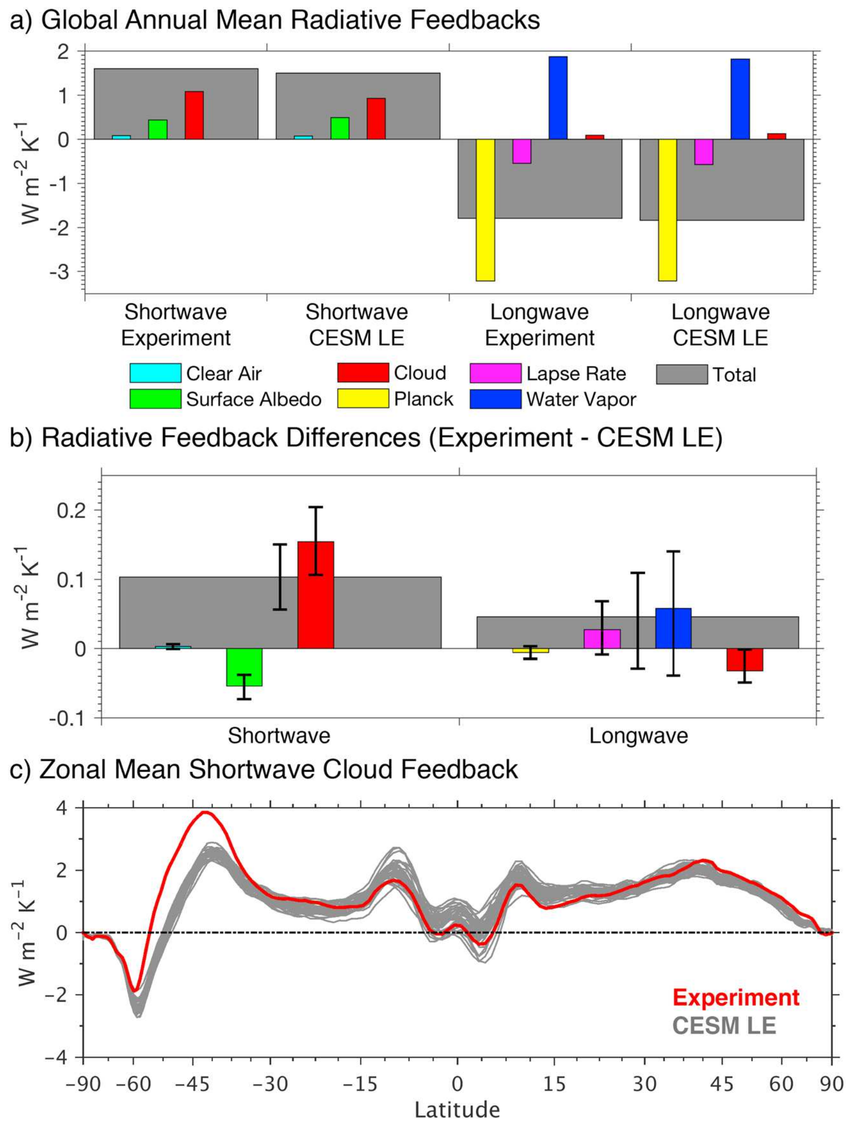

Figure 2. 21st century radiative feedbacks (2091–2100 minus 1996–2005): (a) global annual mean radiative feedbacks normalized by global annual mean surface temperature change. (b) Radiative feedback differences (Experiment minus CESM LE mean); the error bars show the range among CESM LE members. (c) Zonal annual mean shortwave cloud feedback normalized by local surface temperature change: Experiment (red) and CESM LE members (gray). Longwave feedbacks estimated using radiative kernels (Pendergrass et al., 2017). Shortwave feedbacks estimated using the approximate partial radiative perturbation (APRP) method in order to separate cloud and surface feedback (Taylor et al., 2007).

Experiment, which impact the magnitude of shortwave cloud feedback due to phase changes with warming (Frey & Kay, 2017).

The shortwave cloud feedback difference between the Experiment and LE (Figure 2c) is similar to the pattern of difference shown in slab ocean model runs forced with doubled CO_2 used to estimate ECS in Frey and Kay (2017). Yet the increase in 21st century warming (Figure 1a) is much smaller than the ECS increase of 1.5 K in Frey and Kay (2017). Taken together, these results suggest that OHU mutes the surface warming caused by extratropical cloud feedback in the RCP8.5 run and causes the increase in transient warming between the Experiment and LE to be small.

