Understanding Subtropical Low Cloud Response to a Warmer Climate in a Superparameterized Climate Model.

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Abstract

The subtropical low cloud response to a climate with SST uniformly warmed by 2 K is analyzed in the SP-CAM superparameterized climate model, in which each grid column is replaced by a two-dimensional cloud-resolving model (CRM). Intriguingly, SP-CAM shows substantial low cloud increases over the subtropical oceans in the warmer climate. The paper aims to understand the mechanism for these increases, and to test their sensitivity to the coarse CRM resolution (4 km horizontal, 30 vertical levels). The approaches presented also apply to other global climate models or warming scenarios.

The subtropical low cloud increase is analyzed by sorting grid-column months of the climate model into composite cloud regimes using percentile ranges of lower tropospheric stability (LTS). LTS is observed to be well correlated to subtropical low cloud amount and boundary layer vertical structure. The low cloud increase is attributed to boundary-layer destabilization due to increased clear-sky radiative cooling in the warmer climate. This drives more shallow cumulus convection and a moister boundary layer, inducing cloud increases and further increasing the radiative cooling.

The SP-CAM resolution sensitivity is tested with a new CRM analogue to an SP-CAM composite cloud regime. The CRM is run to steady state using composite advective tendencies, winds, and sea-surface temperature from SP-CAM control and +2 K climates. A new ‘weak temperature gradient’ algorithm based on an idealized form of gravity wave adjustment is used to adjust vertical motion in the column to keep the simulated virtual temperature profile consistent with the corresponding SP-CAM composite profile. Humidity is also slowly relaxed toward the SP-CAM composite above the boundary
layer. With SP-CAM grid resolution, the CRM shows +2 K low cloud increases similar to SP-CAM. With fine grid resolution, the CRM-simulated low cloud fraction and its increase in a warmer climate are much smaller. Hence, the negative low cloud feedbacks in SP-CAM may be exaggerated by under-resolution of cloud-topped boundary layers.
1. Introduction

Clouds play a large role in the climate system. Conventional atmospheric general circulation models (GCMs) parameterize unresolved cloud processes. Uncertainties in cloud parameterizations are a major factor in the overall uncertainty in climate model projections. Differences between models in the representation of low clouds have been identified as major cause of model disparity in climate response (Bony and Dufresne 2005). We do not yet have an accepted physical theory for predicting the response of low clouds to a climate change that allows us to prefer one GCM over another.

In most GCMs, multiple interacting parameterizations for turbulence, cloud fraction and microphysics, radiation, etc. determine low cloud properties. Although processes such as turbulence and condensation interact on the scale of individual updrafts and downdrafts, GCMs are forced to represent their interaction via parameterizations communicating on the much larger scale of a GCM grid cell, a challenging problem with no unique solution. Although the solutions to this problem embodied in different GCMs can be tested against current climate, this has so far provided little constraint on the GCMs’ predictions of cloud feedbacks on future climate change.

In this paper, we explore the application of cloud resolving models (CRMs) to improve the simulation and physical understanding of boundary-layer cloud feedbacks on climate. CRMs can directly simulate turbulent cloud processes by using much finer spatial resolution than GCMs, but are typically run for shorter periods with much smaller domains due to their heavy computational requirements. The NICAM model (Tomita et al. 2005) is a global CRM. Because of its computational costs, only runs of seasonal
duration are practical. Another approach is superparameterization (Grabowski 2001; Khairoutdinov and Randall 2001) in which a GCM is run with each grid column replaced by a CRM that interacts with the large-scale fields. With current implementations of either approach, boundary-layer clouds are severely under-resolved by the CRM grid, so the simulated boundary-layer cloud feedbacks on climate are of questionable realism. However, it may soon be computationally practical to use much higher resolution in the embedded CRMs of a superparameterization.

The focus of this paper is low-latitude boundary layer cloud response to a climate perturbation in the SP-CAM superparameterization (Khairoutdinov et al. 2005). SP-CAM uses the two-dimensional version of the System for Atmospheric Modeling (SAM, Khairoutdinov and Randall 2003) CRM embedded in each column of version 3.0 of the Community Atmosphere Model (CAM) GCM (Collins et al. 2006). CAM is run with 30 vertical levels and 2.8° x 2.8° horizontal grid spacing. The CRM replaces the cloud and moist-physics parameterizations within CAM. It has 32 grid columns 4 km apart, aligned in the north-to-south direction, and 30 vertical levels, 28 of which are co-located with CAM model levels. The specific version of SP-CAM used here is further described in Wyant et al. (2006), hereafter referred to as W06. W06 examined the cloud response of SP-CAM to a global 2K sea-surface temperature (SST) increase. This perturbation induced a significant increase in low cloud cover over the oceans in both the tropics and extratropics, implying strong negative cloud feedbacks on climate change. We focus on the cloud increases over the low latitude oceans (30S-30N; hereafter LLO) because the low clouds there vary less from day to day than in the extratropics, so they can be
effectively analyzed using the monthly-mean fields for which we have SP-CAM model output.

Our two specific goals are (1) to using a cloud-regime binning approach to study the warming-induced LLO low cloud increase and its physical mechanism, and (2) to develop a CRM-based column analogue to SP-CAM that roughly matches its cloud climatology and cloud response to climate change in a typical trade-cumulus regime, with which we explore the sensitivity of the cloud response to the grid resolution of the embedded CRM. Our approaches to both goals generalize to conventionally parameterized GCMs, using the single-column version of the GCM to form the column analogue.

Section 2 gives an overview of the SP-CAM low cloud climatology and its response to an SST increase. In Section 3, these are related to lower tropospheric stability (LTS), which is an empirical correlate of boundary-layer cloud amount in the current climate. In Section 4 we gain insight into the low cloud changes by compositing the LLO into cloud regimes using LTS percentile ranges. In Section 5, we use this analysis to propose a physical mechanism for the SP-CAM low cloud response. In Section 6, we review previously proposed single-column analogues for studying LLO boundary-layer cloud responses to climate change and their findings. With this background, in Sect. 7 we develop a column analogue to SP-CAM, and in Sect. 8, we show that it can approximately simulate the mean cloud and +2K cloud response composited over an LTS percentile range that corresponds to a shallow trade-cumulus cloud regime. In Section 9, we use this single-column analogue to explore the sensitivity of the boundary-
layer cloud response to use of a much finer CRM grid resolution than currently used in SP-CAM. Section 10 presents our conclusions.

2. SP-CAM Climatology and Climate Sensitivity

We analyze the two SP-CAM simulations presented in W06. The 3.67-year ‘control’ simulation uses climatological SST, and the 5.25-year ‘+2K’ simulation is identical except that the SST is uniformly increased by 2K. Both control and +2K simulations use identical specified sea-ice climatologies. The simulations start on September 1 and the first 6 months of each simulation are discarded because of model spin-up. Monthly climatologies are created from the remainder of the simulations.

The global climate sensitivity, $\lambda$, can be calculated for this type of experiment based on the relation $\Delta T_s = \lambda G$, where $\Delta T_s$ is the global mean change in surface temperature and $G$ is the change in net outgoing radiation (Cess et al. 1989). W06 found $\lambda = 0.67$ K m$^{-2}$ W$^{-1}$, a low climate sensitivity when compare with other GCMs. This was attributed to increases in low cloud amount and liquid water path (LWP) at both low and high latitudes. These create a global mean change in net cloud forcing (NCF) of -1.77 W m$^{-2}$ when SST is raised 2 K (NCF is defined as the difference between net downward radiative flux and the clear-sky net downward radiative flux at the top of the atmosphere. Shortwave cloud forcing (SWCF), used below, is defined analogously).

