Cloud feedbacks on greenhouse warming in the superparameterized climate model SP-CCSM4

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Abstract.

Cloud feedbacks on greenhouse warming are studied with a superparameterized version of the Community Climate System Model (SP-CCSM4) that explicitly simulates cumulus convection. A 150 year simulation with an abrupt quadrupling of CO$_2$ is branched from a control run. The strongest cloud responses – a large decrease in mid-level cloud, more Arctic water and ice cloud, less low cloud over land, and a slight poleward shift of midlatitude storm track cloud – are also common responses in conventional global climate models and combine to produce moderate positive global cloud feedback and an implied climate sensitivity of 2.8 K. Two companion simulations, one with a uniform 4 K sea-surface temperature increase and one with quadrupled CO$_2$ but fixed SST, suggest that SP-CCSM4’s global-scale cloud changes are primarily mediated by the warming, rather than by rapid adjustments to increased CO$_2$. 
1. Introduction

Recent rounds of the Coupled Model Intercomparison Project, CMIP3 and CMIP5, have emphasized the continuing uncertainty in simulating cloud feedbacks on climate change with coupled atmosphere-ocean general circulation models (CGCMs) [Soden and Held, 2006; Soden and Vecchi, 2011; Zelinka et al., 2013; Vial et al., 2013]. This uncertainty is in part due to the need for complex parameterizations of subgrid variability of clouds, precipitation and radiation due to cloud-forming eddies, including both boundary-layer turbulence and cumulus convection.

Superparameterized (SP) climate models explicitly simulate the largest and most organized circulations within deep cumulus cloud systems using a cloud-resolving model (CRM) to represent the typical atmospheric state within each grid column of the global model. For computational efficiency, the CRM grid is still too coarse (often 4 km horizontal grid spacing and around 30 vertical levels through the atmospheric column) to simulate realistic boundary-layer eddies. Nevertheless, SP has been quite beneficial in simulating aspects of deep cumulus convection that have been challenging to parameterize, for instance the diurnal cycle of deep convection over land and the Madden-Julian Oscillation [Khairoutdinov et al., 2005]. A better simulation of deep convection affects the entire tropical general circulation, summertime continental clouds and precipitation, and parts of midlatitude cyclones. Hence, SP has the potential to improve the simulation of all types of cloud systems, although it cannot be expected to realistically simulate boundary-layer clouds without a careful treatment of much smaller-scale turbulent eddies, e.g. through subgrid turbulence closure parameterizations.
Even given its limitations for boundary-layer clouds, which are very important for global cloud feedback [Bony and Dufresne, 2005; Soden and Vecchi, 2011], SP provides an interesting comparison for cloud feedbacks simulated by conventionally-parameterized atmospheric general circulation models (AGCMs) and CGCMs. However, until recently its computational expense limited the length and types of simulations used for this purpose.

Wyant et al. [2006] compared the cloud distributions in a pair of 3.5 year simulations with a superparameterized adaptation of the Community Atmosphere Model, Version 3 (SP-CAM3), one with an additional uniform 2 K warming of sea-surface temperature (SST). They noted an increase in subtropical low cloud cover in the warm-SST case, which Wyant et al. [2009] attributed to enhanced boundary-layer radiative cooling. Wyant et al. [2012] analyzed cloud and circulation response of a slightly different version of SP-CAM3 to a CO₂ quadrupling with no SST warming, using a pair of 2-year simulations. They found a shift of cloud cover and mean ascent from ocean to land regions and a reduction in marine boundary-layer cloud-top heights, with little net global change in cloud cover.

In this paper, we analyze cloud feedbacks in a superparameterized adaptation of the Community Climate System Model Version 4 (SP-CCSM4), documented in detail by Stan and Xu [2014]. Stan and Xu [2014] and Arnold et al. [2014] have also recently discussed the climate change response of SP-CCSM4 to greenhouse gas perturbations, and Arnold et al. [2014] in particular noted a large increase in Arctic cloud cover in a 4×CO₂ climate, which they found was related in part to increased wintertime cumulus convection over an Arctic Ocean with much decreased sea ice. This paper will present a more comprehensive global view of SP-CCSM4 cloud feedbacks and compare them with the SP-CAM3 studies.
We follow the simulation protocol adopted in CMIP5 [Taylor et al., 2012] in support of Round 2 of the Cloud Feedbacks Model Intercomparison Project (CFMIP2) [Bony et al., 2011]. This includes (1) differencing a 150-year simulation of the coupled response to an abrupt CO\textsubscript{2} quadrupling with a coupled control simulation, (2) differencing two 35-year SP-CAM4 simulations to obtain the uncoupled response to a CO\textsubscript{2} quadrupling with no SST warming, and (3) similarly obtaining the uncoupled response to a uniform 4 K SST warming. The latter simulations are long enough to greatly reduce the cloud response uncertainty due to internal variability compared to the studies of Wyant et al. [2006] and Wyant et al. [2012]. We broadly compare the SP-CCSM4 cloud responses with the CMIP5 CGCMs to test whether the use of SP appears to have a large impact on cloud feedbacks.

Andrews et al. [2012] show that the top of atmosphere global-mean radiative response of a fixed-SST AGCM simulation to CO\textsubscript{2} quadrupling is well correlated across CMIP5 models with the effective radiative forcing derived as the \( y \)-intercept of a ‘Gregory’ plot of net radiative response vs. global surface air temperature difference \( \Delta T \) [Gregory and Webb, 2008]. Ringer et al. [2014] show that the temperature-mediated global cloud feedback of CMIP5 CGCMs is highly correlated across models to the cloud feedback inferred from a specified 4 K SST warming applied to their AGCMs. Gettelman et al. [2012] showed that the equilibrium climate sensitivity of slab-ocean versions of CAM4 and CAM5 were well-predicted by the response of specified-SST simulations with the same SST and CO\textsubscript{2} changes as the corresponding slab-ocean models (in this case the geographical pattern of SST change, not just the global-mean SST change, was specified based on the slab-ocean simulations). With these results in mind, we compare the SP-CAM4 AGCM simulations to the cloud and precipitation responses of the coupled simulation.
In Section 2, we give some brief details of the configuration of the simulations. In Section 3, we discuss the global and zonal-mean cloud and precipitation response in the coupled model to the abrupt CO$_2$ quadrupling. In Section 4 we analyze the resulting cloud feedbacks following Zelinka et al. [2012a], and relate our results to possible cloud feedback mechanisms. Section 5 discusses the coupling of the geographical response of clouds and cloud feedbacks to circulation, precipitation and land surface changes. Section 6 compares our findings to the SP-CAM3 cloud feedback studies, and Section 7 gives conclusions.

