Cloud-Resolving Simulation of Low-Cloud Feedback to an Increase in Sea Surface Temperature

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ABSTRACT

This study investigates the physical mechanisms of the low cloud feedback through cloud-resolving simulations of cloud-radiative equilibrium response to an increase in sea surface temperature (SST). Six pairs of perturbed and control simulations are performed to represent different regimes of low clouds in the subtropical region by specifying SST differences ($\Delta$SST) in the range of 4 and 14 K between the warm tropical and cool subtropical regions. The SST is uniformly increased by 2 K in the perturbed set of simulations. Equilibrium states are characterized by cumulus and stratocumulus cloud regimes with variable thicknesses and vertical extents for the range of specified $\Delta$SSTs, with the perturbed set of simulations having higher cloud bases and tops and larger geometric thicknesses. The cloud feedback effect is negative for this $\Delta$SST range ($-0.68$ to $-5.22$ W m$^{-2}$ K$^{-1}$) while the clear-sky feedback effect is mostly negative ($-1.45$ to $0.35$ W m$^{-2}$ K$^{-1}$). The clear-sky feedback effect contributes greatly to the climate sensitivity parameter for the cumulus cloud regime whereas the cloud feedback effect dominates for the stratocumulus regime. The increase of liquid water path (LWP) and cloud optical depth is related to the increase of cloud thickness and liquid water content with SST. The rates of change in surface latent heat flux are much higher than those of saturation water vapor pressure in the cumulus simulations. The increase in surface latent heat flux is the primary mechanism for the large change of cloud physical properties with $+2$ K SST, which leads to the negative cloud feedback effects. The changes in cloud fraction also contribute to the negative cloud feedback effects in the cumulus regime. Comparison of these results with prior modeling studies is also discussed.

1. Introduction

How the global cloud distributions vary with changes in both the circulation patterns and underlying surface temperature [e.g., sea surface temperature (SST)] in response to anthropogenic forcings is not well understood. The altered cloud distributions lead to an imbalance in global radiative budget. Whether the radiative imbalance can enhance or reduce the changes in the circulation patterns and underlying surface temperature—the cloud feedbacks—is highly uncertain (Stephens 2005; Bony et al. 2006; Randall et al. 2007). This is related to the fact that solar and infrared radiative impacts of low-level clouds are different from those of high-level clouds, variations in the geographic distributions of the different cloud types are sensitive to variations in the atmospheric circulations, and modifications in cloud physical properties such as the altitudes of clouds and their optical thicknesses and inherent relationships may occur as climate changes.

The roles of low-level clouds in the climate system are well known (e.g., Randall et al. 1984), but their climate feedbacks are very uncertain, primarily owing to inadequate representations of physical processes of low-level clouds in GCMs (e.g., Bony and Dufresne 2005;
Randall et al. 2007). These clouds are close to the earth’s surface and are prevalent in the cold regions of subtropical oceans west of the continents (e.g., Xu et al. 2008) where the descending branches of the Hadley and Walker circulations meet. Their main radiative effects are associated with the reflected solar radiation at the top of the atmosphere (TOA) and produce cooling at the surface. Their area-averaged TOA albedo—the ratio of the reflected solar radiation to solar insolation—is determined by many factors. Among them, cloud fraction and cloud liquid water path (LWP) are the primary factors. An increase in the value of either factor increases the TOA albedo and increases cooling at the ocean’s surface. Cloud fraction is associated with the morphology of cloud fields or cloud macrophysical properties, which are sensitive to changes in the atmospheric circulation regimes and thermodynamic stability of the atmosphere. LWP is a product of the cloud liquid water content (LWC) and cloud thickness. It is mostly associated with cloud microphysical properties and determined by cloud-scale dynamical processes such as cloud-top entrainment and turbulent transports in the cloud and subcloud layers. The alterations in the altitudes of these low clouds can also impact the infrared radiative forcing, which is believed to be small compared to the solar radiative forcing for the low-level clouds.

One of the cloud feedback mechanisms is the so-called negative cloud optical depth ($\tau$) feedback (Paltridge 1980; Somerville and Remer 1984), which was based on the dependency of LWC on the ambient temperature ($T$). An empirical relationship between LWC and $T$ was inferred from aircraft measurements of clouds in midlatitude regions (Feigelson 1978). The rate of the increase, $\frac{d \ln(\text{LWC})}{dT}$, was estimated to be 0.04 to 0.05 K$^{-1}$ for the temperature range from $-25^\circ$C to $5^\circ$C, slightly less than that for adiabatic clouds ($\sim$0.06 K$^{-1}$) (Betts and Harshvardhan 1987). Somerville and Remer (1984) obtained a negative cloud feedback using this empirical relationship and a constant cloud thickness assumption. Because $\tau$ is a product of LWC and cloud thickness divided by equivalent droplet radius $r_c$, this relationship implies that $\tau$ increases with increasing $T$ if other cloud properties are unchanged with $T$. As $\tau$ increases, more solar radiation is reflected back to space, which cools the atmosphere and the surface—thus a negative cloud feedback.

On the other hand, recent satellite observations from the International Satellite Cloud Climatology Project (ISCCP), Advanced Very High Resolution Radiometer (AVHRR), and Clouds and the Earth’s Radiant Energy System (CERES) (Tsienouidis et al. 1992; Chang and Coakley 2007; Eitzen et al. 2008) indicate that $\tau$ for oceanic low clouds may decrease, rather than increase, with $T$ (either the cloud-top temperature or the SST), thereby suggesting a positive cloud optical depth feedback to a climate warming. The range of the temperature variations in these oceanic observations is probably as large as that of cloud ambient temperature variations of continental clouds obtained by aircraft measurements (Feigelson 1978; Mazin 1995; Gultepe and Isaac 1997). However, the oceanic low clouds were observed at higher $T$ than the continental clouds. At low temperatures, the variation of $\tau$ (or LWP) with $T$ is likely to be positive whereas it tends to be negative at high temperatures. The different variations of cloud thickness with $T$ can largely explain the different signs of the variation of $\tau$ with $T$ (Del Genio and Wolf 2000; Hu and Stamnes 2000; Lin et al. 2003). For example, Lin et al. (2003) obtained an averaged variation of LWP of 0.03 K$^{-1}$ for arctic clouds using a combination of surface and satellite data due to an increase of cloud thickness. Del Genio and Wolf (2000) obtained negative variations of LWP with $T$ for midlatitude continental low clouds during summer seasons due to a decrease of cloud thickness. These negative variations are consistent with observations of oceanic low clouds (e.g., Tsienouidis et al. 1992). The oceanic observations are also likely to have resulted from transitions between cloud regimes. The optically thin clouds may be more sensitive to the variation of cloud geometric thickness than their optically thick counterparts. Hu and Stamnes (2000) demonstrated that surface temperature increases with $\tau$ when the clouds are very thin but decreases with $\tau$ when solar radiative forcing dominates infrared radiative forcing. Because of the coarse vertical resolutions used in GCMs, the cloud thickness feedback mechanism cannot be simulated owing to the variations in the cloud top and base heights being inadequately captured (Tsienouidis et al. 1998). However, the LWC feedback mechanism can be simulated with the use of prognostic cloud parameterizations (Roeckner et al. 1987; Colman et al. 2001). There are other mechanisms such as the droplet radius feedback, which is even less well understood (e.g., Hu and Stamnes 2000).