Figure 1a shows the annual mean low cloud fraction for the control run, defined as the integrated cloud fraction between the surface and 700hPa.
Figure 1. Annual mean SP-CAM fields and observations. (a) SP-CAM low cloud fraction in the control simulation, (b) SP-CAM net cloud forcing (W m$^{-2}$), (c) ERBE 1986-1989 net cloud forcing (W m$^{-2}$), and (d) SP-CAM lower tropospheric stability (K).

Cloud fraction at a given level is defined as the fraction of horizontal grid points with liquid water content greater than 0.01 g kg$^{-1}$. The realism of the SP-CAM low cloud distribution compares favorably with other contemporary GCMs but does have some notable deficiencies. The high stratocumulus amounts in the eastern subtropical oceans
do not extend far enough westward. Much of the stratocumulus cloud near the coast reaches down to the surface, creating excessive fog-like stratocumulus.

The annual mean net cloud radiative forcing in SP-CAM is shown in Fig. 1b. The large negative NCF over high latitude oceans and subtropical stratocumulus regions are prominent, consistent with the large low-cloud fractions in these regions. Qualitatively the SP-CAM cloud forcing compares well with the annual mean ERBE net cloud forcing from the period 1986-1989 (Fig. 1c). However, the cloud forcing in the western oceans and the polar regions is too negative. Also, the ‘stratocumulus’ regions of strongly negative cloud forcing in the subtropics do not extend far enough westward away from the coasts.

On monthly timescales, low cloud amount has a strong space-time correlation with lower-tropospheric stability over the LLO (Klein and Hartmann 1993). LTS is conventionally defined as the difference between the potential temperature at 700 hPa and at 1000 hPa. For SP-CAM, we slightly modify this definition by replacing the temperature at 1000 hPa with the 2-m surface air temperature to minimize errors due to interpolation between model levels.

The annual mean SP-CAM LTS is plotted in Fig. 1d. The spatial correlation between annual mean LTS and low cloud fraction over the LLO is 0.35 in SP-CAM. The observed annual correlation of low cloud amount from visual surface observations (Hahn and Warren 2007) and LTS from ECMWF ERA-40 reanalysis (Uppala et al. 2005) is 0.48. The positive correlation between LTS and maritime low cloud in SP-CAM is smaller at
higher latitudes, as in observations.

Wood and Bretherton (2006) suggested a modification of LTS, the estimated inversion strength (EIS), that is a better predictor of mid-latitude maritime low clouds and which they speculated might be more applicable to prediction of climate-perturbation induced low cloud changes. We nevertheless use LTS in this paper because in SP-CAM and observations over the LLO, monthly-averaged LTS is somewhat better correlated with low clouds and NCF than is EIS. Furthermore, the statistical relationship between EIS and low cloud cover in SP-CAM shifts in the +2K climate, though to a lesser extent than with LTS. Hence, EIS does not prove to be a low cloud predictor that is invariant to climate perturbations, at least for SP-CAM.

3. LTS and Climate Sensitivity

The +2K low cloud increase in SP-CAM (Fig. 2a) is far from uniform, and induces a spatially anti-correlated change in net cloud forcing (Fig. 2b). Large regions of substantial low-cloud increases exceeding 0.03, with corresponding NCF changes of more than -5 W m⁻², are prominent in the cooler parts of the subtropical oceans. There are also substantial low cloud increases over broad regions at high latitudes. Smaller regions of decreased low cloud and increased NCF can be seen over low-latitude land and ocean storm-track regions.

Based on current climatology, we might hope that +2K changes in LTS would be a good predictor of changes in low-cloud amount and NCF, at least over the LLO.
Indeed, along with the mean +2K increase of 2.3% in LLO low cloud, there is a mean LTS increase of 1.03 K. However, geographic patterns of +2K LTS change (Fig. 2c) suggest a more complex picture. There are significant increases in LTS even in regions with no low cloud increase, such as the tropical west Pacific. These suggest a tendency for a given low cloud fraction to be associated with larger LTS in the warmer climate, as
also seen in conventional GCM simulations by Meideiros et al. (2008). There is a clear
geographical correlation over the LLO \( (r = 0.59) \) between annual-mean changes in LTS
and changes in low cloud. However, the regional variations of +2K LTS increase over
the LLO are too small to be an attractive explanation of the large regional differences in
low cloud change. The largest low cloud increases occur in subtropical belts, while the
large increases in LTS occur poleward of these belts. Over land and at high latitudes the
relationship between LTS changes and low cloud changes is less evident (and less
expected). Hence, we do not regard the +2K changes in LTS as an adequate explanation
of the +2K changes in low cloud cover and NCF. However, we do regard LTS as a
useful analysis tool for low cloud changes, because it can efficiently sort the LLO into
boundary-layer cloud regimes. This is explored in the next section.

4. Sorting By LTS

To understand the processes responsible for +2K subtropical oceanic low cloud increases,
it is helpful to understand the typical changes in thermodynamic profiles that accompany
them. We use a compositing approach to ensure our analysis is representative of the
entire subtropics. LTS is a logical compositing variable because of its strong connection
to observed low cloud amount and NCF. Compositing into LTS bins will sort the LLO
into deep convection regions (low LTS), trade cumulus regions (intermediate LTS), and
stratocumulus regions (high LTS). This approach follows Bony et al. (2004), but we
choose LTS as a binning variable in place of monthly mean 500 hPa vertical velocity (or
SST, which would be another plausible and simple choice) because it explains a
significantly larger fraction of the observed space-time variance in monthly-mean NCF
For the climatology of each simulation, we calculate the LTS for each month at each grid point. Rather than sorting into bins with fixed LTS range, we sort each into 20 equally sized LTS bins. This sidesteps complications associated with the overall ~1 degree tropics-wide increase in LTS associated with the change of the moist adiabatic temperature profile in a warmer climate.

Figure 3 shows bin-means of low and total cloud fraction, net cloud forcing, and updraft cumulus mass flux (in buoyant saturated updrafts > 1 m s\(^{-1}\)) at 800 hPa, plotted against bin-mean LTS. Observed data sorted using LTS from ECMWF ERA-40 are plotted in Figures 3a and 3b (dotted). The bin-mean values of SP-CAM total cloud fraction only range from 0.30 to 0.41 across LTS bins. Above an LTS of 15K the low cloud amount increases with increasing LTS and dominates the total cloud fraction. The seasonal-mean observed low cloud fraction from Hahn and Warren (2007) has a very similar dependence on LTS, but is uniformly about 0.2 larger than SP-CAM. This discrepancy could be partially due to the different methods of determining low-cloud fraction between observation and model. The LTS binning preserves the geographical anti-correlation of SP-CAM net cloud forcing with the low cloud fraction. The ERBE net-cloud forcing is 10-20 W m\(^{-2}\) less negative than SP-CAM, though with broadly similar LTS dependence.