3.3. Ocean Heat Uptake

In both the Experiment and LE, OHU occurs preferentially in the extratropics (Figures 3a and 3b). The Southern Ocean takes up heat over the entire RCP8.5 period maximizing near 60°S. Northern extratropical

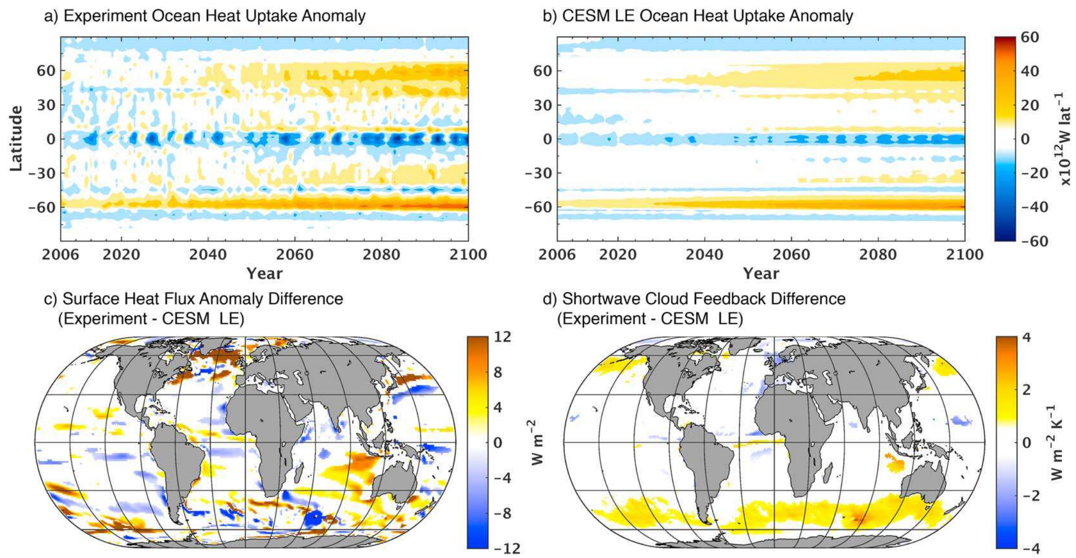

Figure 3. 21st century ocean heat uptake (OHU) anomalies compared to preindustrial (1850–1899): (a) annual mean zonally integrated OHU anomaly for the Experiment. (b) As in Figure 3a for the CESM LE mean. (c) Late 21st century (2081–2100) surface heat flux anomaly difference (Experiment minus CESM LE mean). (d) As in Figure 3c for shortwave cloud feedback. Zonally integrated ocean heat uptake calculated by multiplying surface heat flux by grid cell area and summing zonally. Positive values indicate heat into the ocean. Differences colored in Figures 3c and 3d are statistically significant (99% confidence) (Wilks, 2016). Differences not statistically significant are white.

OHU is not present at the beginning of the run but develops over time. By 2100 there is a broad OHU maximum near 60°N. By the late 21st century (2081–2100) the area-integrated OHU anomaly in the northern extratropics in the Experiment (LE) averages to 782 TW (635 TW) compared with 610 TW (516 TW) in the southern extratropics and only 49 TW (130 TW) in the tropics. In the southern extratropics, the difference in OHU between the Experiment and LE is not uniform but is distributed longitudinally over the Southern Ocean, the same region containing large shortwave cloud feedback differences (Figures 3c and 3d). Northern extratropical differences in OHU between the Experiment and LE occur mainly in the Atlantic and are not collocated with shortwave cloud feedback differences. We next analyze the role of ocean circulation in producing extratropical OHU.

3.4. Ocean Dynamics

In the Southern Ocean, positive heat content anomalies exist throughout the top 2,000 m (Figure 4a, colors). The heat content anomaly pattern is due to mean-state ocean circulation (Figure 4a, contours). The upwelling branch of the meridional overturning circulation (MOC) (Marshall & Speer, 2012) brings cool water to the surface where it gains heat from the atmosphere, producing maximum OHU near 60°S (Figure 3a). The circulation then brings this water equatorward before it sinks near 45°S, producing a heat content maximum and moving heat to depth (Figure 4a) (Armour et al., 2016). The circulation strength changes little over the RCP8.5 period (Figure 4e). Southern Ocean heat uptake is enhanced in the Experiment compared to the LE due to greater heating at the surface from a more positive shortwave cloud feedback (Figure 3d). The Experiment warms more than the LE both at the surface and at depth at southern midlatitudes (Figure 4c).