The main goal of this study is to investigate the physical mechanisms of the low cloud feedback through cloud-resolving simulations of cloud-radiative equilibrium response to an increase in SST, with an emphasis on the shallow cumulus and overcast stratocumulus regimes. This study aims to isolating cloud feedback mechanisms for two particular cloud regimes. This means that the simulated changes of cloud physical properties with surface warming exclude those resulting from transitions between cloud regimes. This exclusion cannot be achieved in either observational or GCM studies. Previous modeling studies used a variety of parameterizations of low clouds in single-column radiative-convective
equilibrium models, energy balance models, or GCMs (e.g., Somerville and Remer 1984; Roecker et al. 1987; Tselioudis et al. 1998; Hu and Stammes 2000; Colman et al. 2001; Zhang and Bretherton 2008, hereafter ZB08; Caldwell and Bretherton 2009; Wyant et al. 2009). The present approach is similar to the single-column modeling approach except for explicitly resolving low clouds using the large-eddy simulation (LES) framework. A LES integration typically lasts for a few hours to a day. In this study, the LES is integrated for 30 days so that an equilibrium state can be reached in which radiative, cloud, and large-scale subsidence processes are balanced. To the authors’ knowledge, this study represents the first attempt to perform long-term cloud-radiative equilibrium integrations of low clouds with a 3D LES even though cloud-resolving models (CRMs) have been used to simulate cloud-radiative equilibrium of deep convective clouds with or without imposed large-scale forcings (e.g., Sui et al. 1994; Grabowski et al. 1996; Tompkins and Craig 1999; Xu and Randall 1999). The results of this study will thus be compared with those obtained with models using parameterized low clouds (e.g., ZB08) owing to the lack of similar CRM studies. Similarities to deep convective clouds will be compared to CRM and mesoscale modeling studies—in particular, Tompkins and Craig (1999), who studied the sensitivity of convective–radiative equilibrium to changes in SST, and Larson and Hartmann (2003), who performed a study similar to that of Tompkins and Craig except with a mesoscale model over a large domain.

2. Model and experimental designs

a. The UCLA LES

The University of California, Los Angeles (UCLA) LES was developed by B. Stevens and collaborators (Stevens et al. 1999, 2005; Savic-Jovcic and Stevens 2008; Stevens and Seifert 2008) based on a cloud and mesoscale model originally developed at Colorado State University by W. Cotton’s group. The model solves the Ogura–Phillips anelastic equations (Ogura and Phillips 1962) using finite differences on a mesh with fixed grid spacing in the horizontal and stretched grid spacing in the vertical. Centered differences in space and time are used to represent momentum advection on the three components of the velocity vector, and monotone-centered, slope-limited upwinding is used for the advection of scalars. The scalars include the liquid water temperature, $T_L = T \left[ 1 - \exp \left( L \frac{q_t}{c_p T} \right) \right]$, defined by Betts (1973); the total water mixing ratio $q_t$; the mass mixing ratio of rainwater $q_r$; and the mass specific number of rainwater drops $n_r$. The cloud water mixing ratio $q_w$ is diagnosed from $q_w = \max(0, q_t - q_s)$, where $q_s$ is the saturation mixing ratio. Details of the cloud and microphysical processes are discussed in Stevens and Seifert (2008) and Savic-Jovcic and Stevens (2008). The Smagorinsky (1963) subgrid model is used to represent subgrid fluxes of both scalars and momentum. The radiative parameterization scheme from the Community Atmospheric Model version 3 (CAM3) (Collins et al. 2006) is employed to represent radiative transfer processes in the LES, which is also used to derive the advective forcings imposed on the simulations (ZB08).

Neither the 3D radiative effect nor the diurnal cycle of solar radiation is considered in the simulations. A cosine solar zenith angle of 0.299 is used to represent the diurnal average corresponding to solar insolation at 20°N at the equinox. The formulation of shortwave cloud optical properties follows that of Slingo (1989), with LWP and specified $r_c$ of 14 μm (i.e., the default value in CAM3) as inputs. The fixed $r_c$ eliminates the droplet radius feedback mechanism in all simulations. The cloud emissivity is only dependent on LWP.

Each of the numerical simulations performed in this study is 30 days long, using a time step of 1 s. Because of such a long integration time and the large number of simulations performed, we chose horizontal and vertical grid spacings that are larger than those typically used in short-term LES integrations. In this study, the horizontal domain size is 6 km × 6 km, with a horizontal grid spacing of 200 m × 200 m. There are 93 layers in the vertical, with stretched grid spacing that is gently stretched with height. For instance, the vertical spacing is 30 m near the surface, 39 m at 500 m, 57 m at 1 km, 91 m at 2 km, and 123 m at 3 km (where the highest clouds are simulated in a pair of experiments presented in section 3). The grid spacing is 350 m at the model top. The top of the domain is at 13 km, and a sponge layer occupies the upper five levels. As noted in Savic-Jovcic and Stevens (2008), the coarse resolution and small domain size may impact the simulation of mesoscale organization, which warrants further study by itself.

b. Experimental designs

The atmosphere to be represented in the simulations imitates the large-scale environmental conditions of low clouds in the eastern Pacific. The details of experimental designs are given in ZB08. Though not explicitly stated in ZB08, a two-box representation of the tropical and subtropical dynamics is adopted, as in Larson et al. (1999), Kelly and Randall (2001), and Caldwell and...
Bretherton (2009). One box represents the convecting, warm SST, high humidity region of the tropics, similar to the mean Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) (Webster and Lukas 1992) condition, and the other box represents the subtropical subsidence region with low humidity, boundary layer clouds, and cooler SST (see Fig. 2 in ZB08). Only the latter, the cool box, is simulated in this study. Interactions between the two boxes are through the prescribed subsidence over the cool box. In the configuration outlined in ZB08, the sum of subsidence-induced warming and horizontal advective heat tendency is balanced by the radiative cooling, which changes with SST, while the subsidence drying is canceled by horizontal advective moistening in the free atmosphere of both the control and perturbed sets of simulations, to be described below. The perturbed simulations represent a warmer climate, compared to the present climate represented in the control simulations.

