Trimodal cloudiness and tropical stable layers in simulations of radiative convective equilibrium

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[1] In this paper, we examine the tropical environment at radiative convective equilibrium using a large-domain cloud system resolving numerical model. As in observed studies of convectively active periods over warm tropical oceans (in particular the tropical western Pacific), we find a trimodal cloud structure that is closely associated with the presence of three distinct stable layers, including a prominent stable layer located near the zero-degree Celsius level. In addition, the simulation exhibits three separate large scale zonal overturning circulations, with two of these circulations located above the trade wind inversion and separated by the freezing level stable layer. At equilibrium, latent heat release associated with freezing and melting processes is dwarfed by that of vapor transitions, and simulation results suggest that this stable layer can be maintained by subsidence in the presence of longwave radiative cooling above the zero-degree level. **Citation:** Posselt, D. J., S. C. van den Heever, and G. L. Stephens (2008), Trimodal cloudiness and tropical stable layers in simulations of radiative convective equilibrium, *Geophys. Res. Lett.*, *35*, L08802, doi:10.1029/2007GL033029.

1. Introduction

[2] It has long been recognized that organized deep convective systems play a key role in regulating the large scale circulations and thermal structure of the atmosphere in tropical oceanic regions. An excess of incoming solar radiation and consequent surface heating produces a potentially unstable stratification, which is balanced by troposphere-deep mixing in convection. As such, the tropical atmosphere is never far from radiative convective equilibrium (RCE). Though the large-scale interaction of convection and radiation at RCE is now relatively well understood (see *Stephens* [2005] and *Tao* [2007] for thorough reviews of observational and modeling studies of RCE, respectively), there are a number of characteristics of the tropical environment that have yet to be explained. Two of the most intriguing features of the tropical atmosphere are the stable layer that has been observed near the freezing level [*Johnson et al.*, 1996], and the presence of cloud at three distinct levels outside of deep convective cores; the trade wind inversion level, the freezing level, and the tropopause [*Mapes and Zuidema*, 1996; *Zuidema*, 1998; *Johnson et al.*, 1999; *Haynes and Stephens*, 2007]. Observational studies have suggested that trimode cloudiness in the tropics is associated with midlevel detrainment from deep convection and with the presence of cumulus congestus clouds [*Johnson et al.*, 1999; *Haynes and Stephens*, 2007], while *Mapes and Houze* [1995] and *Johnson et al.* [1996] suggest that the freezing-level stable layer forms as a result of melting of snow in stratiform precipitation regions. However, the connections between thermal stratification, the large scale overturning circulation, and the trimode distribution of cloudiness have not yet been sufficiently described.

[3] In this paper, we study the tropical atmosphere at RCE using a three dimensional cloud resolving model on a large horizontal domain centered on the equator. Our goal is to examine the thermal and moisture structure of the tropical atmosphere at equilibrium, and to characterize the associated circulations and cloud distributions. The remainder of this paper is organized as follows. We describe the cloud resolving model simulation in detail in section 2, then present results from the simulation in section 3. In section 4, we discuss the implications of the model results and delve into a brief examination of the mechanisms that might be responsible for maintaining the structures produced by the model. We conclude in section 5 with a brief summary along with a suggestion of how we will be expanding on the results reported here in future work.

2. Description of RAMS Simulations

[4] The cloud resolving model used in this study is the Regional Atmospheric Modeling System (RAMS), developed at Colorado State University [*Pielke et al.*, 1992; *Cotton et al.*, 2003]. The model includes a detailed bulk cloud microphysical scheme that assumes a gamma-shaped particle size distribution for three species of liquid and five species of ice. The RAMS advection scheme, turbulence, radiation, cloud microphysics, and surface flux parameterizations are documented by *Cotton et al.* [2003], and a detailed description of recent changes to the model microphysics is given by *Saleeby and Cotton* [2004]. To facilitate development of large-scale overturning circulations and realistic interaction of cold pools associated with the evaporation of precipitation [*Tompkins*, 2001], we employ a three dimensional model domain that spans 9600 km in the zonal direction and 180 km in the meridional. Horizontal grid spacing of 2.4 km is used in both zonal and meridional directions, and the vertical grid consists of 38 levels, which stretch gradually from a depth of 20 meters in the boundary layer to just over one kilometer in the upper troposphere. The model is initiated from a vertical profile of temperature and moisture that is representative of convectively active conditions over the Pacific warm pool, with three dimensional wind set equal