2. Simulations performed and terminology

SP-CCSM4 is documented by Stan and Xu [2014]. For the simulations presented here, its atmospheric component, the SP-CAM4, is run at 2 by 2.5 degree resolution with a finite-volume dynamical core, and the ocean component has 1 degree resolution. The SAM CRM used in each grid column of this model is oriented east-west, has 32 horizontal grid points with a grid spacing of 4 km, and has 30 vertical levels.

Table 1 lists the simulations performed. Nomenclature follows CMIP5 conventions, but for the coupled simulations there are minor deviations from the CMIP5 protocol. The 150 year SP-CCSM4 control and abrupt CO$_2$ quadrupling (abrupt4xCO2) simulations are both initialized from restart files for CCSM4 for Jan. 1, 2006 from an 1850-2010 historical-forcing run; (hence these runs use 2006 CO$_2$ = 367 ppm rather than a pre-industrial value as a baseline and have a short spin-up period in which they adjust to the superparameterized atmosphere).

The 35-year fixed-SST control simulation, amip, is forced with the seasonal cycle of the 1949-2001 mean SST and sea ice dataset described by Hurrell et al. [2008], with the same CO$_2$ concentration as the coupled control simulation. We classify it as AMIP-like
because it uses observed SSTs, but unlike a true AMIP simulation it does not include their interannual variability. The specified-SST perturbation simulations are \textit{amip4CO2} (CO$_2$ quadrupling only) and \textit{amip4K} (4 K SST increase only). Following the CMIP5 protocol, the same sea-ice dataset is used in these simulations as in the fixed-SST control. Unless otherwise indicated, coupled results are means over years 101-150 years, and fixed-SST results are means over years 2-35.

We use the following terminology for ‘experiments’, i.e. differences between simulation pairs:

\[ 4x = \text{abrupt4xCO2} - \text{Control} \]

\[ 4xP0 = \text{amip4CO2} - \text{amip} \quad (\text{CO}_2 \text{ quadrupling with fixed SST}) \]

\[ P4 = \text{amip4K} - \text{amip} \quad (4 \text{ K SST increase without CO}_2 \text{ change}) \]

\[ 4xLC = 4xP0 + 0.85P4 \quad \text{(specified-SST linear combination matching } \Delta T^{4x}) \]

Here, $\Delta T^{4x}$ = 4.5 K is the 100-150 year 4x global mean surface air temperature increase, and $\Delta T^{4xP0} = 0.5$ K and $\Delta T^{P4} = 4.6$ K are the 2-35 year 4xP0 and P4 global mean surface air temperature increases. The coefficient 0.85 is chosen such that

\[ \Delta T^{4x} = \Delta T^{4xP0} + 0.85\Delta T^{P4} \]  

(1) 

to match the 4x temperature and CO$_2$ increases, such that 4xLC fixed-SST responses in different fields can be meaningfully compared with 4x coupled responses.

Fig. 1 shows maps of the temperature response for these four cases. The 4xP0 ($\Delta$CO$_2$-only) warming is due both to small surface air temperature increases over the oceans (a reduced sea-air temperature difference) and larger greenhouse-induced surface air temperature increases over the continents (e.g. Wyant et al. [2012]).
The 4xLC pattern has the same global mean $\Delta \mathbf{T}$ as the coupled 4x case by construction. It also has a broadly similar spatial pattern of warming as 4x, including stronger warming over the continents. However, 4xLC has more muted warming in the polar regions because it uses fixed sea ice. The most striking difference of the coupled 4x SST response from the uniform-SST idealization of 4xLC is the SST cooling response in 4x in the subpolar northeast Atlantic ocean, which is presumably driven by feedbacks with oceanic deep convection. The spatial patterns of the 4x SST increase can be expected to also manifest in its cloud and precipitation-related responses.

3. Global- and zonal-mean cloud and precipitation responses in SP-CCSM4 simulation

Fig. 2 shows 150-year time series of selected global annual means from the SP-CCSM4 control and abrupt CO$_2$ quadrupling simulations. Fig. 2a shows surface air temperature, and Fig. 2b shows net top-of-atmosphere radiation $R$. Much of the difference between the two simulations in these quantities (the 4x response) develops in the first five years, although the difference is still increasing after 150 years. There is also a rapid adjustment of sea-ice cover (Fig. 2c) and total cloud cover (Fig. 2d) within the first year. The sea-ice is reduced more than 60% in abrupt4xCO2 after 100 years.

Fig. 3 shows a Gregory plot of the 4x difference of annual global mean TOA radiation components. All components show reasonably linear behavior with $\Delta \mathbf{T}$. A linear fit of the TOA net incoming radiation difference $\Delta R$ vs. $\Delta \mathbf{T}$ implies an equilibrium $\Delta \mathbf{T}$ at quadrupled CO$_2$ of 5.6 K, i. e. an equilibrium climate sensitivity to CO$_2$ doubling of 2.8 K. The positive slope $dSW_{clr}/dT$ reflects positive clear-sky snow/ice albedo feedback. The slopes $dLWCRE/dT$ and $dSWCRE/dT$ are both close to zero, but these should
not be interpreted as cloud feedbacks because of water vapor and surface albedo masking effects [Soden et al., 2004]. In Section 4, we will show the global shortwave and longwave cloud feedbacks are both slightly positive, 0.3 and 0.2 W m\(^{-2}\) K\(^{-1}\), respectively. The \(y\)-intercept of the SWCRE fit line is slightly positive, consistent with the slight reduction of total cloud cover during the initial rapid adjustment to the CO\(_2\) increase. The \(y\)-intercept of the LWCRE fit line is mainly due to CO\(_2\) masking.

The larger symbols in Fig. 3 show corresponding results from the specified-SST experiments 4xP0 (\(\Delta\)CO\(_2\)-only, magenta outline) and 4xLC (additional uniform SST increase, green outline). For the cloud radiative effects and clear-sky long wave radiation, these are consistent with the 4x coupled experiment. The clear-sky shortwave radiation and net radiation are less sensitive to \(\Delta T\) than in 4x because the sea-ice extent is fixed.

