Simulated Precipitation Diurnal Variation With a Deep Convective Closure Subject to Shallow Convection in Community Atmosphere Model Version 5 Coupled With CLUBB

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Abstract In order to improve the physical consistency between shallow and deep convection, we modify the deep convective closure in the Community Atmosphere Model version 5 (CAM5) coupled with a third-order turbulence closure parameterization (i.e., Cloud Layers Unified by Binormals [CLUBB]). The revised closure reserves a portion of the total convective available potential energy for shallow convection via utilizing the heating and moistening profiles from CLUBB to distribute moisture and energy between shallow and deep convection. Simulations at two resolutions (i.e., 2° and 0.5°, respectively) are conducted to investigate the impacts of convective closure on the simulated precipitation diurnal variations. Results from low-resolution simulations show that the revised closure suppresses deep convection until the lower troposphere is sufficiently moistened by shallow convection, which improves the precipitation diurnal variation simulations compared with the default closure, with the precipitation diurnal peak over tropical lands delayed from 12LST to 19LST. The revised closure better simulates the diurnal variations for precipitation over the Asian monsoon region, such as the delayed precipitation onset, but still fails to well capture the nighttime peak for precipitation there. This deficiency is alleviated to some extent when applying the revised closure in high-resolution simulations, but nighttime precipitation is still underestimated probably because key processes responsible for nighttime convection are missing. Overall, our results indicate that establishing the consistency between shallow and deep convection is critical for the precipitation diurnal cycle simulations.

Plain Language Summary Establishing the consistency between shallow and deep convection is critical for correctly simulating the precipitation diurnal cycle in climate models. In this study, we propose to modify the closure for deep convection in CAM5 coupled with a unified parameterization for boundary layer mixing and shallow convection (i.e., CLU BB). The revised closure reserves a portion of the total convective available potential energy for shallow convection to distribute moisture and energy between shallow and deep convection. The revised closure suppresses deep convection until the lower troposphere is sufficiently moistened by shallow convection, which evidently improves the precipitation diurnal variation simulations over many land areas compared with the default closure.

1. Introduction

Precipitation is featured by pronounced diurnal variations, which are closely related to the fast evolution of near-surface properties and large-scale forcing in the free troposphere (e.g., Bechtold et al., 2014; Dai, 1999; Donner & Phillips, 2003; Rio et al., 2019). However, it is still challenging for climate models to accurately reproduce the observed diurnal feature of precipitation (e.g., Clark et al., 2007; Dai & Trenberth, 2004; Jones & Randall, 2011), as it is governed by various processes associated with boundary layer mixing, cumulus convection, and so on, which interplay with each other and are hard to represent (e.g., Arakawa, 2004; Qiao & Liang, 2016; Stirling & Stratton, 2012; Wu et al., 2015).
Subgrid convective processes have to be parameterized for models with grid spacings ranging from tens to hundreds of kilometers (Bechtold et al., 2004; Betts & Jakob, 2002; Brockhaus et al., 2008; Langhans et al., 2013; Yang et al., 2018). Many models employing convection parameterizations often produce a too early onset of deep convection compared with observations (Dai, 2006; Yuan, 2013). By contrast, many studies have reported the better simulated precipitation diurnal features in convection-permitting models that can, at least in part, explicitly resolve convection (e.g., Dirmeyer et al., 2012; Li et al., 2018; Marsham et al., 2013; Sato et al., 2008; Stirling & Stratton, 2012) and in super-parameterized models that incorporate a 2-D or 3-D cloud-resolving model in each grid cell of the host model (Kooperman et al., 2013; Zhang & Chen, 2016; Zhang et al., 2017).

Despite the better performance in high-resolution and super-parameterized models, development of traditional convection parameterizations is still an important task in the climate modeling community. This is mainly because parameterizations provide a summary of our understanding in physical processes, which potentially allows us to disentangle the roles of different processes in various climate phenomena (Rio et al., 2019). In convection parameterization, a closure assumption is needed to solve equations that determine the intensity of convection (Gregory & Rowntree, 1990; Krishnamurti et al., 1983; Suhas & Zhang, 2015; Tiedtke, 1989; Yano et al., 2013). Many convective closures assume a quasi-equilibrium between the cloud work function (e.g., convective available potential energy [CAPE]) consumed by deep convection and that produced by large-scale forcing (e.g., Arakawa & Schubert, 1974; Zhang & Mcfarlane, 1995). However, the quasi-equilibrium assumption may not hold and non-equilibrium can happen under conditions with rapidly varying surface fluxes or strong upper tropospheric forcing on diurnal time scale, particularly over tropical and midlatitude lands where the imbalance between deep convection and boundary layer forcing should be taken into account (Donner & Phillips, 2003; Zhang, 2002, 2003a). One prominent example for non-equilibrium convection is the precipitation diurnal cycle over lands that are usually associated with the transition from shallow to deep convection (Bechtold et al., 2014).