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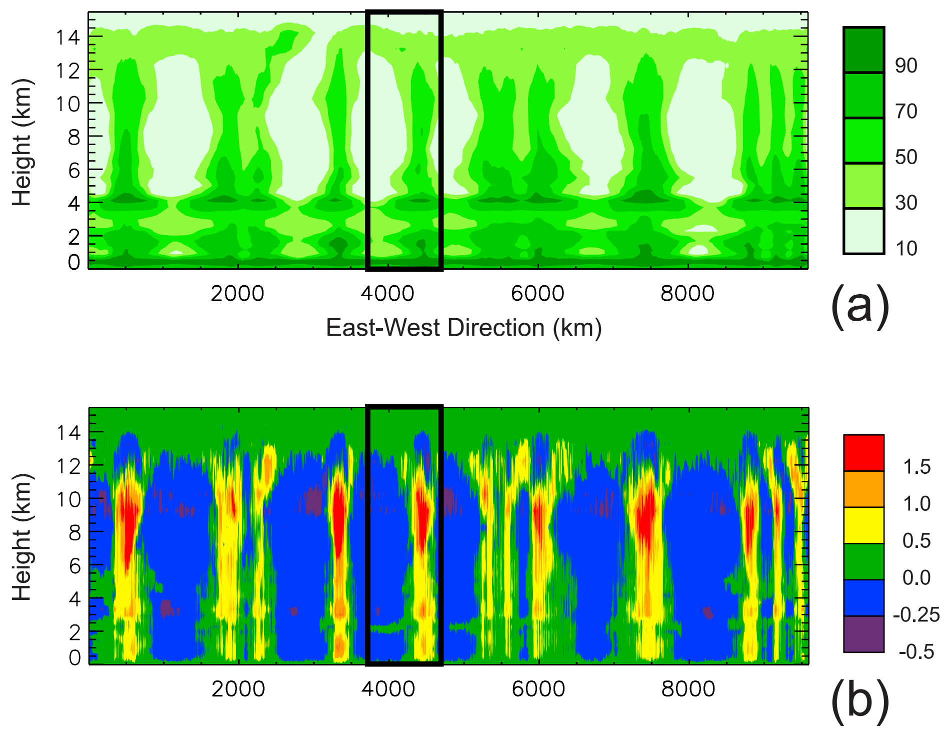

Figure 1. Cross-sections of (a) relative humidity (percent) and (b) vertical velocity (m/s) averaged across the meridional direction and over simulation days 46–55. The box in both Figures 1a and 1b depicts the location of the subset region shown in Figure 2.

to zero. A plot of this sounding is given by *Stephens et al.* [2004], and we note that the stable layers observed in the equilibrium simulation presented below are not present in the initial thermodynamic profile. The model sea surface temperature is fixed at 300 K, and the solar zenith angle is set constant such that the daily insolation is maintained at 447.2 W/m2; roughly equivalent to the mean annual insolation at the equator. To initiate the development of convection, small (less than 0.1 K) temperature perturbations are introduced randomly into the model atmosphere at the initial time in the region below 2 kilometers throughout the model domain. The model attains radiative-convective equilibrium after approximately 40 simulated days (not shown; see *Stephens et al.* [2008] for details).