Figure 3c shows that low LTS regions have high mean cumulus mass flux, corresponding to frequent strong deep convection, while high LTS regions have much less frequent and shallower convection.
Figure 3. LTS-sorted mean SP-CAM fields from the control run and observations. (a) SP-CAM total cloud fraction (solid line), low-cloud fraction (long dashed line) and observed seasonal-mean Hahn and Warren (2007) low-cloud fraction (dotted line). (b) SP-CAM net cloud forcing (solid, W m$^{-2}$) and ERBE (dotted) and (c) upwards cumulus mass flux (kg m$^{-2}$ s$^{-1}$) versus bin-mean LTS. Each symbol represents one 5% bin of column-months.
Figure 4. LTS-sorted mean vertical profiles for the SP-CAM control run. (a) cloud fraction, (b) pressure velocity (hPa day$^{-1}$), (c) relative humidity (%), and (d) radiative heating rate (K day$^{-1}$). LTS increases from left to right. In right column (e-h), the mean difference between the LTS-sorted +2K SST run and the control run is plotted for each variable.

Figure 4a-d shows LTS-composited mean vertical profiles of cloud fraction, pressure velocity $\omega$, relative humidity, and radiative heating rate. There is extensive deep
convective and cirrus cloud, strong mean ascent, and high humidity through the entire troposphere in the weak LTS (warm SST) regions and mean subsidence, little mid and upper-level cloud, and a dry mid-troposphere in the strong LTS regions. Shallow trade-cumulus-like cloudiness maximizing around 900 hPa is present across all LTS bins, but increases at high LTS. At the highest LTS, the ‘boundary layer’ of extensive low cloud and high humidity becomes shallower, resembling fog. Especially for high LTS, there is strong radiative cooling (~2.5 K day⁻¹) in the humid, cloudy air in the upper parts of boundary layer, which helps destabilize it to convection.

We now investigate the +2K LLO cloud response of SP-CAM by applying similar LTS-percentile sorting to the +2K simulation and comparing the sorted results to the control simulation. In the +2K simulation, bin-mean low cloud fraction increases in all tropical LTS bins (dotted line in Fig. 5a), though the largest increases occur in the highest LTS bins. These low cloud changes dominate the changes in total cloud fraction. There is mean negative NCF change across almost all LTS bins; the changes are generally stronger with increasing LTS bin (Fig. 5b) due to the strong low cloud increases.

The increase in mean LTS in each bin is shown in Fig. 5c. The increase in LTS is larger for the higher LTS percentiles, tracking similar trends in NCF and low cloud fraction. In the higher LTS bins from about 14K to 18K (65-95 percentiles), for each 1 K increase in control-run LTS, the SST+2 simulations exhibit an additional increase of 0.07 K in LTS, 1% in low cloud fraction, and -1.4 W m⁻² in NCF. One might conclude that the large increase in low cloud in the high-LTS regimes is due to strengthening of the trade
Figure 5. LTS-sorted changes in SP-CAM. (a) Change in total cloud fraction (solid) and low cloud fraction (dashed), (b) change in net cloud forcing (W m$^{-2}$), and (c) change in LTS (K) versus bin-mean control LTS.

inversion. However, the implied sensitivity of low cloud fraction to LTS in this regime (1% for a 1 K LTS increase) is much stronger than the sensitivity of the control run low cloud fraction to control run LTS (Figure 3 suggests this is less than 0.1% per 1 K LTS increase). This suggests that there is more to the +2K cloudiness changes than changes in inversion stability.

The vertical structure of the low cloud changes is shown in Fig. 4e. For most LTS bins the low cloud increases most at the level with the highest cloud fraction. The total
condensed liquid water content (not shown) increases roughly proportionally to the cloud fraction increases in the shallow clouds. Thus the mean in-cloud liquid water contents of these clouds are not appreciably different in the perturbed climate, in contrast to the cloud feedback mechanism proposed by Somerville and Remer (1984).

The LTS-sorted vertical velocity changes are more varied (Fig. 4f). For high LTS columns, the mean mid-tropospheric subsidence weakens, while near the surface, subsidence stays nearly constant. For moderate LTS columns, where the control vertical velocity is weak, subsidence is strengthened. For weak LTS, where deep convection is concentrated, the mean ascent is strengthened. The low cloud increases occur across all LTS categories, and do not appear to be strongly correlated to these $\omega$ changes.

In the moderate and high LTS columns, the relative humidity increases by 2 or 3% between about 800hPa and 900hPa (Fig. 4g). This increase is centered just above the levels of the largest cloud fraction increase. The radiative cooling also strengthens in these same regions by 0.2-0.4 K day$^{-1}$ (Fig. 4h). Both of these changes are well correlated with the low cloud amount changes.

5. Mechanism of SP-CAM +2K low cloud response

There is little published work convincingly relating the LLO boundary layer cloud response of a GCM to a climate change to a particular physical mechanism operating in that model. Hypothesized physical mechanisms for cloud responses to climate change could be an important organizing tool for testing, analyzing and comparing GCMs, including sharper comparison with observations.
Higher SST

More absolute humidity

More radiative cooling

More convection

More clouds

Figure 6. Schematic of hypothesized SP-CAM +2K low cloud response mechanism.

SP-CAM shows exceptionally large +2K increases in low cloud cover across the subtropics compared to most conventional GCMs. Furthermore, it relies on a CRM (albeit under-resolved) rather than purely a set of interacting physical parameterizations to produce this response. Hence, it seems worthwhile to try to rationalize the low cloud increases in SP-CAM.

In this section, we argue for the following novel radiatively-driven mechanism for this increase, diagrammed in Fig. 6. Higher SST causes a warmer and moister trade-cumulus boundary layer which experiences stronger net radiative cooling. The stronger cooling destabilizes the cumulus layer, leading to more vigorous convection. This fosters a moister boundary layer with more cumulus clouds, which amplifies the anomalous radiative cooling.
Figure 7. LTS80-90 mean vertical profiles. Both control run (black, solid) and +2K SST runs (red, dashed) are plotted. (a) potential temperature (K), (b) relative humidity (%), (c) cloud fraction, and (d) cloud liquid water content (g kg⁻¹). The mean values of LTS, low cloud fraction (CLDLOW) and shortwave cloud forcing (SWCF) in the control run and their changes from the control to the +2K run are shown in panels, (a), (c) and (d), respectively.

The overall +2K increase in lower tropospheric stability helps support this mechanism by keeping the more vigorous convection from enhancing penetrative entrainment of dry air that might evaporate cloud.

We have two pieces of indirect evidence for this mechanism. First, off-line radiation calculations show that even without any cloud or relative humidity increase, the +2K boundary layer would experience significantly stronger radiative cooling. For example, consider the ‘LTS80-90’ composite behavior over the 80-90 percentiles of LTS, chosen as a representative boundary-layer cloud regime with strong low cloud increases, minimal high-cloud effects, and no unrealistic fog. Figure 7 shows the LTS80-90 vertical profiles of θ, relative humidity (RH), cloud cover and cloud liquid water content for the control and +2K runs. These clearly show the +2K increases in boundary-layer cloud cover and
liquid water content, the slight RH increase in the upper boundary layer. The inversion is slightly strengthened, contributing to increased LTS. Figure 8 compares the LTS80-90 composite control and +2K radiative heating profiles (thick lines) with offline calculations of the corresponding clear sky heating rate using LTS80-90 composite temperature and relative humidity profiles and SST (thin lines). The clear-sky radiative cooling in the perturbed climate is 0.1 to 0.25 K day$^{-1}$ stronger than in the control climate throughout most of the cloud layer. This accounts for more than half the overall radiative cooling increase in the cloud layer. The slight +2K RH increase (Fig. 7b) has little impact on the clear-sky cooling. This can be seen by computing the latter using the +2K temperature profile but RH from the control climate (chain-dashed line in Fig. 8).