The initial temperature and water vapor mixing profiles are prescribed in the following manner, which is also used as input for calculating the radiative cooling rate that determines the prescribed subsidence and advective forcings. In the control set of simulations, the SST in the warm box is specified to be 28°C and the surface air temperature is assumed to be 1°C colder than the SST. The temperature in the free atmosphere is determined by adiabatically lifting a surface air parcel with a temperature of 27°C and a relative humidity of 75% at 1000 hPa. In the perturbed set of simulations with the identical SST increase in both the warm and cool boxes, which mimics a warmer climate simulated in global models (ZB08), both SST and surface air temperature are raised by 2°C and the temperature in the free atmosphere is similarly obtained. Over the cool box, the temperature in the free atmosphere is assumed to be the same as that over the warm box (under a weak temperature gradient assumption) while the relative humidity is assumed to be 15% everywhere in either the control or perturbed set. The temperature and moisture profiles over the cool box—in addition to other inputs such as SST, surface albedo, ozone profiles, and solar zenith angle—are used to calculate the radiative cooling rate with the CAM3 radiative transfer model (Collins et al. 2006). No clouds were prescribed for this calculation.

As detailed in ZB08, the chosen subsidence profile, in terms of pressure coordinate vertical velocity, has a peak at 800 hPa (Fig. 1a) and the vertically integrated subsidence warming balances that of the net radiative cooling (owing to the rapid decrease of air density with height, geometric vertical velocity $w$ peaks around 10 km instead; Fig. 1b). As stated earlier, the horizontal advective heat tendency is the residual of the combination of the subsidence warming and net radiative cooling (Fig. 1c), and the horizontal advective moisture tendency is assumed to balance the subsidence drying (Fig. 1d). The wind profiles used in all simulations are identical to those shown in Fig. 3b of ZB08. In the perturbed simulation of a given $\Delta$SST (defined as the difference between the warm and cool boxes) configuration, the net radiative cooling is rather similar to that of its corresponding control simulation, but the large-scale subsidence is slightly weaker above 1.3 km (Fig. 1b), due mainly to the more stable stratification.

With the control (SST = 28°C in the warm box) and perturbed (SST = 30°C) configurations described above, six pairs of simulations are performed with $\Delta$SSTs assigned to be 4, 6, 8, 10, 12, and 14 K, respectively, which crudely represent different distances from the warm pool in the western Pacific toward the cold pool over the eastern oceans. Differences in the imposed large-scale forcings among the six simulations (Fig. 1) are one of the factors in determining the equilibrium states because the forcings are designed to balance the radiative cooling rates in the free atmosphere. The complicated interactions of boundary layer dynamic and thermodynamic processes with surface and radiative processes in response to imposed SSTs may play important roles, as explained later.

The 30-day integrations presented in this study were not initialized with a horizontally homogeneous initial state, the specification of which was described earlier, in order to speed up the transition to an equilibrium state. A few test simulations of 5-day duration were performed under the control configurations with the specified initial conditions for each $\Delta$SST pair, and random perturbations were added to the potential temperature field near the surface to break the slab symmetry of these homogeneous initial conditions. An exception is that a single test simulation was used in the $\Delta$SST = 12 K and $\Delta$SST = 14 K pairs of simulations. The actual 30-day integrations using their corresponding forcing data started from the end of the test simulations when clouds were produced near the top of the mixed layer. The imposed subsidence and the interactive cloud top cooling help to strengthen the capping inversion in all simulations performed in this study, which is not prescribed in the initial soundings.

3. Results

The discussion of the results will be presented in three parts. The temporal evolution toward the equilibrium states, the properties of the equilibrium states, and the analysis of cloud feedback mechanisms will be discussed next.
FIG. 1. Vertical profiles of (a) pressure coordinate vertical velocity, (b) geometric vertical velocity, (c) horizontal temperature advective tendency, and (d) horizontal water vapor advective tendency imposed on the control (dashed) and perturbed simulations (solid) for ΔSSTs of 4, 6, 12, and 14 K. Those for ΔSSTs of 8 and 10 K are not shown, for the sake of clarity.
a. Temporal evolution toward the equilibrium states

All simulations with different ΔSSTs and forcings for either the control or the perturbed configurations reach quasi-steady states, or equilibrium states, after 10 to 20 days of integration. The amount of time needed to reach the equilibrium states is different among the different ΔSSTs and depends on the initial conditions obtained from the test simulations. On the other hand, the equilibrium states are not dependent on the initial conditions used. To further illustrate these two points, Figs. 2 and 3 show the time–height cross sections of domain-averaged cloud amount (or cloud fraction) for 30 days for six pairs of simulations described in section 2b.

For the pairs of simulations with ΔSST = 4 K and ΔSST = 6 K (Figs. 2a,b,d,e), shallow cumuli with cloud amount less than 10% (typically 5%) are simulated and the equilibrium states are reached very quickly (~5 days) for the control configurations. The simulated cumulus clouds in the ΔSST = 4 K and ΔSST = 6 K simulations are characterized by large temporal fluctuations with relatively deep penetrating clouds occurring less frequently. The infrequent occurrence of these cumulus clouds may be due to the small domain size and the coarse grid resolution used in this study. The cloud fraction, however, remains nearly constant with time in the lowest 200-m region of the cloud layer. These thin clouds are associated with large entrainment. These features are similar to previous Barbados Oceanographic and Meteorological Experiment (BOMEX) LES simulations (Jiang and Cotton 2000; Siebesma et al. 2003; Cheng and Xu 2008).