3. Simulation Results

[5] As the model approaches an equilibrium state, broad regions of organized convection develop, which are separated by regions of clear air. Figure 1 depicts ten-day mean fields of relative humidity and vertical velocity averaged in the meridional direction from the time period that spans days 46 through 55 of the simulation. Comparison of fields of relative humidity and vertical velocity reveals regions of ascent that are characterized by large relative humidity, while relative humidity in descending air is generally less than 30%. It is interesting to note that, while the largest mean values of vertical velocity are located in the upper troposphere, there are also regions of strong ascent centered around 1.0 and 3.5 kilometers above the surface. These lower tropospheric regions of ascent lie between layers centered around 2 km, 4 km, and 14 km in which relative humidity is large relative to the surrounding regions; levels that are consistent with the approximate heights of the trade wind inversion, zero degree Celsius level, and tropopause in observational studies [*Johnson et al.*, 1996; *Zuidema*, 1998] (see auxiliary material[1]). Note that in the model, the mean freezing level is located at a height of 4 km (not shown).

[6] The characteristics of the mean thermal structure and circulation can more easily be seen by considering a subset of the full domain that contains one convectively active region and an adjacent region undergoing large-scale descent. Examination of other convective and subsidence regions in the domain yield similar results (not shown). As in the full domain plot (Figure 1), the plot of relative humidity on a reduced domain (Figure 2a) clearly reveals relatively moist layers at 2, 4, and 14 km, as well as the presence of three distinct overturning circulation cells. The first extends from the surface to 2 km, the second is located between 2.5 and 4.5 km, and the third is centered between 6 and 13 km. The ascending branch of each cell is confined to a relatively small area within the region of active convection, while descent occurs over a relatively broad region. The appearance of a region of easterly domain-relative winds at 6 km west of the convective region is due to the presence of convective cells with tops around 6 km in this region. When these cells are removed from the mean, convergence into the convective region from the west is more clearly evident (not shown). Regions in which the temporally and meridionally averaged temperature lapse

[1]Auxiliary materials are available in the HTML. doi:10.1029/2007GL033029.