In the North Atlantic, the mean-state Atlantic MOC (AMOC) brings water northward before sinking at higher latitudes (Figure 4b, contours). As the RCP8.5 scenario progresses the AMOC slows down (Figure 4f) (Gregory et al., 2005; Jahn & Holland, 2013; Rugenstein, Sedlacek, et al., 2016; Stouffer & Manabe, 2003; Weaver et al., 2012) and decreased high-latitude sinking results in less cool water descending and less warm surface water transported northward (Banks & Gregory, 2006; Gregory, 2000; Xie & Vallis, 2012). This produces high-latitude near-surface cooling and warming at depth (Figure 4b). The AMOC slows down more in the Experiment than the LE by the late 21st century (Figure 4f). As a result, heat content at depth increases and near-surface heat content decreases in the extratropical North Atlantic in the Experiment compared to the LE (Figure 4d).

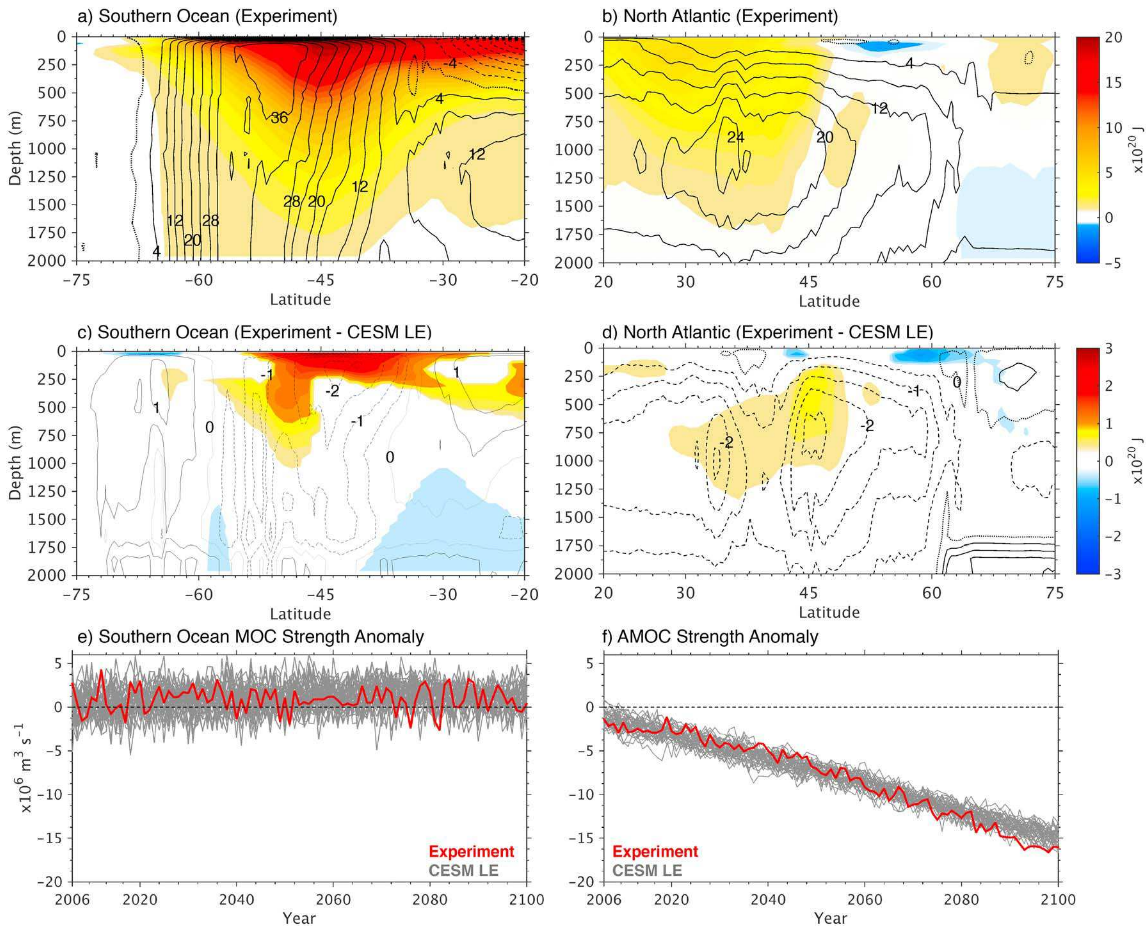

Figure 4. 21st century ocean heat content and meridional overturning circulation (MOC) streamfunction compared to preindustrial (1850–1899): (a) Experiment late 21st century (2081–2100) Southern Ocean zonally integrated ocean heat content anomaly (colors) and preindustrial MOC streamfunction (contours, 10^6 m^3 s^{-1}). (b) As in Figure 4a for the North Atlantic. (c) Late 21st century difference (Experiment minus CESM LE mean) in Southern Ocean heat content anomaly (colors) and MOC streamfunction anomaly (contours, 10^6 m^3 s^{-1}). (d) As in Figure 4c for the North Atlantic. (e) Southern Ocean and (f) North Atlantic MOC strength anomaly. Ocean heat content calculated by multiplying potential temperature by grid cell volume, a constant heat capacity for seawater (3,992 J kg^{-1} K^{-1}), and a constant seawater density (1,035 kg m^{-3}), as in Rugenstein et al. (2013). MOC strength defined as the maximum streamfunction between 30 and 60° south (Figure 4e) and north (Figure 4f). Positive (negative) streamfunction, shown with solid (dashed) contours, indicates clockwise (counterclockwise) circulation. Differences colored in Figures 4c and 4d are statistically significant (99% confidence) (Wilks, 2016). Differences not statistically significant are white.

4. Discussion

Compared to the LE, our Experiment takes up and stores more heat at depth in both the Southern Ocean and North Atlantic (Figures 3 and 4). Southern Ocean heat uptake does not require circulation changes (Armour et al., 2016; Marshall & Speer, 2012) and can be understood as a direct response acting to move heat input by cloud feedback to depth, slowing surface warming. Southern Ocean heat uptake is stronger in the Experiment than the LE not because of circulation differences (Figure 4e) but because more heat is available at the surface from more positive shortwave cloud feedback in the same region (Figure 3d). The efficacy of OHU to mute the impact of increased shortwave cloud feedback over the Southern Ocean suggests that model biases in this region may be less important to 21st century warming than biases in regions without strong OHU. We cannot be sure whether the increase in Southern Ocean heat uptake in our Experiment has reached a limit or if OHU would increase further if the shortwave cloud feedback were even more positive.