Another feature appearing in these four simulations is that there are some significant differences in the simulated vertical extent, cloud-top height, cloud-base height, and the cloud amount within the highest 200-m region of the cloud layer (Figs. 2a,b,d,e). Both the vertical extent of the entire cloud layer and the cloud fraction in the highest 200-m region are larger for the perturbed simulations than for the control simulations. The cloud tops reach 1200–1300 m in both control simulations but the cloud
tops in the perturbed configuration reach 2200 m and 1700 m in the ΔSST = 4 K and 6 K simulations, respectively. These differences are, as shown later, related to higher surface fluxes and increases in the buoyancy of subcloud layer air parcels as the SST increases. As ΔSST changes from 4 to 6 K, the cloud base height is lowered by about 250 m. This is due to a decrease of the lifting condensation level related to more moist surface air parcels in the ΔSST = 6 K simulations.

In contrast to the four simulations of cumulus regimes in which cloud bases do not rise with time during the adjustment phase (Figs. 2a,b,d,e), both the cloud base and top ascend at nearly the same pace with time, albeit somewhat unevenly (i.e., more variation at the cloud base) in the four pairs of simulations with higher ΔSSTs (Figs. 2c, 2f, and 3). This difference is related to the transition of cloud regimes from cumulus to stratocumulus (or stratiform cloud). A transitional cloud regime appears intermittently for the ΔSST = 8 K simulations in which a shallow cumulus layer is overlaid by a higher stratiform cloud layer (with maximum cloud fraction close to 100%). A transitional cloud regime similar to that observed in the Atlantic Trade Wind Experiment (ATEX) (Stevens et al. 2001), which has a much smaller gap between the two layers, is not produced under either the control or perturbed configurations. The large gaps between the two cloud layers are probably unrealistic, arising from the high cloud tops of the upper layer reaching between 2500 and 3000 m. This result suggests that there may be potential deficiencies in the experiment design related to large-scale forcings; for example, the horizontal advective tendencies of heat and moisture were designed to offset the subsidence-induced warming and drying (ω is related to clear-sky radiative cooling), respectively. The domain-averaged radiative cooling rates at cloud top are much higher for

1 An ATEX–like transitional cloud regime was simulated with the UCLA LES using more realistic advective forcing data designed for the Cloud Feedback Model Intercomparison Project by one of the coauthors (Zhang) and the same experimental configuration used in this study.
stratocumulus than for cumulus clouds. The resulting imbalance between the simulated radiative cooling and subsidence contributes to the steady rise of cloud-top heights, coupled with overly efficient cloud-top entrainment related to coarse vertical resolutions (Bretherton et al. 1999). Thus, the results from these two pairs of simulations will not be discussed further.

Under the perturbed configuration, the cloud tops from all ∆SST simulations are higher than those under the control configuration by 200–1000 m. However, the cloud-base heights are also higher by ~200 m for the simulations with ∆SSTs between 8 and 14 K (see Figs. 2e, 2f, and 3). The differences in cloud-top heights in the pairs of ∆SST = 4 K and ∆SST = 6 K simulations are especially pronounced (500–1000 m). The increased turbulent moisture transports related to higher SSTs are directly linked to these differences between the control and perturbed configurations. This will be further explained below. This result is reminiscent of deep convective–radiative equilibrium simulated by Tompkins and Craig (1999) in the sense that cloud profiles were shifted upward at higher SSTs. However, they also found a small reduction in cloud fraction, which did not occur for shallow cumulus regimes simulated in this study.

b. Equilibrium states

Because the amount of time needed to reach the equilibrium states varies from one simulation to another (up to 20 days), the averages between 20 and 30 days are chosen to represent the equilibrium states for all simulations. Their thermodynamics, cloud properties, and vertical transport profiles are discussed below. Table 1 provides a summary of selected 1D (surface, top of the atmosphere, and vertically integrated) characteristics of the simulated equilibrium states.

1) THERMODYNAMIC PROFILES

Vertical profiles of liquid water potential temperature and total water mixing ratio are shown for four pairs of simulations (with ∆SST = 4 K, 6 K, 12 K, and 14 K) in Figs. 4a,b, 5a,b, 6a,b, and 7a,b, respectively. The most distinct feature is the existence of well-mixed boundary layers capped by strong inversions for the stratocumulus regimes (∆SST = 12 K and 14 K), consistent with typical structures of the stratocumulus-topped boundary layer (e.g., Moeng 2000). In contrast, the entire boundary layers of the ∆SST = 4 K and ∆SST = 6 K simulations are not well mixed. These thermodynamic profiles are similar to those of BOMEX cumulus simulations (e.g., Jiang and Cotton 2000; Siebesmas et al. 2003; Cheng and Xu 2008) except that the inversion heights studied here vary greatly from one simulation to another (Figs. 4a,b and 5a,b).

<table>
<thead>
<tr>
<th>Variables</th>
<th>Expt</th>
<th>∆SST</th>
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<tbody>
<tr>
<td></td>
<td></td>
<td>4 K</td>
</tr>
<tr>
<td>LW (W m⁻²)</td>
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<td></td>
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<tr>
<td></td>
<td>∆ + 2 K</td>
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</tr>
<tr>
<td>SW (W m⁻²)</td>
<td>CTRL</td>
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</tr>
<tr>
<td></td>
<td>∆ + 2 K</td>
<td>−4.5</td>
</tr>
<tr>
<td>LW (W m⁻²)</td>
<td>CTRL</td>
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<tr>
<td></td>
<td>∆ + 2 K</td>
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<tr>
<td>Cloud fraction (%)</td>
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<tr>
<td></td>
<td>∆ + 2 K</td>
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<tr>
<td>PW (kg m⁻²)</td>
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</tr>
<tr>
<td></td>
<td>∆ + 2 K</td>
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<tr>
<td>LWP (g m⁻²)</td>
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<tr>
<td></td>
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<tr>
<td>Zι (m)</td>
<td>CTRL</td>
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<td>901</td>
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<td>Cloud thickness (m)</td>
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<td></td>
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<tr>
<td></td>
<td>∆ + 2 K</td>
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</table>