In the past decade, significant efforts have been made to improve the simulation of the non-equilibrium convection such as the precipitation diurnal cycle (e.g., Fuchs & Raymond, 2007; Gerard et al., 2009; Pirioi et al., 2007; Park, 2014b; Rio et al., 2009; Stratton & Stirling, 2012; Tawfik et al., 2017). For example, Xie and Zhang (2000) and Xie et al. (2004) emphasized that a favorable large-scale dynamic condition is an important factor for convection triggering, and their trigger function based on large-scale dynamic forcing can evidently improve the precipitation simulation. An alternative approach is taken by Tawfik et al. (2017) who introduced a convective initiation based on subgrid surface properties in the Community Earth System Model (CESM). The new subgrid trigger delays the onset of convection by 1–4 h but has relatively weak impacts on the peak time of precipitation. Closures modified from the traditional quasi-equilibrium assumption have also improved the simulated precipitation diurnal cycle. For example, Zhang (2003b) showed that a CAPE closure assuming quasi-equilibrium for the free troposphere works for non-equilibrium convection driven by boundary layer forcing.

Despite the different constraints used, all approaches shown above act to prevent deep convection from responding too quickly to the boundary layer forcing, which potentially allows the model to better simulate the transition from shallow to deep convection, a key element for the precipitation diurnal cycle simulation (Rio et al., 2019). To couple deep convection with shallow convection in a more explicit way, Bechtold et al. (2014) developed a CAPE closure under the free troposphere quasi-equilibrium assumption, considering that CAPE production from boundary layer forcing is partly consumed by shallow convection. The new closure evidently improves the diurnal phases of precipitation in the European Centre for Medium-Range Weather Forecasts Integrated Forecasting System. In Rio et al. (2009), the triggering and closure for deep convection are linked to boundary layer thermals associated with shallow convection. The updated parameterization greatly improves the precipitation diurnal cycle simulations in both 1-D (Rio et al., 2009) and global-scale experiments (Rio et al., 2013). Instead of using separate parameterizations, Park (2014a, 2014b) developed a unified framework for both shallow and deep convection, which can better simulate the precipitation diurnal cycle over lands because it correctly captures the seamless transition from shallow to deep convection and the feedback of mesoscale organized flow on convection.

Recently, a third-order turbulence closure parameterization, Cloud Layers Unified by Binormals (CLUBB) (Golaz et al., 2002; Larson & Golaz, 2005), has been implemented into several global climate models.
including CESM (Bogenschutz et al., 2013). CLUBB provides a unified treatment of boundary layer turbulence, shallow convection, and stratiform cloud macrophysics, which can lead to better simulated low clouds than the traditional approaches using separate parameterizations (Bogenschutz et al., 2013; Guo et al., 2014, 2015). In CESM, the Zhang-McFarlane (ZM) deep convection scheme (Zhang & Mcfarlane, 1995) is still needed to parameterize deep convection partly because CLUBB was originally designed for warm cloud processes. Recent work by Lin et al. (2019) showed that the unified treatment of CLUBB for boundary layer turbulence and shallow convection helps improve the simulations of the precipitation diurnal cycle of Asia in CESM. However, apparent biases still exist in many regions. CLUBB is also applied in Department of Energy’s Energy Exascale Earth System Model (E3SM). Xie et al. (2019) modified the convective triggering function in E3SM atmosphere model version 1 (EAMV1) (Rasch et al., 2019; Xie et al., 2018) by introducing a dynamic constraint on convection initiation and a looser restriction on the launch level for convection updrafts. The new trigger was found to markedly improve the simulations of the precipitation diurnal cycle over tropical and midlatitude lands.

Here, we propose to modify the ZM deep convective closure subject to shallow convection in order to improve the physical consistency between shallow and deep convection and the simulation of precipitation diurnal variation in CESM. In the revised closure, a portion of the total CAPE is reserved for use by shallow convection prior to calling the deep convection scheme. This is achieved via utilizing the heating and moistening profiles from CLUBB. Section 2 introduces the proposed modification to deep convective closure. Section 3 describes the model experiments and observational data sets. In section 4, we analyze the impacts of deep convective closure on the simulated precipitation diurnal variation. We first briefly evaluate the simulated precipitation diurnal features in a global context. Then, we further focus on results over the Asian summer monsoon region where the precipitation diurnal features exhibit strong regional variations (Yu et al., 2007; Zhou et al., 2008). Model results at two different resolutions are examined. The main findings are summarized in section 5 along with discussions for future studies.

2. Deep Convective Closures

In the ZM parameterization, the column-integrated CAPE is relaxed to a reference value (i.e., $CAPE_0$; corresponding to the “capelmt” parameter in the code in CESM) within an adjustment time scale $\tau$ (Zhang & Mcfarlane, 1995):

$$
\left( \frac{\partial CAPE}{\partial t} \right)_{cu} = \frac{CAPE - CAPE_0}{\tau},
$$

where $\left( \frac{\partial}{\partial t} \right)_{cu}$ means the tendency due to deep convection. Bechtold et al. (2014) updated this type of closure, assuming that CAPE production from boundary layer forcing is partly consumed by shallow convection.