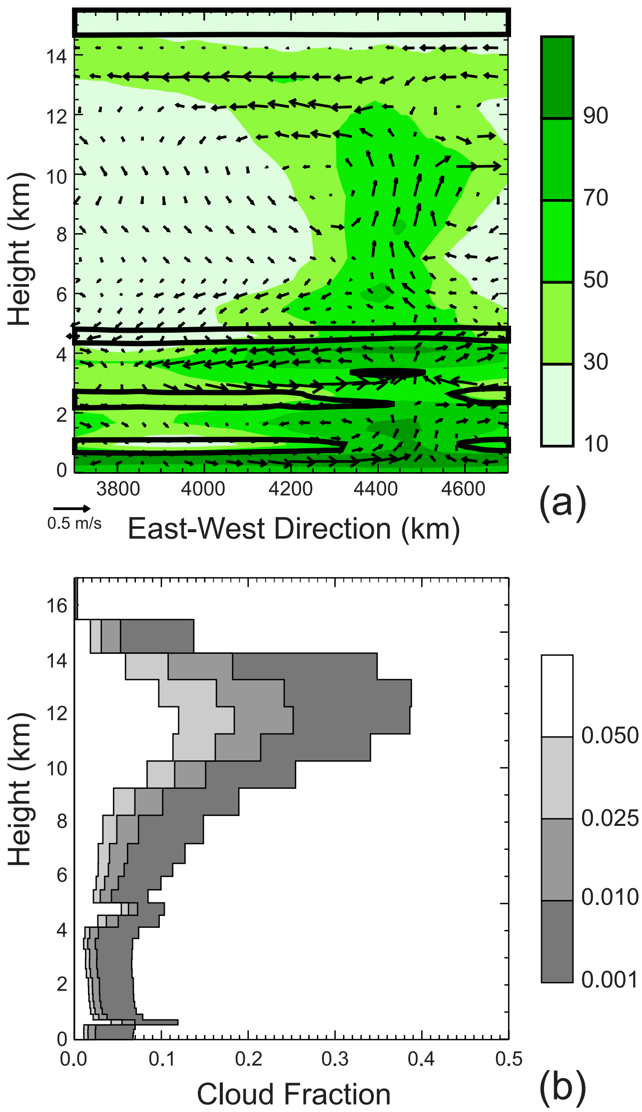

Figure 2. (a) Plot of relative humidity, winds, and temperature lapse rate for a subset of the full horizontal domain. Filled contours correspond to the relative humidity (percent), while the solid unfilled contour encloses a region in which the temperature lapse rate exceeds −4.0 K/km. Wind vectors are plotted in meters per second, and a reference vector is plotted for comparison. To highlight the vertical branches of each overturning circulation, the vertical component of the velocity has been multiplied by a factor of 10. (b) Domain averaged cloud fraction, temporally averaged over simulation days 46–55 for different cloud mass mixing ratio thresholds (depicted in grayscale shading).

rate exceeds −4 K/km (statically stable) are enclosed within a heavy black line in Figure 2a, and it can be seen that regions of relatively large static stability at approximately 2 and 4 km above the surface are co-located with the boundaries between circulation cells in the free troposphere. The stable layer atop the marine boundary layer does not divide the lowest circulation into two branches due to the fact that the marine boundary layer is the primary source region for parcels that ascend in convective updrafts. Since these updrafts are located above 1 km, mass continuity constraints require subsidence through the top of the boundary layer.

[7] As in observational studies of the tropical environment [*Mapes and Zuidema*, 1996; *Zuidema*, 1998; *Johnson et al.*, 1999; *Haynes and Stephens*, 2007], the domain and time mean cloud fraction (Figure 2b) output from the model also exhibits three distinct layers, though the observed peak at the trade wind inversion is conspicuously absent. Instead, the model generates shallow clouds atop the tropical boundary layer at a height of approximately 500 meters. Examination of the mean relative humidity fields (Figure 2a) reveals a moist layer at the 2 km level, and it is possible that the model is simply too dry at this level to support persistent cloud.

[8] The existence of two distinct circulation cells above the trade wind inversion is consistent with the results obtained by *Mapes and Houze* [1995], who described persistent convergence into mesoscale convective systems atop the freezing level during the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA-COARE). In addition, many studies have noted detrainment as a possible mechanism for the formation of the frequently-observed mid-level clouds in the tropics, and various authors have noted the existence of both mid-level stratiform and cumulus congestus clouds [*Mapes and Houze*, 1995; *Johnson et al.*, 1996]. However, the popular conception of the large scale tropical overturning (e.g., Walker) circulation is of a cell that extends in a largely unbroken fashion from the trade wind inversion to the tropopause [e.g., *Johnson et al.*, 1999]. In the current simulation, the zero degree stable layer acts to inhibit not only penetrative deep convection, but also the descending branch of the large-scale overturning circulation. As such, the vertical region around the zero degree level becomes a preferred level for both detrainment from and convergence into deep convection, as indicated in the mean wind field.

4. Discussion

[9] We now briefly discuss possible mechanisms for the formation and maintenance of the thermal and cloud structure described in the previous section. In particular, we focus on the stable layer at the freezing level, as it is this structure that divides the upper-tropospheric overturning circulation into two branches. Since previous studies have hypothesized that this stable layer arises as a result of melting of frozen stratiform precipitation [*Mapes and Houze*, 1995; *Johnson et al.*, 1996], we partition simulated latent heat release into a component associated with freezing and melting processes, and a component that includes all vapor phase transitions. Figures 3a and 3b depict domain and time mean profiles of latent heat release from ten days of simulated equilibrium for freezing/melting and vapor transition processes, respectively. A dipole of heating and cooling in the latent heat of freezing centered around the zero degree Celsius layer is clearly visible, with cooling concentrated below and heating distributed over a broader region from four to ten km. Examination of the latent heat release associated with vapor processes reveals three regions of significant heating, the lowest two of which appear to be coincident with the trade-wind and freezing-level stable layers. Examination of the time and domain