### 3.1. Zonal-mean cloud and precipitation responses

Fig. 4 shows selected zonal mean 4x differences averaged over years 100-150 (the period shown by the red bar in Fig. 2a.); panels (b) and (c) instead show the Control and abrupt4xCO2 precipitation and total cloud cover, so that their differences can be compared with the control climatology. Fig. 4a shows that the surface temperature difference is nearly uniformly 4 K equatorward of 40\(^\circ\) latitude in both hemisphere, with more than two-fold polar amplification in both hemispheres after 100 years. Precipitation is enhanced in the equatorial belt and the high latitudes (Fig. 4b), and total cloud cover shifts slightly poleward in the mid-latitudes of both hemispheres (Fig. 4c). These behaviors are both qualitatively and quantitatively within the range of CMIP5 CGCMs [Collins et al., 2013].

Fig. 4d-f shows the high (< 400 hPa), mid-level (400-700 hPa) and low (> 700 hPa) cloud responses. As with precipitation, there is a large increase in high cloud in high
latitudes, with a slight decrease in the subtropics and a slight increase near the equator. Low cloud also increases in polar ocean regions at the expense of the mid-latitudes, likely in connection with enhanced surface-driven convection over open water, as suggested by Arnold et al. [2014]. Mid-level cloud decreases at almost all latitudes, except right over the polar caps.

3.2. Latitude-binned Gregory plots of cloud types

Fig. 5 shows Gregory plots of the responses of high, middle and low cloud, as well as ice water path (IWP) and liquid water path (LWP) to the abrupt CO\textsubscript{2} increase. Tropical (\(<30^\circ\text{N/S}\)), midlatitude (30–60° N/S) and high-latitude (\(>60^\circ\text{N/S}\)) annual-mean responses are plotted for each simulated year. As in Fig. 3, the magenta and green symbols show corresponding 4xP0 and 4xLC specified-SST results.

All of the plots show an approximately linear behavior of cloud properties with surface temperature difference \(\Delta T\) with a \(y\) intercept that is close to zero. Following Gregory and Webb [2008], this suggests that these large-scale measures of SP-CCSM4’s cloud response are mainly temperature-mediated; rapid adjustment plays only a minor role. Except for low cloud and LWP, the 4xP0 and 4xLC results are also approximately on the same fit lines, supporting this interpretation and suggesting that regional SST differences are not crucial to these responses. The consistency between the specified-SST and coupled cloud responses on these very large space scales is consistent with the findings of Andrews et al. [2012] and Ringer et al. [2014].

A Gregory plot of precipitation (Fig. 5d), on the other hand, shows negative intercepts in the tropics and midlatitudes, corresponding to the expected decrease in global precipitation that accompanies a abrupt CO\textsubscript{2} increase (e. g. Bala et al. [2010]). It is
interesting that this is not accompanied by a larger corresponding rapid adjustment in cloud properties.

Figs. 5a-d corroborate the zonal-mean plots of cloud and precipitation response in Fig. 4, showing increases in polar high cloud and precipitation, decreases in mid-level cloud in the tropics and especially mid-latitudes, and low cloud increases in polar regions and decreases in the tropics. The ice water path (Fig. 5e) increases in polar regions in association with the high cloud increase, while the liquid water path (Fig. 5f) increases in polar regions in association with the low cloud increase. Shifts in the freezing level must also contribute to the LWP/IWP partitioning, but do not dominate the impacts of cloud fraction changes.

The 4xLC LWP in Fig. 5f is far above the coupled result for the midlatitude and polar regions; we do not have a good explanation for this.

4. Diagnosis of cloud feedbacks in SP-CCSM

We use the method of Zelinka et al. [2012a] to diagnose cloud feedbacks in SP-CCSM4. This method uses International Satellite Cloud Climatology Program (ISCCP) simulator diagnostics in combination with a radiative kernel to diagnose the TOA longwave and shortwave radiative feedback of cloud changes partitioned by location, cloud-top height and optical depth. Dr. Mark Zelinka kindly provided software to perform this analysis. We do not break the feedback into fast-adjustment and temperature-mediated components, since for SP-CCMS4 we have already shown the latter component dominates the global-scale cloud response. Also, as the ISCCP simulator is not run during polar night, the annual average of longwave cloud feedback at these latitudes is computed only from
months when the sun is above the horizon. Zelinka et al. [2012a, App. B, sec. d] suggest that the errors induced by this approximation are not significant.

Fig. 6 shows the geographical patterns of the net, longwave and shortwave cloud feedbacks. There is a positive global longwave cloud feedback of $0.30 \text{ W m}^{-2} \text{ K}^{-1}$ and a positive shortwave cloud feedback of $0.19 \text{ W m}^{-2} \text{ K}^{-1}$. These values are quite similar to the CMIP5 multimodel mean cloud feedbacks (without rapid adjustments separately broken out) in Fig. 3 of Zelinka et al. [2013]. For reference, Fig. 7 shows the easier-to-compute longwave and shortwave cloud radiative effect (here normalized by $\Delta T$), as well as the change in 500 hPa vertical pressure velocity $\omega_{500}$, which we use as a measure of circulation changes. As expected, the CRE changes and cloud feedbacks are highly spatially correlated for both longwave and shortwave, with a spatially-variable offset of their difference that reflects cloud masking effects. Globally, the longwave and shortwave feedbacks are $0.65$ and $0.16 \text{ W m}^{-2} \text{ K}^{-1}$ larger than the respective global CREs.

The longwave feedback is spatially correlated with the $\omega_{500}$ change, since high-topped clouds are favored in regions of mean ascent. It is rather insensitive to land-ocean boundaries. The shortwave feedback, on the other hand, is strongly positive over all low-latitude land masses except Australia (where there is very little low cloud to start with); we will relate this to low cloud reductions over land in SP-CCSM4 in a warmer climate. It also tends to spatially anticorrelate with the longwave feedback due to changes in middle and high-topped cloud cover.

Fig. 6 can be compared with CMIP3 multimodel mean cloud feedbacks deduced using a different technique by Soden and Vecchi [2011] (their Fig. 1). There is a reassuring similarity in the broad geographical pattern of feedbacks, e. g. the positive longwave
feedbacks in the wetter parts of the tropics, as well as in high latitudes, as well as the large positive shortwave feedbacks over the low-latitude continents outside Australia, though there are many regional differences. However, examination of Fig. 1 of Soden and Vecchi [2011] suggests that the SP-CCSM4 net and shortwave cloud feedbacks averaged over ocean areas are somewhat smaller over ocean compared to the CMIP3 composite, and that in the CMIP3 composites there is no clear difference between the land-mean vs. ocean-mean cloud feedbacks.