The effects of the perturbed SST increase on the thermodynamic profiles are threefold. First, both θl and qt values are nearly uniformly increased at any level within the boundary layer below the capping inversion layer of the control simulations. Second, the capping inversion layers (see also Table 1) are raised. Third, the amplitude of the upward jump of the capping inversion increases as ∆SST decreases, which results in much larger increases of qt in the boundary layer of the perturbed simulations of the cumulus regime, compared to those of the stratocumulus regime. The large turbulent transports of moisture associated with the cumulus regime (Figs. 4e,f and 5e,f), as discussed later, are the major reason, although the upward movement of the peak θl transports in the cloud layer (Figs. 4e, 5e, 6e, and 7e) is tied to the changes in the θl profiles (Figs. 4a, 5a, 6a, and 7a). Slightly weaker large-scale subsidence, as discussed earlier, may play a role in the upward jump of the capping inversion. The change in the inversion height is expected to impact the cloud properties presented below, along with the significant increase of water vapor in the boundary layer, which
FIG. 4. Vertical profiles of (a) liquid-water potential temperature, (b) total water mixing ratio, (c) cloud water mixing ratio, and (d) cloud fraction, and vertical fluxes of (e) liquid-water potential temperature and (f) total water mixing ratio for the simulations with ΔSST, between the warm and cool boxes, of 4 K under the control (solid) and perturbed (dashed) configurations.
FIG. 5. As in Fig. 4, but for ΔSST of 6 K.
FIG. 6. As in Fig. 4, but for ΔSST of 12 K.
FIG. 7. As in Fig. 4, but for ΔSST of 14 K.
can increase clear-sky cooling (≈0.2–0.7 K day**−1** in the ΔSST = 6 K and ΔSST = 4 K pairs, not shown). This cooling is the key component for the hypothesized physical mechanism presented in Wyant et al. (2009). This will be further discussed later.

2) CLOUD PROPERTY PROFILES

Vertical profiles of cloud water mixing ratio and cloud fraction are shown in Figs. 4c,d, 5c,d, 6c,d, and 7c,d. For the stratocumulus simulations, clouds are formed in a thin layer located below the capping inversion with a thickness between 400 and 600 m (Figs. 6c,d and 7c,d). The maximum LWC is located near the cloud top with a maximum cloud fraction of 100%. The cloud thickness of the cumulus simulations, however, varies between 1000 and 2000 m (Figs. 4c,d and 5c,d). The LWC increases with height, but cloud fraction has two peaks in these simulations. The primary peak is located at the cloud base and the secondary peak is at the cloud top. The secondary peaks are related likely to interactive radiative destabilization, which is absent in the BOMEX simulations with prescribed radiation (Siebesma et al. 2003). It is also interesting to note that the maximum cloud water mixing ratio varies by two orders of magnitude (from 0.003 to 0.6 g kg**−1**) between the cumulus and stratocumulus simulations. The small magnitudes in cloud water mixing ratio are related to the small cloud fractions in the cumulus simulations (Figs. 4d and 5d).

The effects of the perturbed SST increase on the cloud property profiles are exhibited by (i) the upward jump of the cloud layers in all pairs of simulations (as discussed in section 3a), (ii) the increase of the maximum LWC, (iii) the increase/decrease of cloud fraction at different heights in the cumulus regime only, and (iv) the increase of precipitation in the cumulus simulations only (not shown). The increase in the maximum LWC (Figs. 4c, 5c, 6c, and 7c) is consistent with the changes in the thermodynamic profiles with the SST increase discussed earlier because the moister atmosphere under the perturbed configuration can hold more condensed water than in the control configuration (Table 1) if precipitation is negligible. In the stratocumulus simulations, there is an ≈10% increase in the maximum cloud water mixing ratio whereas the increase is 25%–50% in the pairs of cumulus simulations. Both are much larger than changes due to adiabatic increase of LWC (Betts and Harshvardhan 1987). These large maximum LWCs are associated with higher cloud tops (Figs. 4c, 5c, 6c, and 7c).

In the perturbed set of stratocumulus simulations, the maximum cloud fractions are unchanged at 100% (Figs. 6d and 7d) whereas those of the cumulus simulations at the cloud base (cloud top) decrease (increase) with the perturbed SST. Although the cloud fraction decreases in the lower regions of the cloud layers in the pairs of cumulus simulations (Fig. 5d), the cloud layers extend upward with the maximum cloud fractions in the upper regions, being as large as the primary maximum near the cloud base. The results for the cumulus regime may suggest that updrafts are stronger in the perturbed configuration (Table 1), which can produce stronger compensating subsidence that reduces the cloudy area in the lower cloud layer. This explanation was also given in Tompkins and Craig (1999) and Eitzen and Xu (2008) for deep convective-equilibrium simulations. The increased detrainment is also responsible for both the increase in cloud fraction and cloud water mixing ratio in the upper region of the cloud layer. Similar features also appear in Blossey et al. (2009) except for smaller magnitudes owing to a slightly different configuration and the use of a 2D CRM. An estimate of entrainment rate based on the surface latent heat flux and the moisture jump across the inversion of the equilibrium states suggests a 15%–20% increase in the cumulus simulations but no change in the stratocumulus simulations.

Caldwell and Bretherton (2009) found from a simple mixed-layer model that the equilibrium boundary layer depth and cloud thickness increase in the perturbed configuration, which is consistent with the present study. They attributed their result to the decrease of large-scale subsidence and entrainment in the more stable, warmer environment. ZB08 found that stratiform cloud amount is slightly increased in the perturbed configuration but neither the boundary layer depth nor the cloud thickness is changed. This might be related to the coarser vertical resolution used in their simulation and, perhaps, to deficiencies in the parameterizations of CAM3, which did not produce the cumulus regime for the same ΔSST = 4 K setup as in the present study.

3) VERTICAL TRANSPORTS

It is well known that vertical transports of liquid-water potential temperature (wθl) and total water mixing ratio (wql) are distinctly different for cumulus (Figs. 4e,f and 5e,f) and stratocumulus (Figs. 6e,f and 7e,f) regimes. As in Siebesma et al. (2003), the wql profiles of the ΔSST = 4 K and ΔSST = 6 K simulations can be subdivided into two regions (Figs. 4f and 5f). Between the surface up to near the inversion, the fluxes only marginally decrease with height. This is followed by a strong decrease in the inversion where most of the moisture surface flux is deposited and used to moisten the cloud layer. The perturbed simulations have larger magnitudes of these fluxes (but similarly decrease slowly with height) in the lower regions and slightly stronger decreases in the upper regions, compared to the control
The inversion. Here $q_L$ is due to the potential temperature $\theta$, defined as $\theta = T / (g / C_p)$. As explained in Siebesma et al. (2003), this minimum corresponds to a level where the condensation rate equals the evaporation rate. Above this level, evaporation exceeds condensation owing to entrainment of drier air from above the inversion.