In contrast, North Atlantic OHU develops as warming progresses (Figures 3a and 3b) and is not collocated with differences in shortwave cloud feedback between the Experiment and LE (Figure 3d). Circulation changes necessary for North Atlantic OHU (Banks & Gregory, 2006; Gregory, 2000; Xie & Vallis, 2012) result from an AMOC slowdown (Figures 4d and 4f), a consequence of increased warming (Gregory et al., 2005; Jahn & Holland, 2013; Weaver et al., 2012). In a sense, North Atlantic OHU acts like a negative feedback (Trossman et al., 2016; Winton et al., 2013). The Experiment warms more than the LE over time (Figure 1c), which causes the AMOC to slow more in the Experiment (Figure 4f). As a consequence, North Atlantic OHU increases more in the Experiment than the LE by the late 21st century (Figure 3c), slowing surface warming.

Improved cloud phase in CESM produces a large (1.5 K) increase in ECS (Frey & Kay, 2017) but only a small increase in 21st century warming (Figure 1a). While this is not astonishing, it is also not obvious based on comparison between ECS and 21st century warming among CMIP5 models (Figure S1). One difference between our models and the CMIP5 ensemble is the cause of intermodel ECS spread. Among modern ensembles, ECS spread has been attributed to tropical and subtropical feedbacks (Sherwood et al., 2014; Tian, 2015; Vial et al., 2013; Webb et al., 2013) while extratropical feedbacks drive the ECS increase in our Experiment (Figure 2). This geographical difference in feedbacks may be important because of differences between tropical and extratropical OHU. Specifically, tropical OHU is important on short timescales (Clement et al., 1996; Held et al., 2010), but does not persist for long periods. Rose et al. (2014, Figure A1) show that in years 1–5 after quadrupling CO_2 OHU occurs in both the tropics and extratropics, but by years 96–105 tropical OHU is virtually zero while extratropical OHU persists (see also Marshall et al., 2015; Rugenstein, Caldeira, et al., 2016). Similarly, in transient runs OHU is much greater in the extratropics than tropics (Figure 3). When ECS increase is driven by feedbacks collocated with OHU in the southern extratropics, e.g., our Experiment, the ocean takes up heat and moves it to depth (Figure 4) slowing transient warming. In contrast, we hypothesize that tropical feedback contribute more to surface warming in transient runs because they are not collocated with persistent OHU. Therefore, ECS and transient warming may be more closely related for models which differ primarily in tropical shortwave feedbacks. Future work is required to determine whether collocation of feedbacks and OHU is necessary to slow transient warming and determine its relevance more generally.