Fig. 8 shows the global, land-only, and ocean-only annual average 4x cloud cover changes binned by cloud top pressure and optical depth. The large global increase of high-top cloud in the uppermost cloud top pressure bin is due partly to the rise of the tropopause in a warmer climate [Hansen et al., 1984; Hartmann and Larson, 2002], and partly to the previously-noted increase in high cloud at high latitudes. Consistent with our earlier findings, there is a decrease in mid-top cloud of all optical depths. Low cloud cover of all optical depths also decreases over land, but low cloud of intermediate optical depths increases over the oceans and in the global mean. For mid-top and high-top cloud, the 4x cloud changes are fairly similar between land and ocean, though as previously noted they do depend on latitude.

The global, land-only, and ocean-only longwave, shortwave and net cloud feedbacks due to these cloud changes are also given in Fig. 8. Since the main difference in ocean vs. land cloud changes is the large decrease in low cloud over land, the longwave cloud feedback is similar over ocean and land, but the shortwave cloud feedback is strongly positive over land, and in fact weakly negative over ocean.
We also partitioned the global net feedback into contributions from altitude, cloud fraction and optical depth changes, following Zelinka et al. [2012b]; of these the altitude increase of high clouds explains 0.30 W m\(^{-2}\) K\(^{-1}\) of the global net cloud feedback, all in the longwave, while the cloud fraction and optical depth changes and a residual term all make smaller positive contributions to the feedback.

### 4.1. Relation of global cloud feedbacks to cumulus mixing efficiency metrics

Sherwood et al. [2014] introduced two metrics, \(S\) and \(D\), of lower to mid-tropospheric cumulus mixing efficiency in the current climate that empirically are correlated with climate sensitivity and hence global cloud feedbacks in the CMIP3/CMIP5 models. These metrics can be construed as measures of how well GCM-simulated shallow cumulus convection (reaching no higher than the freezing level) can keep up with deep cumulus convection in the regions in which deep convection is most prevalent. They might also be relevant to how aggressively shallow cumulus convection vertically mixes in the subsiding branch of the Hadley circulation, which supports most of the boundary layer cloud and is responsible for much of the intermodel spread in cloud feedbacks. SP-CCSM4 does not use a cumulus parameterization, relying instead on an explicit, albeit under-resolved simulation of cumulus convection. Hence, it is of interest (a) how well SP-CCSM4 agrees with the observed value of these metrics, and (b) whether its climate sensitivity lies on the CMIP-based regression lines between climate sensitivity and each of \(S\) and \(D\).

Following the methods section of Sherwood et al. [2014], at each month and tropical (30N-30S) ocean grid point, we computed \(s = 0.5(\delta R - \delta T/9)\), where \(\delta\) denotes a difference between monthly-mean values at 700 hPa and 850 hPa, \(R\) is relative humidity, and \(T\) is temperature in K. This is a measure of low-tropospheric temperature and moisture
stratification, which Sherwood et al. [2014] argue is regulated by convective mixing efficiency. For each month, we select the tropical ocean grid points in regions of 500 hPa mean ascent $-\omega_{500} > 0$, spatially average $s$ over the quartile of these grid points with strongest mean ascent (which mainly lie in the west Pacific warm pool). We then average the resulting quantity across months to get a statistic $S$, which we find is 0.40 for SP-CCSM4. Based on Sherwood et al. [2014]’s Fig. 5a, this is within the range 0.35-0.5 suggested by radiosonde observations and reanalyses. SP-CCSM4’s climate sensitivity of 2.8 K is also within the fairly broad range (2-4.5 K) of climate sensitivities of GCMs whose $S$ is within the range 0.35-0.45. For this metric, SP-CCSM4 is consistent with observations and behaves similarly to conventional GCMs.

The second metric $D$ is a ratio of shallow vs. deep overturning in regions of mean low-level ascent over the ITCZ sectors (160W-30E) of the tropical oceans, where ‘bottom-heavy’ mean vertical motion profiles are common. In the discussion below, ‘800 hPa’ will be used as a proxy for an average of 850 and 700 hPa data, and ‘500 hPa’ as a proxy for an average of 400, 500, and 600 hPa data, to simplify the description of Sherwood et al. [2014]’s ratio. For each month, those grid columns within this region with 800 hPa mean ascent and net lateral outflow of mass from the lower troposphere (500-800 hPa) are selected. The shallow overturning is the area integral of this lateral mass flux. Then, those grid points within the ITCZ region with 500 hPa monthly-mean ascent are selected, and the deep overturning is defined as the area integral of this upward mass flux through the 500 hPa level. The monthly ratio of shallow to deep overturning is averaged across the annual cycle to obtain $D$. 
For SP-CCSM4, following this prescription we obtain $D = 0.32$, which is at the upper end of the CMIP5 model range and somewhat less than two reanalysis-based estimates of 0.38 and 0.45. We note that most of the contribution to $D$ comes from ‘marginal’ precipitation regions outside the core of the ITCZ, and is sensitive to the extent of these regions. We also computed $D$ using the specified-SST control run of SP-CAM4, and found a significantly smaller value $D = 0.21$ that is below the median of CMIP5 models shown in Sherwood et al. [2014]’s Fig. 5b. To the extent that $D$ is an indicator of simulated convective mixing, it is not obvious why it should be substantially different in the coupled and specified-SST simulations, raising concerns about the robustness of $D$ for this purpose.

The climate sensitivity of SP-CCSM4 is on the low end of the CMIP5 models with $S + D$ values similar to the 0.72 of SP-CCSM4, but it would be comparable to CMIP5 models with if we used $D = 0.21$ to calculate $S + D = 0.61$ instead.

### 4.2. Subtropical cloud feedbacks poorly predicted by CGILS column tests

The CGILS intercomparison [Zhang et al., 2013; Blossey et al., 2013] aimed to understand mechanisms of subtropical low cloud feedback using single column experiments. Realistic steady ‘control’ (CTL) large-scale advective forcings and boundary conditions corresponding to mean summertime conditions in three northeast Pacific locations, S6, S11 and S12, were applied to obtain simulated equilibrium cloud-topped boundary layer for those locations, which in reality support shallow cumulus cloud regimes, cumulus rising into stratocumulus, and well-mixed stratocumulus, respectively. Then the column model was brought into equilibrium with a perturbed forcing (P2S), including a 2 K SST and column temperature increase and a 10% decrease in subsidence, to mimic effects of climate change. The hope was that differences ($\Delta P2S$) of low cloud between these two column
experiments would approximately correspond to those simulated by the full GCMs. This
did not prove to be the case for conventional climate models [Zhang et al., 2013], because
the equilibrium responses of single column models do not respond smoothly to small per-
turbations in the forcings. In fact, the single-column model responses were dominated
by grid-locking effects in which the cloud response of a single column model sensitively
depended on whether or not the capping inversion shifted up or down a grid level due to
the forcing perturbation.