The positive $w\theta'$ value of the stratocumulus simulations below the cloud layer is influenced by subcloud-layer dynamics because of the strong coupling between the subcloud and cloud layers. The convergence of $w\theta'$ is balanced by the cloud-top radiative cooling ($\sim 4$ K h$^{-1}$) and cloud-base radiative warming ($\sim 0.7$ K h$^{-1}$). The cloud-base radiative warming is strong in all stratocumulus simulations but weak in the cumulus simulations (not shown) due to small cloud fractions (Figs. 4d and 5d).

c. Cloud–climate feedback analysis

To understand what contributes to negative low-cloud feedback, we relate the rates of change of cloud physical properties to the strengths of climate feedback. Two types of climate feedback effects are analyzed. One is the clear-sky feedback effect, which includes the effects of water vapor, temperature, and lapse rate feedbacks. This is defined as

$$\alpha_{\text{clr}} = -\frac{\Delta G_{\text{clr}}}{\Delta T_s}$$

where $\Delta T_s$ is the change of surface temperature and $\Delta G_{\text{clr}}$ is the change of clear-sky net upward radiative fluxes at the TOA (Table 2). (The clear-sky radiative fluxes shown in Table 2 are obtained from offline calculations.) The net upward radiative flux $G$ at the TOA is defined as

$$G = LW - SW,$$

where LW is the emitted infrared flux and SW is the net downward shortwave flux. The other is the cloud feedback effect, defined as

$$\alpha_{\text{cld}} = \frac{\Delta \text{CRF}}{\Delta T_s} = \frac{\Delta (G_{\text{clr}} - G_{\text{all}})}{\Delta T_s},$$

where CRF stands for cloud radiative forcing, which is the difference between clear-sky (“clr”) and all-sky (“all”) fluxes. The LW and SW components of the CRF are defined in such a way that the net CRF is the sum of the LW and SW CRFs; that is,

$$\text{CRF}_{\text{LW}} = LW_{\text{clr}} - LW_{\text{all}}$$

and

$$\text{CRF}_{\text{SW}} = SW_{\text{all}} - SW_{\text{clr}}.$$

As stated earlier, $\Delta T_s$ is specified to be 2 K, that is, the difference in SST between the perturbed and control simulations, and $\Delta G$ is the radiative response at the TOA to 2 K change in SST. They are related to the climate sensitivity parameter $\lambda$, defined as $\lambda = -\Delta G / \Delta T_s$, where the negative sign is related to the definition of $G$. Similar to the GCM studies of Cess et al. (1990) and Zhang et al. (1994) and many others, the CRM studies of Lau et al. (1994) and Tompkins and Craig (1999), and the mesoscale modeling study of Larson and Hartmann (2003), this study represents a reversed climate response problem since the atmosphere reaches a new equilibrium state with a specified SST change. The climate sensitivity parameter $\lambda$ is related to clear-sky

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**Table 2. Differences in clear-sky upward TOA longwave, net downward shortwave radiative flux, and net (SW – LW) radiative flux (W m$^{-2}$) for the equilibrium states between the perturbed and control simulations; ΔSSTs as in Table 1. Clear-sky feedback effects are the flux differences divided by 2 K.**

<table>
<thead>
<tr>
<th>ΔSST</th>
<th>LW</th>
<th>SW</th>
<th>Net</th>
</tr>
</thead>
<tbody>
<tr>
<td>4 K</td>
<td>0.22</td>
<td>0.91</td>
<td>0.69</td>
</tr>
<tr>
<td>6 K</td>
<td>3.59</td>
<td>0.68</td>
<td>-2.90</td>
</tr>
<tr>
<td>12 K</td>
<td>2.96</td>
<td>0.44</td>
<td>-2.52</td>
</tr>
<tr>
<td>14 K</td>
<td>3.11</td>
<td>0.47</td>
<td>-2.64</td>
</tr>
</tbody>
</table>

---
and cloud feedback effects through $\lambda = (\alpha_{clr} + \alpha_{cld})^{-1}$ in the analysis presented below. An offline calculation is performed to partition the effect of clear-sky feedbacks into that due to change in either water vapor ($\alpha_q$) or temperature ($\alpha_T$). The latter also includes the effect of the changing lapse rates, which are mostly related to the vertical jump of the inversion layer in this study.

In contrast to the cloud feedback effect, the positive clear-sky feedback effect is well understood and is a robust feature of GCMs and a range of models of different complexity and scope (Randall et al. 2007). The strong positive water vapor feedback in the GCM arises from changes in the upper tropospheric humidity, while the negative lapse rate feedback is due to changes in the temperature profile as SST increases. In addition, the increase of the mean temperature (the Stefan–Boltzman emission) as SST increases also produces a negative feedback. For the tropical deep convective simulations, Tompkins and Craig (1999) and Larson and Hartmann (2003) found that the clear-sky feedback effect is negative (−1.44 and $-2.01 \text{ W m}^{-2} \text{ K}^{-1}$, respectively) and mostly determined by the lapse rate feedback, since the effect of the mean temperature feedback is nearly cancelled by that of the water vapor feedback.

In this study, the clear-sky feedback effect (Table 3) is negative (−1.26 to $-1.45 \text{ W m}^{-2} \text{ K}^{-1}$) for the 6, 12, and 14 K pairs, but positive for the 4 K pair (0.35 W m$^{-2}$ K$^{-1}$). This result may be due partly to the fact that most changes in water vapor occurred in the lower troposphere instead of the upper troposphere (Figs. 4b, 5b, 6b, and 7b). There are larger differences (−1.80 to $-0.11 \text{ W m}^{-2} \text{ K}^{-1}$) in the LW clear-sky feedback effects than for the SW counterparts (0.22 to 0.46 W m$^{-2}$ K$^{-1}$), especially between the two pairs of cumulus simulations. This suggests that contrasting changes in both water vapor and lapse rate determine the sign and magnitude of clear-sky feedbacks. The rate of increase of precipitable water with SST are similar (9.5%–10.3% K$^{-1}$, Table 4) for the 6, 12, and 14 K pairs; the positive clear-sky SW feedbacks (0.22 to 0.34 W m$^{-2}$ K; Table 2) and clear-sky wa-