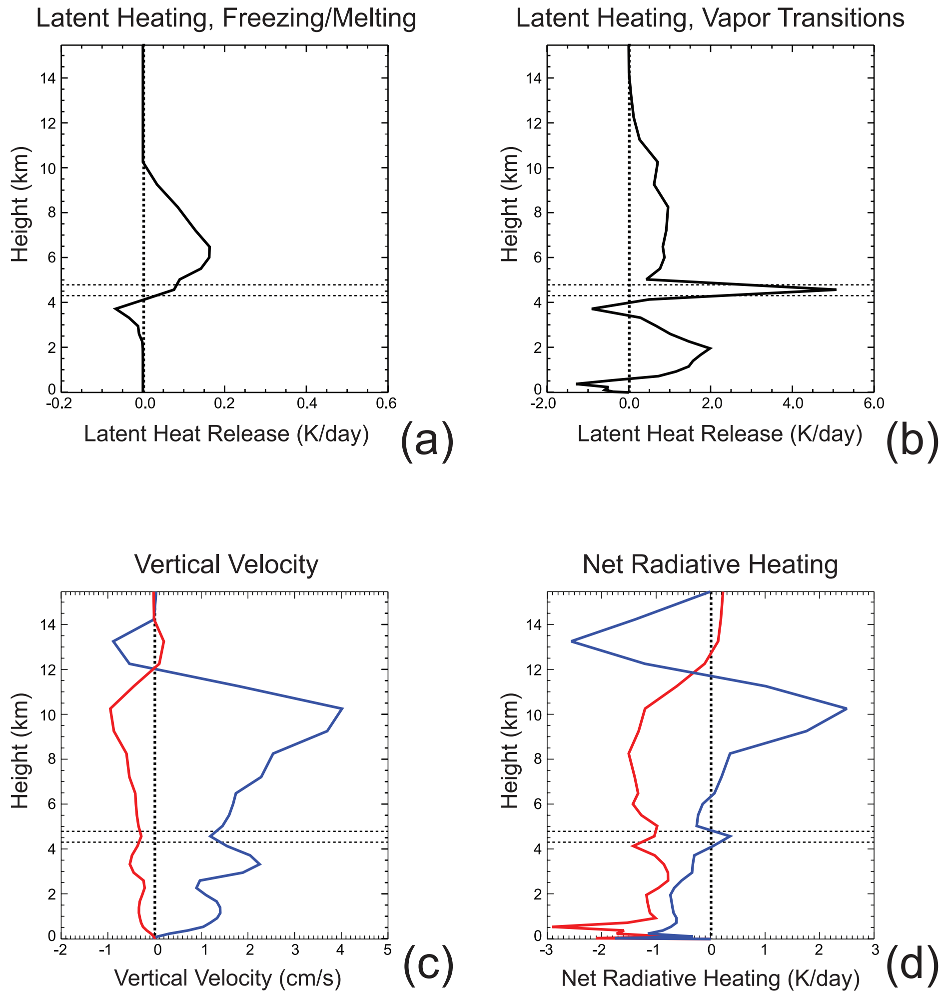

Figure 3. Profiles of profiles of domain and time averaged latent heat release due to (a) freezing and melting processes and (b) vapor transitions, along with domain and time mean (c) vertical velocity and (d) net radiative heating. Thick solid lines in Figures 3c and 3d correspond to the disturbed region, while thick dashed lines correspond to the undisturbed region (see text for details). The thin horizontal dashed lines in each plot depict the vertical extent of the stable layer at 4.5 km, defined as the region in which the temperature lapse rate exceeds -4.0 K/km. Note the difference in scale in plots Figures 3a and 3b.

mean vertical profile of vertical velocity (Figure 3c) in convective regions indicates vertical mass convergence into these regions, as does the cross-section of the mean wind (Figure 2a). Vertical mass convergence is associated with significant vertical moisture convergence (not shown), which leads to a spike in the latent heating profiles at the 2 and 4 km stable layers.