In the superparameterized SP-CAM, each grid column of the AGCM is simulated by
an entire CRM, so one might imagine the mean state of the corresponding atmospheric
column model would have more freedom to evolve smoothly in response to a small forcing
perturbation. Thus, we used a 2D version of the SAM CRM with a grid, time step,
and physical parameterizations configured as closely as possible to that in SP-CAM4, to
perform the CGILS tests. We call this configuration SP-SAM.

Table 2 summarizes our findings. For brevity, SWCRE is the sole metric used for
cloud response; following Blossey et al. [2013] this is calculated for the CGILS cases from
simulated 8-10 day means. We find that both the mean state and perturbation response
of SP-SAM are significantly different from those of the full-resolution three-dimensional
version of SAM presented by Blossey et al. [2013]. We also compare the SP-SAM CGILS
results with the summertime (June-July-August or JJA) climatology of the SP-CAM4
amip and amip4K runs at the same three locations. This makes a better comparison than
the coupled results, for two reasons. Firstly, the specified-SST control run enforces the
same observed SST climatology as used in the CGILS boundary conditions, Secondly, the
P4 experiment, like the CGILS P2S perturbation, only includes a warming perturbation.
and not a CO₂ perturbation, while allowing mean subsidence to respond to the SST warming. Because the P4 SST perturbation is twice as strong as that used for CGILS, in Table 2 we halve the P4 cloud-radiative response for comparison with the CGILS ∆P2S response.

We find that SP-SAM has much more SWCRE in the control case, and much more positive cloud feedbacks, than the full SP-CAM4 at all three locations. These results bear further study. Inspection of the time series and vertical structure of the control and perturbed cloud responses (not shown) suggests that grid locking remains an important issue for SP-SAM, like conventional single-column models, because the vertical grid spacing is coarse and the strong simulated inversions remain flat across the CRM grid. For instance, the very large SP-SAM warming response in the S11 case is due to the inversion moving up a layer and causing partial cloud dissipation, while in the S12 case, the inversion also moves up a layer and this induces decoupling of the boundary layer, also contributing to stratocumulus thinning.

For now, we conclude that (a) as currently configured, the CGILS cases are not a good guide to subtropical cloud feedback for SP-CCSM4, and (b) that a much higher vertical and horizontal grid resolution (or at least a more sophisticated sub grid cloud and turbulence parameterization) in the embedded CRMs of SP-CCSM4 could substantially change its cloud feedbacks. Lastly, we note that the thermodynamic and radiative mechanisms of positive marine subtropical low cloud feedback deduced from the CGILS LES results [Bretherton et al., 2013], respectively due to overall warming of the lower atmosphere and to the increased longwave emissivity of the overlying free troposphere, are not apparent in the SP-CCSM4 results. Instead, SP-CCSM4 shows very little marine subtropical low
cloud change in the face of increased greenhouse gases and a warmer atmospheric column. A key issue for future study is to resolve whether the weak subtropical low cloud changes in SP-CCSM4 are due to inadequate CRM resolution or to other confounding consequences of greenhouse warming.

5. Changes in the geographical distribution of cloud and precipitation

As seen above, the regional response of clouds to a changing climate is tightly coupled to circulation changes [Bony et al., 2004], as well as differential changes between land and ocean. These relationships affect zonal means, but the full regional response, which is relevant to climate impacts, is best visualized with selected geographical maps, presented in this section.

5.1. Mid-level cloud response

Fig. 9a shows that in the control run, mid-level cloud is much more prevalent at latitudes poleward of 45° in both hemispheres. Fig. 9b shows the coupled response of mid-level cloud to the abrupt CO$_2$ increase. As suggested by the zonal-mean behavior, mid-level cloud decreases nearly everywhere except the polar caps and the central equatorial Pacific, with typical relative reductions of 10-20%. The spatial modulations in mid-level cloud change correlate well with $-\omega_{500}$ (Fig. 7a). There is an offset, reflected in regions such as the North Pacific storm track with increased upward motion but mid-level cloud decreases, suggesting that the ratio of mid-level cloud to upward mass flux in storm systems may tend to decrease in SP-CCSM4 in a warmer climate.

In contrast to SP-CCSM4, Zelinka et al. [2013] showed that the CMIP5 multimodel mean exhibits a fast CO$_2$-forced reduction of midlevel cloud, but only a weak overall
further temperature-mediated decrease. We speculate that this may reflect differences in
the microphysical character of mid-level cloud changes in SP-CCSM4 compared to some
CMIP5 GCMs.

5.2. Low cloud changes

Fig. 10a shows the spatial distribution of low cloud in the control simulation. The
persistent low cloud over the midlatitude oceans is well captured, but the subtropical
stratocumulus regions less well so [Stan and Xu, 2014]. Fig. 10b shows the 4x low
cloud response. There is a low cloud decrease over the midlatitude oceans and there
are increases in the central subtropical oceans, corresponding to a westward shift of the
subtropical stratocumulus regions.

The most striking feature is the low cloud decrease over most land areas. Fig. 10c
shows the 4x near-surface relative humidity response. There is a strong spatial correlation
between near-surface relative humidity and low cloud cover changes over both land and
ocean. As noted by Joshi et al. [2008] and others, the near-surface relative humidity
decreases sharply over land as the climate warms because the land surface becomes more
moisture-limited. This inhibits low cloud development since it requires a deeper layer of
boundary layer turbulent mixing before condensation will occur.

5.3. Coupled vs. specified-SST cloud-radiative and low cloud responses

In this and the following subsection, we will compare the spatial structure of the coupled
(4x) response of selected quantities to that due to the specified-SST perturbations, 4xP0
(ΔCO₂-only) and P4 (ΔSST-only), and their temperature-matched linear combination
4xLC. We have already learned from the Gregory plots (Fig. 5) that for averages of
most cloud quantities over the entire tropical, midlatitude, or polar latitude belts, the
\( \Delta \text{CO}_2 \)-only response is fairly small, and the 4xLC combination matches the coupled run
fairly well. On these nearly global scales, then, the SP-CCSM4 coupled cloud response
is dominated by temperature-mediated changes which are not sensitive to the spatial
patterns of SST increase.