<table>
<thead>
<tr>
<th>$\Delta$SST</th>
<th>4 K</th>
<th>6 K</th>
<th>12 K</th>
<th>14 K</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha_q$</td>
<td>4.74</td>
<td>3.79</td>
<td>4.19</td>
<td>4.16</td>
</tr>
<tr>
<td>$\alpha_T$</td>
<td>−4.40</td>
<td>−5.24</td>
<td>−5.45</td>
<td>−5.47</td>
</tr>
<tr>
<td>$\alpha_{clr}$</td>
<td>0.35</td>
<td>−1.45</td>
<td>−1.26</td>
<td>−1.32</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>4 K</th>
<th>6 K</th>
<th>12 K</th>
<th>14 K</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saturation water vapor pressure</td>
<td>6.0</td>
<td>6.0</td>
<td>6.3</td>
</tr>
<tr>
<td>Surface latent heat flux</td>
<td>10.9</td>
<td>13.2</td>
<td>5.7</td>
</tr>
<tr>
<td>Precipitable water</td>
<td>16.7</td>
<td>10.3</td>
<td>9.5</td>
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<tr>
<td>Liquid water path</td>
<td>54.5</td>
<td>24.6</td>
<td>16.1</td>
</tr>
<tr>
<td>Maximum cloud water mixing ratio</td>
<td>50.9</td>
<td>24.3</td>
<td>10.0</td>
</tr>
<tr>
<td>Inversion height</td>
<td>21.7</td>
<td>14.2</td>
<td>5.0</td>
</tr>
<tr>
<td>Cloud thickness</td>
<td>25.0</td>
<td>13.2</td>
<td>10.1</td>
</tr>
</tbody>
</table>

The LW and SW CRFs for the control set of simulations are listed in Table 5. As expected, the SW CRFs are negative while the LW CRFs are positive. The magnitudes of SW CRFs (−4.6 to $-132.2 \text{ W m}^{-2}$) are much larger than those of the LW CRFs (0.4 to 20.3 W m$^{-2}$) for all simulations and higher for the stratuscumulus regime than for the cumulus regime by almost two orders of magnitude, consistent with observations (e.g., Etten et al. 2008). The absolute values of both CRFs are well correlated with cloud optical depth (τ), with the highest values in the $\Delta$SST = 12 K simulation. Because of low cloud fractions, the CRFs are small for the $\Delta$SST = 4 K and 6 K simulations.
The magnitudes of SW (ΔCRF_{SW}) and net CRFs (ΔCRF_{net}) systematically increase owing to the +2 K SST from the ΔSST = 6 K to 14 K pairs of simulations (1.4 to 11.0 W m\(^{-2}\)) despite the fact that Δτ does not systematically increase with ΔSST (Table 5). For example, Δτ is largest for the ΔSST = 12 K pair, but the magnitudes of ΔCRF_{SW} and ΔCRF_{net} are largest for the ΔSST = 14 K pair. The reason is that solar absorption increases with lnτ (cf. Δlnτ between the two pairs in Table 5). The magnitudes of ΔCRF_{SW} and ΔCRF_{net} are proportionally larger for the ΔSST = 4 K pair than for the ΔSST = 6 K pair with respect to Δlnτ, due to a higher percentage of the increases of both LWC and cloud geometric thickness with SST (Table 4). Note that the cloud feedback effect α_{cl} is approximately 0.68, 2.22, 4.21, and 5.22 W m\(^{-2}\) K\(^{-1}\) for the 6, 4, 12, and 14 K pairs, respectively. The magnitudes of α_{cl} are, as expected, determined mostly by the SW cloud feedbacks (Table 5).

In the following, the relationships of ΔCRF_{net} with other cloud variables will be discussed. Changes in a variety of cloud physical and radiative properties due to +2 K SST (e.g., surface latent heat flux, precipitable water, LWP, inversion height, and cloud thickness) basically increase as ΔSST decreases (Table 1). The rates of change (dlnA/dT) for a few selected properties are compared to the rate of change of saturation water vapor pressure (dln_{e} /dT, Table 4). The latter varies between 6.0% and 6.4% K\(^{-1}\) for the ΔSST = 4 K to 14 K pairs of simulations. Table 4 shows that precipitable water, LWP, and maximum LWC vary with significantly higher rates than dln_{e} /dT in all four pairs of simulations, particularly in the cumulus simulations. The surface latent heat fluxes vary at approximately twice the rates as dln_{e} /dT for the cumulus simulations but at the same rates for the stratocumulus simulations. This difference is due to strong dynamic circulations associated with cumulus clouds, which bring drier air to the surface to enhance surface evaporation. The inversion height and cloud thickness also vary similarly at roughly the same rates as the surface latent heat flux except for the ΔSST = 4 K pair, whose rates are twice as high as those of the ΔSST = 6 K pair.

The primary drivers for these increases of cloud physical and radiative properties with +2 K SST are the surface latent fluxes, which increase the turbulent transports in the boundary layer (Figs. 4e, 5e, 6e, and 7e). The fact that the cloud layers become higher and thicker (Figs. 4d, 5d, 6d, and 7d) also suggests that the decrease of large-scale subsidence in the perturbed simulations (Fig. 1) may play a role, as discussed in Caldwell and Bretherton (2009). The increases in water vapor and LWC enhance both clear-sky (Table 2) and cloudy-sky LW radiative cooling (Table 1). The increased solar reflection resulting from optically thicker clouds would inhibit further warming at the surface (if an interactive ocean were included), creating a negative low-cloud feedback. The associated changes in cloud physical properties shown in Table 4 reflect this negative feedback, although the magnitudes of the cloud feedback effects are highly dependent on the cloud types simulated. Although the results presented in this study resemble those shown in Wyant et al. (2009), especially for the cumulus regime, the physical mechanisms operating in most of these simulations are slightly different—namely, the turbulent moisture transports in the boundary layer are emphasized while the increases in cloud fraction and convection are probably less important, particularly for the stratocumulus regime which is overcast in this study.

Finally, the sign of the climate sensitivity parameter λ is negative and is determined by that of cloud feedback effects in all four ΔSSTs because the magnitudes of cloud feedback effects are larger than those of clear-sky feedback effects except for the 6 K pair (Tables 3 and 5). For the ΔSST = 4 K pair, the positive clear-sky feedback effect is weak, compared to the negative cloud feedback effect. The clear-sky feedback effect is negative in the ΔSST = 6 K, 12 K, and 14 K pairs of simulations. These differences in feedback strengths among the four pairs of simulations, which decrease with increasing ΔSSTs \([-0.56, -0.45, -0.18, \text{ and } -0.15 \text{ K (W m}^{-2}\text{)}^{-1}\)], in turn explain why the rates of change in the cloud parameters shown in Table 4 are much higher for the cumulus simulations than for the stratocumulus simulations.