Instead, the +2K clear-sky cooling increase is driven by the large boundary-layer specific humidity increase due to the warmer temperature profile. The remaining increase in the +2K radiative cooling increase is due to increased low cloud.

A second line of evidence supporting our hypothesized mechanism for low cloud increase comes from diagnosing the monthly-mean net convective heating $Q_{1c}$ in SP-CAM, which we interpret as a measure of the intensity of moist convection. $Q_{1c}$ was calculated as a residual in the SP-CAM heat budget, as we now explain. SP-CAM and SAM use moist-conserved prognostic variables, the liquid-ice static energy $s_{li} = c_p T + gz - L(q_l + q_i) - L_f q_i$ and the total water $q_t = q_v + q_l + q_i$. Here $T$ is temperature, and $q_v$, $q_l$ and $q_i$ are the mixing ratios of water vapor, liquid, and ice.
Figure 8. Vertical profiles of diurnal-mean radiative heating rate for LTS80-90. Full-sky SP-CAM (thick) calculated clear-sky (thin) profiles are plotted for the control run (black, solid) and the +2K SST run (red, dashed). An additional clear-sky profile from the +2K case is calculated (blue chain-dashed line) that uses the RH profile from the control case.

$L$ and $L_f$ are the latent heats of condensation and freezing, respectively. In each SP-CAM grid column (i.e. horizontally-averaged across the CRM simulating that grid column) the budget equations of $s_{li}$ and $q_t$ can be written:

\[
\frac{\partial s_{li}}{\partial t} = c_p Q_t - \left( \mathbf{u} \cdot \nabla s_{li} \right)_h - \omega \frac{\partial s_{li}}{\partial p}
\]

(1)

\[
\frac{\partial q_t}{\partial t} = Q_{qt} - \left( \mathbf{u} \cdot \nabla q_t \right)_h - \omega \frac{\partial q_t}{\partial p}
\]

(2)

Here $Q_t$ [K day$^{-1}$] is the diabatic heating, composed of radiative heating $Q_R$ and convective/turbulent heating $Q_{tc}$, while $Q_{qt}$ [g kg$^{-1}$ day$^{-1}$] is the diabatic moistening. The other terms on the right hand side are the large-scale horizontal and vertical advective
heat and moisture tendencies.

The monthly averages of the total (vertical + horizontal) advective heat and moisture tendencies used to force the CRM in each grid column were saved in the SP-CAM simulations. These include the rectified effect of transients. The composite heat storage is very small, so the LTS-composited diabatic heating rate \( Q_1(p) \) must balance the composite advective heating (and similarly for moisture). The convective heating \( Q_{1c} \) is inferred by subtracting the composite radiative cooling rate \( Q_R(p) \) (Fig. 8) from \( Q_1(p) \).

Figure 9 shows the LTS-binned profiles of composite \( Q_{1c} \) for the control simulation and their change from the control to the +2K simulation. The +2K simulation has 0.2 K (10%) stronger convective heating at 900 hPa for LTS80-90, the same level as the maximum cloud cover increase and just below the level of maximum radiative cooling increase. Similar results are seen in other high-LTS bins. This suggests in the +2K simulation that the cloud increases are associated with more convection.

Finally, the +2K boundary-layer changes have interesting interactions with the large-scale circulation. Within the cloud layer, up to half of the +2K radiative cooling increase is compensated by more convective heating. The remaining cooling drives subsidence. Figure 10a shows the control and +2K LTS80-90 composite vertical motion (thick lines, the thin lines represent forcing for the CRM experiments discussed in Sect. 7b below). In the mid-troposphere (600 hPa), radiation dominates the diabatic cooling. There is no +2K radiative cooling change, and since the +2K thermal stratification is stronger, the unchanged diabatic cooling drives weaker subsidence.
Figure 9 Monthly LTS-sorted profiles of convective heating $Q_{1c}$ (K day$^{-1}$). (a) Control run and (b) the change from the control to the +2K SST simulation.

Figure 10. (a-b) SP-CAM LTS80-90 composite profiles. Profiles are shown for SP-CAM control (black, solid) and the +2K simulation (red, dashed). (a) Pressure-velocity ($\omega_0$, hPa d$^{-1}$, thick) and perturbed pressure-velocity ($\omega_0 + \omega'$, thin) used in the CRM, (b) temperature advection, and (c) humidity advection. The blue chain-dashed line in (c) is the horizontal humidity advection predicted for the +2K run if relative humidity from the control run is used.

In the cloud layer (900 hPa), strengthened diabatic cooling balances the stronger stratification and subsidence remains just as strong in the +2K climate. Subsidence
is viewed here as a feedback on the column diabatic processes rather than a fundamental external control.

Could changes in horizontal advection drive the cloud response? The dashed lines in Figs. 10b and 10c show the LTS80-90 horizontal advective forcing of temperature and specific humidity estimated as the difference between the composite SP-CAM total advective forcing and the vertical advective forcing. The horizontal advective heating profile (Fig. 10b) shows very little +2K change. Since these changes are small compared to the clear-sky radiative cooling changes (Fig 8, thin lines), they probably are not the main driver of the +2K SP-CAM low cloud increase.

The +2K increase in amplitude of the horizontal advective moistening (Fig. 10c), including stronger drying in the cloud layer and at the surface, is largely attributable to the Clausius-Clapeyron effect. When the +2K moisture advection is instead calculated using temperatures and winds from the +2K run but relative humidity fields from the control run (blue chain-dashed curve), the result is very similar to the original +2K moisture advection (red dashed curve). The +2K increase in advective drying also does not help explain the low cloud increase.

6. Column modeling of +2K LLO boundary layer cloud response

In the next sections, we introduce a column modeling framework that allows us to test the sensitivity of the SP-CAM +2K LLO low cloud response to the CRM resolution. We will deliberately use the term ‘column modeling’ rather than ‘single-column modeling’, because as in SP-CAM, we run a CRM with prescribed large-scale advective forcings
Our framework is related but not identical to those used in recent studies by Zhang and Bretherton (2008) and Caldwell and Bretherton (2008). Two key assumptions that we share with these two studies are: (a) in low latitudes, gravity waves adjust the temperature profile in each tropospheric column towards a roughly moist-adiabatic profile that is controlled remotely by deep convection over the warmest oceans, and (b) in the synoptically quiescent subtropical trades, the boundary layer cloud response to climate change is mainly due to changes in the mean state, rather than in the transient variability. Like these studies, our column methodology is applicable to GCMs other than SP-CAM. We adapt their methodology to more quantitatively replicate the climatology and +2K cloud response of the studied GCM in a boundary-layer cloud regime defined using the LTS80-90 percentile range. We then test whether this cloud response is robust to a better representation of the boundary layer than used in the GCM.