We now test whether the same is true for cloud-related responses at smaller regional
scales. Since tropical circulations are sensitive to SST gradients, and SST does not increase
in the 4x run exactly uniformly, like it does in the 4xLC analogue, some regional differences
between the 4x and 4xLC patterns of cloud-related response are to be expected. But are
these large enough to make the 4x vs. 4xLC comparison meaningless?

We start by considering the \textit{cloud radiative response}, which we define as the difference
in TOA radiative flux (defined positive downward) between a pair of simulations due only
to the changes in the three-dimensional cloud distribution, computed from the ISCCP-
binned cloud histograms following Zelinka et al. [2012a]. For the 4x (coupled) case, this
is the cloud feedback multiplied by the temperature change (recall that in this paper, we
are including fast cloud adjustment within the cloud feedback). For all simulation pairs,
it differs from the CRE response due to cloud masking effects, and gives a truer measure
of the importance of cloud changes alone to the TOA radiation budget.

Fig. 11 compares the spatial structure of the 4x (coupled) net (longwave plus shortwave)
cloud radiative response with the \( \Delta \text{CO}_2 \)-only, P4 and combined (4xLC) responses. The
\( \Delta \text{CO}_2 \)-only response is generally weak, consistent with the Gregory plots, but with positive
net radiative response over many land areas leading to a positive global-mean \( \Delta \text{CO}_2 \)-only
response of 0.8 W m\(^{-2}\). The temperature-mediated cloud radiative response is stronger,
with generally positive values at lower latitudes and slightly negative values at higher latitudes, especially over oceans. The 4xLC net cloud radiative response has a fairly similar zonal-mean structure to the coupled 4x case, and both are positive over nearly all land regions. 4x and 4xLC also both tend to show negative cloud radiative response in the subtropical stratocumulus to cumulus transition regions. The negative equatorial cloud radiative response in the coupled simulation is not apparent in the 4xLC analogue. This presumably results from more upward motion and cloud due to a slight equatorial Pacific enhancement of the SST warming in 4x that is robustly seen in CMIP5 models and has been attributed to weakening of the Walker circulation [Chadwick et al., 2013].

Fig. 12 shows the analogous plots for low cloud response. All panels look strikingly similar to the net cloud radiative response, except with reversed sign. This echoes the well-known correlation between net cloud radiative effect and low cloud amount in the current climate [Klein and Hartmann, 1993]. Thus, low cloud changes explain almost all the cloud-radiative changes in the previous figure; in particular the low cloud decrease over land in both 4x and 4xLC explains the positive net cloud radiative response over land in these simulations. Both the $\Delta CO_2$-driven and $\Delta SST$-driven low cloud responses contribute to these low cloud decreases over land.

Using a subset of CMIP5 models, Zelinka et al. [2013] found low cloud reduction over land that in the multimode mean is weaker than SP-CCSM4 and develops mainly in the initial $\Delta CO_2$-induced rapid adjustment. Since low cloud reduction over land is important to the positive low cloud feedbacks in SP-CCSM4, and since shallow convective cloud over land is very difficult to parameterize well in conventional GCMs, this issue bears further
study, perhaps using a superparameterized model with a higher-resolution CRM that can better simulate shallow convection over land.

5.4. Coupled vs. specified-SST responses in $\omega_{500}$ and precipitation

Fig. 13 compares the 4x $\omega_{500}$ response to the specified-SST cases. The CO$_2$-only response is weak, though with systematic ascent ($\omega_{500} < 0$) over land and subsidence over ocean [Wyant et al., 2012]. The P4 response is stronger in most regions, and dominates the 4xLC response. The spatial pattern of $\omega_{500}$ response in 4xLC is fairly similar to the 4x (coupled) case. An important common feature between the two cases is a reduction of the tropical overturning circulation (a positive $\omega_{500}$ response in the rainy regions and a negative $\omega_{500}$ response in the subsidence regions) [Chadwick et al., 2013]. For the 4x case, there is also more ascent along the equatorial Pacific, also like most CMIP models, for reasons noted in the previous subsection. The 4xLC case also exhibits this response, but to a lesser extent.

The 4x and specified-SST precipitation responses, shown in Fig. 14, have spatial structure that is tightly anticorrelated with $\omega_{500}$. This is particularly evident in the $\Delta$CO$_2$-only response. In the other cases, the atmospheric warming also introduces a thermodynamic component into the precipitation response [Bony et al., 2013], due to Clausius-Clapeyron-driven increase of the precipitation produced by a given upward mass flux. Precipitation increases in the equatorial Pacific in 4x (the coupled case) are dynamically driven [Bony et al., 2013].

The 4xP0 and P4 precipitation responses in Figs. 14b, d can be approximately compared with the CMIP5 multimodel mean responses in Fig. 2d and 2e of Bony et al. [2013], keeping in mind that the latter were derived from Gregory regressions rather than
fixed-SST simulations. As with other fields, the ΔCO₂-only response is weaker in SP-
CAM4 than the CMIP5 multimodel mean, although it has a broadly similar geographical
structure. SP-CAM4 is not unique in this respect. Bony et al. [2013] note that the in-
termodel variability between CMIP5 GCMs in the ΔCO₂-only response accounts for a
much larger fraction of the intermodel variability in 4x precipitation changes that does
the temperature-mediated response.

6. Comparison with SP-CAM3 cloud responses

SP-CAM4 is not dramatically different from SP-CAM3. The CRM is nearly unchanged
between the two models. Their main difference is the use of a finite-volume dynamical
core in the global host model of SP-CAM4, vs. a semi-Lagrangian tracer advection scheme
in SP-CAM3. Hence, one might anticipate that the cloud response of the two models to
a climate perturbation would be similar.