[10] To assess the properties of subsident regions, we divide the model domain into "disturbed" and "undisturbed" regions, with disturbed regions defined as areas in which cloud top height is greater than 5 km. The domain averaged profile of net radiative heating (shortwave plus longwave) in undisturbed regions (Figure 3d) indicates a rate of cooling of between 1.0 and 1.5 K/day in a layer that extends from 5 to 12 km above the surface. In addition, a close examination of the undisturbed net radiative cooling rate (Figure 3d) reveals a 0.4 K/day difference across the 4.5 km stable layer; a direct consequence of the increased relative humidity associated with detrainment from convection at and below the 4 km level. Adiabatic descent coupled with the simulated net radiative cooling rate noted above would reduce the lapse rate of a one kilometer thick layer by 2.8 K/km, while each day of differential radiative cooling across the 500 meter deep stable layer would cause the lapse rate to decrease by an additional 0.8 K/km. Hence, four days of differential radiative cooling, in combination with clear-air radiative cooling between 5 and 12 km, would be sufficient to produce a 6 K/km deviation from the adiabatic lapse rate in a layer just atop the freezing level. Though we acknowledge the importance of dry air intrusions into the tropics, it appears that strong temperature inversions near the zero-degree level can be supported in their absence through the combined action of subsidence and clear-air radiative cooling, provided cooling is allowed to act over a sufficiently long time period.

5. Summary and Conclusions

[11] In this study, simulations of radiative convective equilibrium were performed using a cloud system resolving

model with a large three dimensional horizontal domain. It was found that the model generated cloud structures and stable layers that were representative of features observed in studies of the tropical oceanic environment. Specifically, at RCE, the model domain exhibited narrow regions of organized convection separated by broad areas of clear air and descent. Each convective region was associated with three distinct large-scale overturning circulations connecting ascent in convection with subsidence in adjacent areas, and stable layers at the top of the boundary layer, two km, the freezing level, and the tropopause served as vertical boundaries for these circulations. To investigate the formation and maintenance of the freezing level stable layer, we partitioned latent heat release into contributions from freezing/melting and vapor transitions. While melting processes may be important in the initial formation of a freezing-level stable layer, the model results suggest that once it has been established, clear air subsidence in the presence of radiative cooling is sufficient to maintain it.

[12] Though the cloud-resolving radiative convective equilibrium experiments described above describe many features that have been observed in convectively active regions in the tropical oceanic environment, several outstanding questions remain. First, the RCE simulation was run without shear, and hence organized convective regions in the model did not propagate out of disturbed regions, but instead remained in relatively fixed locations. Because subsidence in the presence of longwave cooling was free to act in clear regions without interference from propagating convective systems, stable layers in the model may be stronger than those that occur in nature. We intend to conduct an observational study of the characteristics of tropical freezing-level stable layers to confirm whether or not this is the case. We also note that computational constraints limited our experiments to a model grid spacing of 2.4 km, which allows for simulation of the interaction between mesoscale convective circulations and the large scale environment, but does not accommodate realistic simulation of towering cumulus clouds or shallow convection. It has been suggested that entrainment at the tops of cumulus congestus clouds may be important in modifying the zero-degree stable layer [*Johnson et al.*, 1999], but because horizontal grid spacings of less than one kilometer may be necessary to resolve individual convective cells [*Bryan et al.*, 2003], we leave such an investigation for future work. Finally, setting a fixed solar zenith angle is necessary for the development of an equilibrium state, but does not allow for a diurnal cycle in the model. It is anticipated that inclusion of a diurnal cycle could produce convection that is more episodic in nature, leading to weakened stable layers at the 2 km and 4 km levels, and hence to smaller mass convergence and latent heat release at these levels as well. Detailed comparison of simulation results with observations over the tropical oceans, as well as investigation of the role of shear and a diurnal cycle is left for future work.

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