Zhang and Bretherton (2008) examined the +2K response of subtropical stratocumulus in a single-column version of a conventional GCM, the CAM. Caldwell and Bretherton (2008, manuscript submitted to J. Climate, hereafter CB08) analyzed the climate change response of a mixed-layer model of the stratocumulus-capped boundary layer. Both studies adjusted the free-tropospheric temperature toward a moist adiabat related to warm-ocean SST (with significant methodological differences), assumed a constant free-tropospheric relative humidity, and let the column model drive the boundary-layer turbulent mixing and cloud. Both models made strong simplifying assumptions about
horizontal advection and they computed mean vertical velocity in the column based on a heat budget. That is, mean subsidence in each column adjusts to keep the temperature roughly anchored to the same moist adiabat. CB08 used a more sophisticated representation of these feedbacks than did Zhang and Bretherton (2008).

Each of these studies, like SP-CAM, predicted a boundary-layer cloud albedo increase (negative cloud feedback) in the warmer climate. However, the physical mechanisms were quite different, both from each other and from the radiatively-driven mechanism for low cloud increase that appears to operate in SP-CAM. Zhang and Bretherton (2008) found that a complex and somewhat unphysical interaction between many moist physical parameterizations produces a +2K boundary-layer cloud albedo increase in the single-column version of CAM in a cloud regime comparable to our LTS80-90 case. This increase is also seen in global CAM simulations, suggesting that the +2K cloud response in the global simulations may be artificial. However, both the mean state and +2K change in the boundary-layer clouds were rather different between their single-column model and the global CAM simulations of the corresponding subtropical cloud regime. CB08 found that stratocumulus clouds thicken in a +2K climate due to ‘subsidence lapse-rate’ feedback, in which the inversion strengthens (suppressing dry air entrainment and lowering cloud base) and mean subsidence weakens in the warmer climate (so the cloud top can nevertheless rise). The net result is an increase in cloud thickness. Their result applies only to the small fraction of the subtropical oceans usually covered by stratocumulus-capped mixed layers, a regime poorly simulated in SP-CAM due to inadequate CRM resolution.
7. SP-CAM column analogue – description.

Because our SP-CAM analysis suggests the low cloud increases are quite pervasive across the subtropical oceans, we could pick some individual LLO grid column for this test, compute mean forcings for this column, and run our column model with these forcings. To more convincingly connect the results to average trends across the entire LLO, we instead apply this approach to a composite over all LLO grid column-months in a particular climate regime. We focus hereafter on the LTS80-90 regime analyzed in Section 5, which has strong +2K low cloud increases, minimal cirrus, and no coastal fog.

We run the CRM to steady state with the composite forcings for the control climate and repeat for the +2K climate. Let C be a key summary cloud statistic (e.g. low cloud cover or SWCF), with value $C_{CM}$ in the control-forced CRM and value $C_{GCM}$ in the control GCM composite. Let the difference in $C$ between the two CRM simulations be $\Delta C_{CM}$. If $C_{CM}$ is comparable to $C_{GCM}$ and $\Delta C_{CM}$ is comparable to $\Delta C_{GCM}$ then the column analogue has skill and further analysis of its low cloud sensitivity of the single column model to the +2K forcing change is warranted. Otherwise, the time averaging and LTS-compositing has removed essential ingredients of the +2K low cloud changes.

a. CRM configuration and boundary conditions

We configure the SAM CRM to use the same algorithms as in SP-CAM for microphysics, radiation, surface fluxes. SST and downwelling solar radiation at the top of the model domain are taken from the SP-CAM LTS composite, with the solar zenith angle fixed at the annual-mean diurnal-mean for 22°N.
We use 2 two-dimensional configurations of domain and resolution for the CRM, as listed in Table 1. The ‘4kmL30’ configuration matches the SP-CAM CRM in domain size and resolution. A natural advantage of using a GCM with an embedded CRM is the ability to more realistically examine cloud processes and cloud changes using identically forced offline CRM experiments with higher resolution. These can help establish whether the CRM configuration used in SP-CAM is adequate to examine the low cloud response. Hence, we also consider an ‘LES’ configuration of the identical CRM with much higher horizontal/vertical grid spacings of 100/40 m in the lower troposphere and a shorter domain length of 51.2 km. These grid spacings were chosen to match Siebesma et al. (2003)’s intercomparison of nonprecipitating trade cumulus convection, which showed little sensitivity of the simulated cloud statistics to choice of CRM or further increases in resolution. Tests with a longer LES domain did not show significant differences in the mean profiles achieved in the steady state. A three-dimensional LES configuration was too computationally expensive for the long integrations performed here.

Table 1. CRM Configuration parameters.

<table>
<thead>
<tr>
<th>Configuration</th>
<th>Domain Size (km)</th>
<th>Resolution (m)</th>
<th>Grid Points</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$x$</td>
<td>$z$</td>
<td>$x$</td>
</tr>
<tr>
<td>4kmL30</td>
<td>128</td>
<td>40</td>
<td>4000</td>
</tr>
<tr>
<td>LES</td>
<td>51.2</td>
<td>27</td>
<td>100</td>
</tr>
</tbody>
</table>
An idealized vertical profile of winds is constructed for the CRM simulations (Fig. 11). The horizontal velocity components are not directly composited because this would produce unrealistically weak wind speed and vertical shear. Instead the monthly-mean wind speed (which includes rectification from transients) is composited at each level. An idealized profile of wind directions, based on the RICO wind profile used by Abel and Shipway (2007), is combined with the composite wind speed profiles to create $u$ (E-W) and $v$ (N-S) velocity profiles. These are used to initialize the runs and to nudge their domain-mean velocity profiles with a 10-minute timescale.

The CRM model domain is oriented N-S, as in SP-CAM. Identical 4kmL30 CRM simulations except with an E-W oriented domain produce mean cloud cover and +2K
cloud changes that are both considerably closer to those simulated by SP-CAM, showing that the orientation of the wind profile with respect to the domain does quantitatively affect the results. This is because only the shear component along the horizontal domain axis affects the simulation.

b. Vertical velocity feedback for the CRM

The CRM experiments are initialized with the SP-CAM LTS80-90 composite profiles of temperature and moisture (Fig. 7) with small-amplitude white noise in temperature added to initiate convection. The steady state reached depends on how the model is forced. This requires one to make choices about the specification of advective forcings, wind direction, and nudging. Unlike the past idealized studies discussed in Sect. 6, we advectively force and nudge the mean state toward the SP-CAM thermodynamic profiles rather than a moist adiabat, so that our column simulation creates a cloud environment as close as possible to the LTS80-90 composite.

We force the CRM with the LTS80-90 composite of the SP-CAM large-scale horizontal advective tendencies and vertical pressure-velocity shown in Fig. 10. In SP-CAM, the composite diabatic heating and moistening rates closely balance the sum of the resulting horizontal and vertical advective tendencies. However, the CRM diabatic tendencies inevitably will not exactly balance the SP-CAM composite advective tendencies, so the CRM mean profiles start to drift away from the SP-CAM profiles to which they are initialized. Slow but persistent drifts in the free troposphere must be controlled so that the simulated boundary layer evolves under free-tropospheric profiles that remain close to the SP-CAM composite. For instance, the SP-CAM composite has slight
advective moistening in the upper troposphere to balance drying from ice precipitating out of occasional high cirrus, even though the composite sounding is not ice-saturated at upper levels. This advective moistening slowly moistens the CRM simulation in the upper troposphere. If uncompensated, this drift would ultimately lead to an unrealistic sheet of thin cirrus near the tropopause.