Comparison of our results with those published for SP-CAM3 suggests otherwise. As
discussed in the introduction, Wyant et al. [2006] examined the cloud response of SP-
CAM3 to a uniform 2 K SST increase (P2), using two years of output after spinup. They
found an increase in marine low cloud cover and negative shortwave and net CRE changes
at most latitudes in the warmer simulation. Our SP-CAM4 results suggest low cloud
decreases in some latitude belts, especially the tropics, both for the coupled (4x) case and
for the P4 case analogous to Wyant et al. [2006]’s study. Similarly, Wyant et al. [2012]
noted slight low cloud increases over land in their SP-CAM3 fixed-SST CO₂-quadrupling
study, using 3.5 years of control and perturbed simulation, while our 4xP0 experiment
shows cloud decreases over land.
In this section, we test whether these differences are meaningful or whether they can be explained purely by interannual variability of the cloud response. The SP-CAM3 papers attempted to address this issue using the interannual differences between their simulated years, but with only 2-3 years of model output this is not very reliable. Our 35-year long SP-CAM4 specified-SST experiments are long enough to accurately estimate a year-to-year standard deviation in any cloud or radiation statistic, which we further assume is not substantially different in SP-CAM3.

6.1. Response to a uniform SST increase

Fig. 15 compares the shortwave, longwave, and net CRE from the SP-CAM3 P2 climate perturbation (adapted from Fig. 1 of Wyant et al. [2006]) and the SP-CAM4 P4 climate perturbation. The latter is plotted with a doubled vertical scale for a fair comparison. In both plots, the grey shading is a ±2σ uncertainty range of net CRE over the averaging period. Here, the estimated σ = N^{-1/2}σ_i, where σ_i is the corresponding interannual standard deviation based on years 2-35 of the SP-CAM4 P4 experiment, and N is the number of simulated years (3 for SP-CAM3, 34 for SP-CAM4). The uncertainty range is much larger for SP-CAM3 because fewer years are represented in the average.

The longwave CRE is similar between the models. Throughout most of the tropics, the shortwave and net CRE are negative in SP-CAM3, while they are near zero or positive in SP-CAM4. These differences in net CRE somewhat exceed the ±2σ uncertainty range, and probably reflect real effects from the changes in model version.
6.2. Response to a CO$_2$ quadrupling with fixed SST

The spatial patterns of longwave and shortwave cloud adjustment for SP-CAM3 in response to CO$_2$ quadrupling with fixed SST (Fig. 2 of Wyant et al. [2012]) are qualitatively similar to the SP-CAM4 4xP0 results in most locations. However, there are quantitative differences in mean tropical (30S-30N) cloud and radiation changes, given in Table 3. Even the tropical cloud climatologies of the two SP-CAM control runs are different. SP-CAM4 has larger fractional cover than SP-CAM3 in all vertical layers, low, medium and high. Table 3 includes $\pm 2\sigma$ uncertainty ranges for the time-mean 4xP0 cloud and radiation responses for both models, calculated as in the previous section, but now with $N = 2$ for SP-CAM3.

The mid-level and high cloud changes from both SP-CAM4 and SP-CAM3 are positive over land and negative over ocean, and their $\pm 2\sigma$ uncertainty ranges nearly overlap. For low cloud over land, SP-CAM4 shows decreases over land, while SP-CAM3 suggests highly uncertain increases. The change in net CRE is correspondingly less negative over land in SP-CAM4 than in SP-CAM3 (to translate net CRE change into a tropical-mean cloud radiative response, add a CO$_2$-masking correction of 1.1 W m$^{-2}$ for either model [Wyant et al., 2012]).

Overall, SP-CAM4 appears to have stronger temperature-mediated low cloud reduction, somewhat more positive cloud feedbacks, and a stronger reduction in low cloud over land than SP-CAM3.

7. Conclusions

We have analyzed the cloud response of the super-parameterized coupled climate model SP-CCSM4 to climate perturbations following CFMIP protocols, using an abrupt 4xCO$_2$
simulation and related specified-SST perturbation experiments. We use a combination of Gregory regression plots and ISCCP-simulator cloud feedback analysis [Zelinka et al., 2012a].

In most respects, SP-CCSM4’s cloud adjustment and feedbacks lie within the typical range of conventionally parameterized CMIP5 GCMs. Since SP-CCSM4 does not employ a cumulus parameterization, and cumulus parameterization details can substantially affect global cloud feedbacks [Zhao, 2014; Sherwood et al., 2014], this is reassuring (with the caveat that the embedded cloud-resolving models in SP-CCSM4 do not resolve shallow cumuli or boundary-layer turbulence that dominate cloud feedback uncertainty).

Compared to most CMIP5 GCMs, SP-CCSM4’s simulated clouds respond weakly to CO₂ changes alone. This is corroborated by fixed-SST simulations with quadrupled CO₂. As global surface temperature increases, SP-CCSM4 simulates strong reductions in mid-level cloud worldwide and strong increases in Arctic high and low cloud, as noted by Arnold et al. [2014]. Both of these are common in CMIP5 models. SP-CCSM4 also simulates strong reductions in boundary layer cloud over land, more so than most CMIP5 models. These reductions are highly spatially correlated to reduced near-surface relative humidity driven by reduced moisture availability of land in a warmer climate, and are the main driver of SP-CCSM4’s weakly positive global shortwave cloud feedbacks. SP-CCSM4 also shows the upward shift of high cloud seen in CMIP5 models and the associated positive longwave cloud feedback [Zelinka et al., 2012b]. Overall, SP-CCSM has a moderate positive total cloud feedback of 0.5 W m⁻² K⁻¹, consistent with its equilibrium climate sensitivity of 2.8 K.
As with the CMIP5 models [Ringer et al., 2014], the cloud response of SP-CCSM4 at global and near-global scales is also captured in atmosphere-only simulations with a specified uniform 4 K SST increase. These simulations even reproduce some regional features of the coupled cloud response, such as the low cloud reduction over the continents and slight marine low cloud increases downstream of the subtropical stratocumulus to cumulus transition regions.

As superparameterized models are developed with more skillful sub-grid turbulence parameterizations or better CRM resolutions, it will be important to see how this affects both the overall magnitude and spatial pattern of their cloud feedbacks. Based on the current study, even just the first 20 years of a coupled abrupt $4 \times CO_2$ simulation pair or 10 years of specified-SST simulation pairs will give an excellent guide to the large-scale cloud response of an SP-CCSM version to various combination of $CO_2$ and surface warming perturbations.

Acknowledgments. We acknowledge support from the NSF Science and Technology Center for Multi-Scale Modeling of Atmospheric Processes (CMMAP), led by David Randall and managed under the leadership of by Colorado State University under cooperative agreement No. ATM-0425247. The computations underlying this research used resources of the National Energy Research Scientific Computing Center, which is supported by the Office of Science of the U.S. Department of Energy under Contract No. DE-AC02-05CH11231, and the Extreme Science and Engineering Discovery Environment (XSEDE), which is supported by National Science Foundation grant number OCI-1053575. The model output data used to produce the results of this publication can be obtained upon
request to C. Stan. We would like to acknowledge Marat Khairoutdinov for his sustained contributions to the development of SAM and SP-CAM.