Our work identifies interesting discrepancies in ECS estimates obtained with two common techniques. ECS is usually estimated rather than explicitly diagnosed due to the cost of running a fully coupled climate model to equilibrium. We estimate ECS for the Experiment and LE with mixed-layer "slab" ocean models (SOM) run to equilibrium with doubled CO_2. In contrast, CMIP5 ECS estimates use linear regression (Gregory et al., 2004) applied to fully coupled models with full-depth oceans. ECS is commonly estimated with both SOM (Meehl et al., 2007) and linear regression (Andrews et al., 2012; Flato et al., 2013; Gregory et al., 2004), and both techniques have been shown to produce reasonable estimates (Danabasoglu & Gent, 2009; Jonko et al., 2013; Li et al., 2013). Nevertheless, both methods have limitations. SOMs do not account for ocean circulation changes and produce different spatial patterns of feedback and warming compared with fully coupled models (Boer & Yu, 2003; Jonko et al., 2013; Williams et al., 2008). Linear regression uses simulations that have not reached equilibrium and thus lack the Southern Ocean warming our SOM simulations achieve (Armour et al., 2013; Frey & Kay, 2017). As a result, linear regression likely underestimates true ECS (Andrews et al., 2015; Gregory et al., 2004; Gregory & Andrews, 2016; Knutti & Rugenstein, 2015; Knutti et al., 2017). Despite these limitations, the impact of estimation method is thought to be minor compared to intermodel ECS spread (Flato et al., 2013).

Our Experiment is one case where these two methods produce inconsistent estimates. Using linear regression (Gregory et al., 2004) to estimate ECS for the Experiment and LE (Frey & Kay, 2017) reduces the ECS difference between the two versions of the model compared to SOM ECS estimates (Figure S1). We believe that our SOM ECS estimate is closer to the true ECS of our Experiment than ECS estimated with linear regression. This is in part because it compares favorably with Tan et al. (2016), who made modifications to CESM similar to our Experiment and found an ECS increase of 1.3 K compared to default CESM by running a fully coupled climate model with a full-depth ocean until the global top-of-atmosphere radiation budget was balanced with doubled CO_2. The large difference between SOM and linear regression ECS estimates for the Experiment suggests that the spatial differences in warming and feedback, which differentiate the Experiment from the LE, may impact the accuracy of ECS estimates (Andrews et al., 2015; Armour et al., 2013; Gregory et al., 2004; Murphy, 2010).

Our study is nominally limited by its design in that we completed only one RCP8.5 Experiment run to compare to 38 existing LE members. Thus, while we can show our Experiment falls outside of the internal variability-generated range of the LE (Figure 1a), we cannot show how a distribution of Experiment runs would compare. We consider this limitation to be second order because the difference in 21st century warming between the Experiment and LE is so much smaller than the ECS difference. Assuming that the internal variability in an Experiment ensemble would be of the order of the LE, our main conclusions would be unchanged. Notwithstanding this limitation and the discussion above, the main result of this study stands. Observational constraints on cloud phase that imply a large increase in ECS (Frey & Kay, 2017; Tan et al., 2016) do not imply a large increase in 21st century warming.

5. Conclusion

Cloud phase improvements in a climate model that decrease radiation biases produce a large (1.5 K) increase in ECS via extratropical cloud feedback (Frey & Kay, 2017). Despite this, 21st century warming under the RCP8.5 forcing scenario increases by a small, though statistically significant, 0.3 K compared to the default model, an increase just above the warming range due to internal variability (Figure 1a). The shortwave cloud feedbacks that drive increased ECS occur in the extratropics where the ocean is most effective at taking up heat (Figure 3). As a consequence, in the RCP8.5 scenario, the ocean takes up a portion of the heat from more positive extratropical shortwave cloud feedbacks (Figures 2 and 3) and moves it to depth (Figure 4), slowing surface warming compared to the default model. These processes are demonstrated with a state-of-the-art global climate model including a full-depth dynamic ocean and a realistic forcing scenario complementing previous work, which has identified the impact of extratropical OHU using idealized model configurations (i.e., Rose et al., 2014; Rugenstein, Caldeira, et al., 2016; Trossman et al., 2016). The ability of extratropical oceans to take up heat implies that extratropical cloud biases may not be as important to 21st century warming as biases in other regions. Observational constraints on cloud phase and shortwave radiation that produce a large ECS increase do not imply large changes in 21st century warming projections.

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