4. Conclusions and discussion

This study has investigated the physical mechanisms of the low cloud feedback through cloud-resolving simulations of cloud-radiative equilibrium response to a
+2 K SST. Our experimental design has been to isolate cloud feedback mechanisms for cumulus and stratocumulus regimes, which means that the simulated changes of cloud physical properties with surface warming exclude those resulting from transitions between cloud regimes. Six pairs of perturbed and control simulations were performed using the SST differences ($\Delta$SST) in the range from 4 to 14 K between the warm tropical and cold subtropical regions. In the perturbed set of simulations, SSTs were increased by 2 K in both the warm and cold regions from those in the control set of simulations. Large-scale advective heat and moisture tendencies in the cool box are derived so that the clear-sky radiative cooling is balanced by large-scale subsidence (and horizontal advection) in the free atmosphere above the boundary layer. The boundary layer clouds respond to both the imposed large-scale forcing and the surface turbulent fluxes resulting from the perturbation of SST. The latter were interactively calculated from the bulk formula for each simulation with a prescribed SST. Quasi-steady states were reached after 10 to 20 days of integrations.

With $\Delta$SST = 4 K and $\Delta$SST = 6 K pairs, shallow cumuli with cloud amount typically less than 5% were simulated while stratocumulus clouds with cloud amount close to 100% were simulated with the higher $\Delta$SST pairs. Equilibrium states have variable thicknesses and vertical extents for the range of specified $\Delta$SSTs, with the perturbed set of simulations having higher cloud bases and tops and larger geometric thicknesses. The increased turbulent moisture transports in the boundary layer resulting from higher SSTs are primarily responsible for the upward jump of the capping inversion layers since the imposed large-scale subsidence in the perturbed simulations is similar to (but slightly weaker than) the control simulations, although overly efficient entrainment due to coarse resolution cannot be ruled out (e.g., Bretherton et al. 1999).

Negative clear-sky feedback effects were obtained for the 6, 12, and 14 K pairs, while the 4 K pair had weak positive effects. This result may be due partly to the fact that most changes in water vapor occurred in the lower troposphere instead of the upper troposphere. The large increase in precipitable vapor (relatively large positive $\alpha_q$) and the reduced negative lapse rate feedback (relatively small negative $\alpha_T$) may explain why the clear-sky feedback effect is weakly positive in the 4 K pair. The clear-sky feedback effects of the 6, 12, and 14 K pairs are comparable to the negative effects in the tropical deep convective simulations of Tompkins and Craig (1999) and Larson and Hartmann (2003).

As expected, the SW cloud radiative forcings (CRFs) are negative, while the LW CRFs are positive. The magnitudes of SW CRFs are much larger than those of the LW CRFs for all simulations and higher for the stratocumulus regime than the cumulus regime, consistent with observations (e.g., Eitzen et al. 2008). Negative cloud feedback effects were obtained for all four pairs of simulations ($-0.68$ to $-5.22$ W m$^{-2}$ K$^{-1}$), with major contributions from SW ($-0.79$ to $-5.50$ W m$^{-2}$ K$^{-1}$) and opposite effects from LW (0.11 to 0.49 W m$^{-2}$ K$^{-1}$).

The negative cloud feedback effect arises from the increase of cloud geometric thickness, LWC, LWP, $\tau$, and inversion height with the increase of SST. Although this result is basically consistent with the cloud feedback mechanism proposed by Paltridge (1980) and Somerville and Remer (1984), their assumptions were based on the increase of LWC with cloud ambient temperature, not SST. In the cumulus regime, the increase of cloud fraction in the upper regions of the cloud layers with SST may also play a significant role, consistent with the findings in Blossey et al. (2009). However, the increases of LWP and $\tau$ with the increase of SST simulated in this study are just the opposite of those obtained from satellite measurements of oceanic low clouds (Tselioudis et al. 1992; Chang and Coakley 2007). The range of the temperature variations in these oceanic observations is much greater than 2 K used in the present study, but these variations are mostly due to spatial and temporal variabilities, so the transition from stratocumulus to cumulus regimes may be responsible for the decrease of LWP and $\tau$ with temperature. The present study did not simulate a decrease in cloud geometric thickness with the increase of surface temperature, which was observed in continental clouds (DelGenio and Wolf 2000); that is, both the cloud-top and cloud-base heights simulated in this study increase with SST. To produce the decrease of cloud geometric thickness with the increase of SST, the imposed large-scale subsidence in the perturbed simulations has to be stronger to keep the cloud top at the same heights as in the control simulations. This implies that the Hadley circulation should be stronger in the warmer climate. However, a weaker and poleward-expanded Hadley circulation was simulated in a global CRM climate sensitivity study (Miura et al. 2005) and in doubling CO$_2$ simulations of GCMs (e.g., Lu et al. 2008). Wyant et al. (2009) found that the large-scale subsidence at high stability bins (corresponding to stratocumulus regimes) is weakened, but it is strengthened at moderate stability bins (cumulus regimes) in a warmer climate. This implies that the increase of cloud-top heights in the $\Delta$SST = 4 K and $\Delta$SST = 6 K pairs may be overestimated. However, the changes in cloud fraction and the sign of cloud feedback effects are consistent between the two studies.

The physical mechanisms for the negative cloud feedbacks are only slightly different between Wyant et al. (2009) and this study. This is primarily because only the
cumulus regimes were studied in Wyant et al. while the overcast stratocumulus regimes were also simulated in this study. The primary drivers for the increases of cloud physical properties with SST are the surface latent fluxes, which increase the turbulent transports in the boundary layer and move the cloud layers higher and make the cloud layers thicker. The water vapor and LWC are increased. These, in turn, increase both clear-sky and cloudy-sky LW radiative cooling. The increased solar reflection would inhibit further warming at the surface (if an interactive ocean were included), creating a negative low-cloud feedback. The magnitudes of the cloud feedback effects vary greatly among the pairs of simulations (cloud regimes). The clear-sky feedback effects contribute differently to the climate sensitivity for the lowest $\Delta$SST pair, which may be related to the highest rates of change in selected physical properties with $+2 \text{ K SST}$ for the cumulus simulations.

Further work is underway to simulate the transition cloud regimes using more realistic advective forcing data from the Cloud Feedback Model Intercomparison Project (CFMIP) case study so that the gaps in the $\Delta$SST parameter space can be filled. Higher-resolution simulations will be performed also to examine the robustness of the results presented here. It is expected that the equilibrium cloud-top heights of stratocumulus regimes will be slightly lower owing to reduced entrainment at higher resolutions, but we expect that the physical processes described in the paper and the climate sensitivity results will not be greatly affected. These will be the subjects of future research.

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