The scheme used to control these drifts can affect the +2K sensitivity of the results. Ideally, the column model should be able to properly represent the same dynamical feedbacks that accompany LLO low cloud changes in the global model – in particular the nonlocal control of the temperature profile via stratified adjustment, and the corresponding adjustment of mean vertical motion to compensate changes in diabatic heating or stratification. In this spirit, we use the following adaptation of CB08’s algorithm for relating vertical velocity changes to diabatic heating changes in a column setting. Let subscript ‘0’ denote a reference profile (in our case, from the SP-CAM LTS80-90 composite), and let a prime denote a perturbation of a CRM horizontal mean quantity from that reference profile. We use the reference horizontal advective tendencies, but we let the mean vertical velocity used to compute vertical advective tendencies respond to the instantaneous profile of column virtual temperature (i. e. density) anomalies. That is, we take $\omega = \omega_0 + \omega'$, where $\omega_0(\rho)$ is the SP-CAM reference profile and $\omega'(\rho, t)$ is a diagnosed pressure velocity perturbation.

Following CB08, the approach is to consider the column to be representative of the center of an anomalous region of some half-width $\Lambda$. The density profile across this region is assumed to have a horizontally sinusoidal variation with wavenumber $k = \pi/(2 \Lambda)$ such
that the density anomaly vanishes at a distance $\Lambda$ from the column. We compute $\omega'$ from the linear quasi-steady damped inertia-gravity-wave adjustment of this horizontally-varying density field to the reference density profile. The tropopause and surface are approximated as rigid boundaries. We assume a Rayleigh damping rate $a_m$ for momentum perturbations. This results in the following boundary-value problem, derived in Appendix A, which is solved at each timestep for $\omega'(p)$:

$$\frac{\partial^2 \omega'}{\partial p^2} \approx r \frac{T'}{p}, \quad \omega'(p_S) = \omega'(p_T) = 0.$$  

(5)

where $r = a_m R_d / (f^2 + a_m^2)$. Here $f = 5 \times 10^{-5}$ s$^{-1}$ is a representative subtropical value of the Coriolis parameter, and $p_S = 1016$ hPa and $p_T \approx 100$ hPa are the surface and tropopause pressure, respectively. We choose $a_m = 0.5$ d$^{-1}$ to be broadly characteristic of the lower troposphere, where the virtual temperature perturbations from the reference profile are the largest. We also somewhat arbitrarily choose $k = 2.4 \times 10^{-6}$ m$^{-1}$, corresponding to a half-width $\Lambda = 650$ km for the cloud regime we are considering. Together, these parameter choices imply $r = 3.8 \times 10^{-6}$ K$^{-1}$s$^{-1}$.

An independent estimate $r \approx 4.3 \times 10^{-6}$ K$^{-1}$s$^{-1}$ can be obtained by comparing LTS-sorted bin-mean profiles of $\omega$ and $T_v$ in the lower troposphere, now treating the perturbation as being the difference between the bin-mean and the tropics-wide mean. The similarity of the estimate to our choice of $r$ is reassuring, and suggests that the magnitude of vertical velocity feedback we are using is comparable to that apparent in the mean tropical Hadley-Walker circulation. It also suggests that Eq. (5) may also be suitable for more
idealized frameworks for studying cloud feedbacks such as proposed by Zhang and Bretherton (2008), without large changes in $r$.

Our approach bears some resemblance to Raymond and Zeng’s (2005) algorithm for enforcing the weak temperature gradient approximation in a cloud-resolving model. They assumed $\omega'$ is proportional to $-a_{WTG}T_v'$, with a relaxation rate $a_{WTG}$ that goes to zero at the surface and tropopause and maximizes in the mid-troposphere. With an appropriate choice of $a_{WTG}$, this will roughly the same effect as (5) for gravity waves with a half-wavelength that spans the troposphere, but then it will distort the $\omega'$ response to sharper (shorter vertical wavelength) perturbations in $T_v'$ in a manner that depends on how far they are from the top or bottom boundary. Unlike Raymond and Zeng’s approach, our algorithm (5) has the attraction of being formally derivable from a physically-based model of the dynamical feedbacks on horizontally localized density perturbations.

We believe that the approach proposed here is superior to that proposed by CB08 for our application. While the two approaches start with a similar underlying model of the dynamical adjustment process, CB08 neglected momentum damping. After some approximation, they were ultimately able to relate $\omega'$ to the profile of perturbation diabatic heating rate in the simulated column. At least in the lower troposphere, momentum damping is probably quite significant for gravity waves of short vertical wavelength propagating thousands of kilometers across the tropics. For instance, a hydrostatic gravity wave of vertical wavelength 3 km (which projects strongly onto the strong and sharp horizontal temperature variations seen around the trade inversion) has a horizontal phase speed of roughly 5 m s$^{-1}$. In the damping timescale $a_m^{-1} = 2$ d, it will
only propagate 1000 km, so it experiences considerable momentum damping in the regime half-width $\Lambda = 650 \text{ km}$. Thus, CB08’s neglect of momentum damping was probably not fully justified for our application. In addition, an approach based on temperature rather than heating rate perturbations has practical advantages when used with a CRM of finite domain size. The CRM-simulated domain-mean heating rate varies much more from time step to time step than does the domain mean temperature, violating the quasi-steady assumption underlying the $\omega'$ feedbacks derived by CB08, even while the $\omega'$ derived from Eq.(5) varies only slowly from timestep to timestep. That said, for this study we also tried implementing variants of the CB08 approach to diagnosing $\omega'$, and with appropriate parameter choices we were able to obtain results quite similar to those shown in this paper (but not shown here due to space limitations).

c. CRM moisture nudging

The $\omega'$ feedbacks controls temperature drifts in a physically appropriate way, but will not entirely prevent systematic drift of the upper-tropospheric humidity from the reference profile, which is an inevitable consequence of using a quasisteady CRM simulation to represent a composite over many times and locations. Hence, we add a moisture relaxation term with a height dependent rate

$$a_q(z) = a_{FT}\{1 + \text{erf}[(z - z_0)/b]\}/2,$$  \hspace{1cm} (6)

where $a_{FT} = (1 \text{ d})^{-1}$, $z_0 = 3 \text{ km}$, and $b = 1.5 \text{ km}$. This nudging rate ramps from close to zero at the surface to $a_{FT}$ in the upper troposphere, with the transition mainly occurring
just above the trade inversion in the height range $z_0 \pm 2b/3 = 3\pm1$ km.

We also tried pure temperature nudging instead of $\omega$-feedback to control temperature drift, using a damping rate profile similar to Eq. (5). This approach gives LTS80-90 simulated boundary layer cloud and +2K response that are fairly similar to those for vertical velocity feedback. In corresponding LTS70-80 simulations (based on a composite of the 70-80th LTS percentiles of SP-CAM), temperature nudging is less successful than $\omega$-feedback in reproducing the SP-CAM +2K cloud changes though both methods replicate the control-state clouds fairly well. The $\omega$-feedback also produces a slightly better match than temperature nudging between the CRM-simulated cloud-layer relative humidity profile and the SP-CAM composite.