References


### Table 1. Simulations performed

<table>
<thead>
<tr>
<th>Case</th>
<th>Model</th>
<th>Description</th>
<th>Length</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>SP-CCSM4</td>
<td>Coupled control</td>
<td>150 yrs</td>
</tr>
<tr>
<td>abrupt4xCO2</td>
<td>SP-CCSM4</td>
<td>abrupt 4×CO₂</td>
<td>150 yrs</td>
</tr>
<tr>
<td>amip</td>
<td>SP-CAM4</td>
<td>Specified-SST control</td>
<td>35 yrs</td>
</tr>
<tr>
<td>amip4CO₂</td>
<td>SP-CAM4</td>
<td>Quadrupled CO₂, specified-SST</td>
<td>35 yrs</td>
</tr>
<tr>
<td>amip4K</td>
<td>SP-CAM4</td>
<td>Uniform 4 K SST increase</td>
<td>35 yrs</td>
</tr>
</tbody>
</table>

*a* All specified-SST simulations have the same sea-ice extent
Table 2. CGILS control and P2 SWCRE responses [W m\(^{-2}\)] for SP-SAM vs. high-resolution 3D SAM, SP-CAM4, and satellite observations

<table>
<thead>
<tr>
<th>Model(^a)</th>
<th>S6</th>
<th>S11</th>
<th>S12</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>CTL</td>
<td>ΔP2S</td>
<td>CTL</td>
</tr>
<tr>
<td>SP-SAM</td>
<td>-100</td>
<td>8</td>
<td>-210</td>
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<tr>
<td>Hi-res 3D SAM(^a)</td>
<td>-30</td>
<td>3</td>
<td>-145</td>
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<tr>
<td>SP-CAM4 (JJA)(^b)</td>
<td>-93</td>
<td>-1.4</td>
<td>-107</td>
</tr>
<tr>
<td>Observed (JJA)(^c)</td>
<td>-25</td>
<td>-83</td>
<td>-84</td>
</tr>
</tbody>
</table>

\(^a\) From Blossey et al. [2013].

\(^b\) From sampling of the SP-CAM4 P4 experiment at the CGILS locations. ΔP2S perturbation is estimated as half the P4 perturbation.

\(^c\) Based on CERES climatology from 2000–2013 (current climate only).
Table 3. SP-CAM3 vs. SP-CAM4 tropical-mean (30S-30N) cloud cover and net cloud radiative effect, and their sensitivities over land and ocean regions to CO$_2$ quadrupling with fixed SST

<table>
<thead>
<tr>
<th>Model</th>
<th>Case</th>
<th>Low cld [%]</th>
<th>Med cld [%]</th>
<th>High cld [%]</th>
<th>Net CRE. [W m$^{-2}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>SP-CAM3</td>
<td>CTL</td>
<td>27</td>
<td>7</td>
<td>17</td>
<td>-29</td>
</tr>
<tr>
<td></td>
<td>$\Delta_{\text{land}}$</td>
<td>0.6 ± 0.8</td>
<td>0.6 ± 0.5</td>
<td>2.0 ± 1.0</td>
<td>-1.0 ± 1.3$^b$</td>
</tr>
<tr>
<td></td>
<td>$\Delta_{\text{ocn}}$</td>
<td>-0.1 ± 0.3</td>
<td>-0.3 ± 0.2</td>
<td>-0.4 ± 0.4</td>
<td>-0.9 ± 0.5</td>
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<tr>
<td>SP-CAM4</td>
<td>amip</td>
<td>31</td>
<td>11</td>
<td>28</td>
<td>-38</td>
</tr>
<tr>
<td></td>
<td>$\Delta_{\text{land}}$</td>
<td>-0.9 ± 0.2</td>
<td>1.2 ± 0.1</td>
<td>1.9 ± 0.2</td>
<td>0.0 ± 0.4</td>
</tr>
<tr>
<td></td>
<td>$\Delta_{\text{ocn}}$</td>
<td>-0.3 ± 0.1</td>
<td>-0.6 ± 0.04</td>
<td>-0.8 ± 0.1</td>
<td>-0.8 ± 0.1</td>
</tr>
</tbody>
</table>

$^a$ From Table 1 of Wyant et al. [2012].

$^b$ Estimation of ±2σ uncertainty ranges for both models is described in text.
Figure 1. Surface air temperature increase relative to control simulations in 100-150 year mean of coupled (4x) and in 2-35 year mean of specified SST experiments (4xP0: fixed SST, 4xCO₂ only; P4: 4 K SST increase only). P4LC is a linear combination of the 4xP0 and P4 responses with the same global surface air temperature and CO₂ increase as 4x.
Figure 2. SP-CCSM4 global-mean ‘4x’ response to abrupt quadrupling of CO$_2$. 
Figure 3. Gregory plot of 4x global-mean TOA radiative flux changes. Symbols show corresponding results from specified-SST experiments 4xP0 (magenta) and 4xLC (green).
Figure 4. 4x zonal mean change in (a) temperature and (d-f) high, mid-level, and low cloud. (b) and (c) show the Control and abrupt4xCO2 mean precipitation and total cloud cover, respectively. Color bands in panel a show the low, middle and high-latitude averaging regions used in Fig. 5.
Figure 5. Gregory plot of tropical, mid-latitude and high-latitude changes in selected cloud-related quantities.
Figure 6. Maps of SP-CCSM4 net, longwave and shortwave cloud feedback deduced from 4x changes in ISCCP simulator cloud fraction.
Figure 7. Maps of SP-CCSM4 4x responses of $\omega_{500}$, longwave and shortwave cloud radiative effect.
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Figure 9. (a) Mid-level cloud in Control, and (b) 4x changes in mid-level cloud.
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Figure 12. Low cloud change: coupled vs. specified SST experiments.
Figure 13. 500 hPa omega change: coupled vs. specified SST experiments.
Figure 14. Precipitation change: coupled vs. specified SST experiments.
Figure 15. Zonal mean CRE change (a) for P2 climate perturbation using 3.5 year SP-CAM3 simulations, and (b) for P4 SP-CAM4 climate perturbation, with doubled vertical scale. The grey shading on both plots bounds the ±2σ uncertainty on the mean net CRE due to interannual variability estimated from the 34 years of the SP-CAM4 P4 experiment.