8. SP-CAM column analogue –4kmL30 results

Each simulation reaches statistically-steady horizontal-mean temperature and moisture profiles within 20 days. Low cloud fraction (shown in Fig. 12 for the 4kmL30 simulations) adjusts even more rapidly. Once equilibrated, its daily mean is uncorrelated between successive days, and its daily standard deviation in both runs is approximately 0.02. This allowing the random sampling uncertainty to be estimated for an averaging period of any length. Because we are interested in the differences between two rather similar simulations, we must be sure to use a sufficiently long averaging period so their differences overwhelm the sampling uncertainty in this and other cloud-related variables. The ‘steady-state’ profiles presented below are averaged over the period 20-60 days.
Figure 12. Time series of column cloud fraction for 4kmL30 runs. Control (black, solid) and +2K (red, dashed) runs are shown. Cloud fraction is defined as the fraction of domain columns with optical depth > 0.3 from 700hPa to the surface.

For cloud fraction, the sampling uncertainty for a 40-day period is approximately $0.02/(40)^{1/2} = 0.3\%$ for each time series, and $2^{1/2}$ as large, or roughly 0.4%, for the cloud cover difference of the +2K and control runs. This is about 10% as large as the mean cloud cover difference over this time interval, which is sufficiently small to consider our results to be representative.

Figure 13 plots the steady-state profiles of potential temperature, relative humidity, cloud fraction, and liquid water content reached by the LTS80-90 forced control and +2K 4kmL30 runs. These can be compared with the corresponding SP-CAM composite profiles shown in Fig. 7.
Figure 13. Horizontal mean profiles of the 4kmL30 runs for LTS80-90, with variables as in Fig. 7. Control (black, solid) and +2K run (red, dashed) profiles are averaged over days 20-60.

The 4kmL30 temperature and relative humidity profiles above the boundary layer match the SP-CAM composites fairly closely because of the strong nudging and relatively slight imbalance between CRM advective and diabatic forcings at these levels. The CRM deviates more from the SP-CAM composite profiles in the boundary-layer cloud layer. The cool, moist boundary layer becomes deeper and is capped by a more pronounced inversion in 4kmL30 than in SP-CAM. We speculate that this occurs because in SP-CAM, synoptic transients episodically depress the trade inversion, replacing moist cool air by warmer and drier air. By construction, this cannot occur in the steadily forced CRM.

Nevertheless, the cloud fraction, liquid water profile, SWCF and their +2K sensitivity (Fig. 13c) agree fairly well between 4kmL30 and SP-CAM. Table 2 summarizes statistics of all of the CRM runs over the 20-60 day period for both the control and the +2K
changes. The CRM has a somewhat higher mean and a weaker +2K sensitivity of low cloud fraction, cloud liquid water path, and SWCF. Table 2 also shows corresponding results for the CRM run with LTS70-80 forcing, which show similarly good agreement with SP-CAM. Hence, we regard the 4kmL30 simulation as a useful column analogy to the SP-CAM composite cloud response.

Table 2 also shows that the control SP-CAM composite rainfall rate is about 40% larger than in 4kmL30 for LTS80-90 and 170% larger for LTS70-80. This is because the SP-CAM composites include occasional episodes of deeper convection as well as the light shallow convective rainfall simulated by the CRM. The heating and drying from these episodes affect the SP-CAM composite heat and moisture budgets.
Figure 14. Horizontal mean profiles of the 4kmL30 run. (a) Radiative heating (K d$^{-1}$), (b)–(c) tendencies of temperature (K d$^{-1}$) and humidity (g kg$^{-1}$ d$^{-1}$) from perturbation vertical advection, and (d) moisture relaxation tendency (g kg$^{-1}$ d$^{-1}$).

However, in both cases the precipitation is much smaller than evaporation from the sea surface, and similarly for its +2K changes. Hence, the episodic precipitating deep convection does not appear to be seriously perturbing the shallow cumulus layer heat and moisture budgets and their +2K response. This is consistent with the similarity in the SP-CAM +2K cloud response across the high LTS percentiles, despite substantial differences in their mean precipitation rate.

Figure 14 shows the radiative heating and the nudging terms in the 4kmL30 CRM simulations. The mean control radiative cooling (Fig. 14a) is enhanced compared to SP-CAM (Fig. 8) in the 850-900 hPa layer in both the control and +2K simulations because of higher RH and cloud. The +2K 4kmL30 CRM simulation has slightly stronger +2K radiative cooling in the cloud layer than the control, though the increase is much less than that of the SP-CAM (Fig. 8). The control vertical advective temperature tendency (Fig.
14b) compensates for most of the radiative cooling, but the +2 K increase in subsidence warming is not as large as the radiative cooling increase. Subsidence drying (Fig. 14c) balances upward convective and turbulent moisture transport. Within the cumulus layer (below 800hPa), the moisture nudging (Fig. 14d) is much weaker than the vertical (Fig. 14c) and horizontal (Fig. 10c) moisture advection and also shows little +2K sensitivity. We infer that the moisture nudging does not seriously distort the cloud-formation processes in the cumulus layer. Above the cumulus layer, the nudging acts to moisten the 700-800 hPa layer where the CRM does not generate as much cumulus-induced moistening as does SP-CAM (again probably due to the lack of synoptic variability that can produce somewhat deeper cumuli). Overall, the physics of the 4kmL30 +2K response seems similar to that diagnosed for SP-CAM as a whole – more radiative cooling leads to more shallow convection and associated cloud, and is partially balanced by stronger subsidence warming which also helps keep the boundary layer from deepening.

9. Sensitivity to CRM resolution

We applied the column analogue to study the sensitivity of the SP-CAM cloud climatology and +2K response to the CRM resolution. To do this, we ran the CRM to steady state using the same LTS80-90 control and perturbed forcing, but with the ‘LES’ resolution given in Table 1. Figure 15 shows steady state profiles for the control and +2K LES cases. Comparison of Figs. 15a and 15b with Figs. 7a and 7b shows that the control temperature and relative-humidity profiles of LES and 4kmL30 are somewhat similar. However, a sharper inversion is clear in the LES simulation.
Furthermore Figs. 15c-d show that the LES simulation has much less cloud and condensate at all heights, producing a low cloud fraction is only 40% as large and the SWCF only 22% as large as for the 4kmL30 resolution. In both simulations, the cloud cover, condensate and relative humidity have relative maxima at the inversion base, suggestive of a tendency toward of a cumulus-under-stratocumulus. Similar results (not shown) were found with LTS70-80 forcing, but the LES cloud and condensate profiles, like their 4kmL30 analogues, were more weighted toward the bottom of the cloud layer, indicative of a pure shallow cumulus structure with less residual cloud at the trade inversion base.

Given the differences in simulated mean state, it is no surprise that the +2K cloud response in the LES simulations also differs considerably from the 4kmL30 simulations. For the LTS80-90 case, the LES-predicted +2K response in low cloud fraction and SWCF (Fig. 15c-d and Table 2) is comparable to that of the 4kmL30 simulation. However, its vertical structure is different than the 4kmL30 simulation, with the cloud
increase localized in the layer below the inversion. There is also a notable increase in the cloud-layer relative humidity in the +2K LES case not seen at 4kmL30 resolution.

For the LTS70-80 case the LES-predicted cloud and condensate (not shown) also increase near the trade inversion in the +2K simulations, but the increase is much smaller, so that the cloud cover and SWCF (given in Table 2) are only slightly larger than in the control run. This result contrasts strongly with the strong +2 K low cloud increase seen in the 4kmL30 simulations and the SP-CAM in this climate regime.

The coarse-resolution cloud biases can be rationalized as follows. The 4 km horizontal resolution forces cumulus updrafts to be far too broad and weak. Updrafts that are too weak must have an excessive fractional area at cloud base in order to flux enough moisture out of the boundary layer to balance the surface evaporation. This causes the coarse-resolution simulation to overestimate cloud fraction just above cloud base. On the other hand, cumulus updrafts that are too weak are insufficiently penetrative when they overshoot their level of neutral buoyancy. This keeps the cumulus layer too shallow at SP-CAM resolution. In addition, the SP-CAM vertical resolution is very coarse, especially at heights above 2 km. Other CRM resolution sensitivity tests that we have performed suggest that this vertical under-resolution slightly inhibits the deepening of the shallow cumulus cloud field.

10. Discussion and Conclusions

The SP-CAM exhibits large increases in low cloud cover in a climate in which SST is artificially warmed by 2 K. Much of this cloud increase occurs in subtropical marine
boundary layers. Lower tropospheric stability (LTS) is used to analyze the cloud changes. In the Tropics, high LTS is closely correlated with cool SST. The overall tropical LTS increases by 1 K even in regions of no low cloud increase. Over the cooler subtropical oceans, there is a slightly larger LTS increase and a low cloud cover increase of over 5%.

We composited SP-CAM monthly-mean output from all oceanic low-latitude grid columns into climate regimes defined using percentiles of LTS. By comparing the SP-CAM control and +2K climates, we argued that the low cloud increase is ultimately driven by an increase in clear-sky radiative cooling in the relatively humid boundary layer air. This destabilizes the boundary layer, stimulating more convection, more cumulus cloud and further amplifying the radiative cooling. The vertical structure of the clouds shows little change in the warmer climate due to compensation between increased radiative destabilization of convection and inhibition of convective deepening by increased lower-tropospheric static stability. Horizontal advective forcing and boundary-layer winds are insensitive to the +2K forcing changes and appear to play little role in the SP-CAM cloud response.

The boundary-layer radiative feedbacks discussed here may be quite sensitive to the type of climate perturbation. For instance, the increase in clear-sky boundary-layer radiative cooling in a climate warmed by doubled CO₂ might be rather weaker due to the system-wide radiative balance constraint. This might lead to much weaker cloud feedbacks than in the +2K case studied here. Conventional AGCMs often show rather different cloud responses to these two climate perturbations (e.g. Wyant et al 2006b). Studying the
response of SP-CAM to an instantaneous CO₂ doubling with fixed SST could yield preliminary insights without the computational expense of a 20+ year simulation over a mixed-layer ocean (Gregory and Webb 2008).

A column analogue to SP-CAM is constructed and compared with the SP-CAM in two trade cumulus regimes defined by the 70-80ᵗʰ and 80-90ᵗʰ percentiles of LTS. It uses a single CRM interacting with a free troposphere in which subsidence is allowed to adjust to the column diabatic cooling to keep the temperature profile close to the SP-CAM composite, with additional humidity relaxation above the cumulus layer. For both regimes, the column analogue produces an encouragingly similar cloud climatology and +2K cloud response to SP-CAM when the CRM is run with the same coarse 4 km, 30 level resolution. However, the mean cloudiness dramatically decreases in this column analogue if the CRM is instead run with a typical LES resolution used for shallow cumuli. The +2K cloud response also is resolution-dependent. Unlike with coarse resolution, the CRM with LES-like resolution predicts a substantial cloud increase at the warmer SST only in the more stable 80-90ᵗʰ LTS percentile regime, and this cloud increase occurs almost exclusively at the trade inversion. The implication is that a superparameterized GCM or global CRM likely needs very high resolution to realistically simulate subtropical boundary layer cloud feedbacks on climate. While the required resolution is computationally infeasible for a global CRM for the foreseeable future, it might be within reach of a suitably designed superparameterized GCM.

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**Appendix. Vertical velocity feedback equation**

As in CB08, we consider a system in which domain-mean anomalies of virtual temperature and diabatic heating in the simulated column are assumed to be representative of a region of characteristic half-width $\Lambda$. The resulting horizontal buoyancy gradients generate vertical motions through linear gravity wave dynamics. We modify CB08’s derivation (which did not include momentum damping) by allowing a momentum damping rate $a_M$. For simplicity, we do not consider the advection of anomalies by the reference-state vertical and horizontal winds, which (as shown via scaling analysis in CB08) should not influence the result.

We assume a constant Coriolis parameter $f$. Our vertical velocity equation is derived assuming the column density anomalies are ‘quasi-steady’, i. e. varying on timescales much longer than $a_m^{-1}$ and $f^{-1}$, so that we can neglect time-derivative terms in the governing equations. This is not an important restriction, since we run the column model to a statistical steady state.

We consider an arbitrary small perturbation $F^*(x, p, t)$ of all variables $F$ about the reference state $F_0$. For simplicity, this perturbation includes variability in only one horizontal dimension, but the same result would also be obtained assuming sinusoidal horizontal variability in two horizontal dimensions. The linear, damped, hydrostatic, quasi-steady momentum and mass conservation equations in pressure coordinates can be
written

\[ a_m u^* - f v^* = -\partial \phi^* / \partial x, \]  
(A1)

\[ a_m v^* + f u^* = 0, \]  
(A2)

\[ \partial \phi^* / \partial p = -R_d T_v^* / p, \]  
(A3)

\[ \partial u^* / \partial x + \partial \omega^* / \partial p = 0. \]  
(A4)

These equations can be combined into a single equation relating \( \omega^* \) to the virtual temperature perturbation \( T_v^* \):

\[ \frac{\partial}{\partial p} a_m^{-1} \left( f^2 + a_m^2 \right) \omega^* = -\frac{R_d}{p} \partial^3 T_v^*. \]  
(A5)

At the surface and the tropopause, we assume rigid-lid boundary conditions and neglect Eulerian pressure tendencies so

\[ \omega^*(x, p_s) = \omega^*(x, p_r) = 0. \]  
(A6)

We assume the anomalies of all variables \( F \) have the spatial structure

\[ F^*(x, z, t) = F'(z, t) \cos(kx), \]  
(A7)

where \( k = \pi/(2\Lambda) \) is a user-specified characteristic horizontal wavenumber. Primes denote the column anomalies from the reference state. Substituting (A7) into (A5),
\[
\frac{\partial}{\partial p} a_m^{-1} \left( f^2 + a_m^2 \right) \frac{\partial \omega'}{\partial p} \approx \frac{R_f k^2}{p} T_v'.
\] (A8)

Assuming \(a_m\) to be constant, Eq. (5) is straightforwardly obtained from (A8) and (A6). In (A5), \(a_m\) can be vertically varying. It is plausible that \(a_m\) should be chosen larger near the surface than in the upper free troposphere. However, for now we regard this as unduly complex for our highly idealized model.
References