In CLUBB, boundary layer mixing and shallow convection are treated in a unified way. It is reasonable to assume that in CLUBB, the temperature ($T$) and moisture ($Q$) tendencies above the top of the boundary layer are produced by shallow convection. The boundary layer top is computed according to the Holtslag-Boville scheme (Holtslag & Boville, 1993), which is based on the bulk Richardson number. With energy and moisture conservation (for column-integrated values) for non-precipitating shallow convection (before water phase change), we have

$$
\int_{P_{sat}}^{P_{top}} \left( \frac{\partial \theta}{\partial t} \right)_{sh} dp = -\int_{P_{sat}}^{P_{top}} \left( \frac{\partial Q}{\partial t} \right)_{sh} dp, \quad (2)
$$

and

$$
\int_{P_{sat}}^{P_{top}} \left( \frac{\partial Q}{\partial t} \right)_{sh} dp = -\int_{P_{sat}}^{P_{top}} \left( \frac{\partial \theta}{\partial t} \right)_{sh} dp, \quad (3)
$$

in which $\theta$ is potential temperature and $p$ is pressure. $P_{sat}$, $P_{sh}$, and $P_{top}$ are pressures at surface, boundary layer top, and top of atmosphere, respectively. $\left( \frac{\partial}{\partial t} \right)_{sh}$ means the tendency due to shallow convection. Assuming that these tendencies due to shallow convection do not change with height within the boundary layer, we can derive the vertical profiles of these tendencies. Then, we can diagnose the CAPE tendency produced by shallow convection (i.e., $\left( \frac{\partial CAPE}{\partial t} \right)_{sh}$; computed using the “closure"
subroutine in the ZM scheme), which is used to update the CAPE closure in the ZM deep convection scheme as follows:

\[
\frac{\partial (\text{CAPE})}{\partial t} = \frac{\text{CAPE} - \text{CAPE}_0}{\tau} - \alpha \frac{\partial (\text{CAPE})}{\partial t}
\]

A tunable parameter \( \alpha \) is included here. When \( \alpha \) is smaller than 1, it means within the adjustment time of deep convection, a fraction (i.e., \( 1 - \alpha \)) of \( \frac{\partial (\text{CAPE})}{\partial t} \) is supplied by newly generated CAPE during this period. Here we set \( \alpha \) to 1 considering that deep and shallow convection remove CAPE on similar time scales (Donahue & Caldwell, 2018; Kain, 2004). Note that \( \frac{\partial (\text{CAPE})}{\partial t} \) is negative, that is, shallow convection consumes CAPE. Thus, the modification in Equation 4 reduces the amount of CAPE available to deep convection, reflecting a non-equilibrium relationship between deep convection and CAPE production from other sources. On diurnal time scale, CAPE responds fast to surface heating from radiative fluxes. However, deep convection does not respond as fast. By making a portion of CAPE available to shallow convection, it serves to delay the initiation of deep convection.

Note that all the above calculations are completed in the ZM scheme with the CLUBB scheme kept unchanged. Given that the ZM scheme is called before CLUBB in CESM, the \( T \) and \( q \) tendencies from CLUBB at the previous time step are passed to ZM to calculate \( \frac{\partial (\text{CAPE})}{\partial t} \).

### 3. Model Experiments, Observational Data Sets, and Method

#### 3.1. Experimental Design

We use the CESM version 1.2.1 with the Community Atmosphere Model version 5 (CAM5) physics (Neale et al., 2010) coupled with CLUBB (referred to as CAM5_CLUBB). Three CAM5_CLUBB experiments, that is, CNTL, REV, and REV2, are conducted (see Table 1). The CNTL experiment uses the default convective closure (Equation 1), and the two sensitivity experiments REV and REV2 use the revised closure (Equation 4). In ZM, CAPE is determined by the atmospheric conditions of both the updraft’s source layer and its environment (Zhang, 2002). In the REV experiment, we consider only the impacts of shallow convection on the updraft’s source layer, while in REV2, the impacts of shallow convection on the whole atmosphere column are included. All three experiments are forced by observed monthly sea surface temperature and sea ice conditions with a horizontal resolution of 2.5° longitude × 1.9° latitude and 30 vertical hybrid levels. The experiments are integrated for 6 years from 2000 to 2005, with the last 5 years analyzed here.

Deep convection can also be suppressed via different approaches such as tuning some key parameters in the ZM scheme. For example, the entrainment rate for deep convection is one of the most influential parameters that determine the intensity of deep convection (Qian et al., 2015; Yang et al., 2013). To examine whether the modified closure just simply acts as a way to reduce the intensity of deep convection, we conduct two experiments (i.e., ENT and ENT_REV; Table 1) using the default and revised closures but with a stronger entrainment rate for deep convection, that is, \(-0.002 \text{ m}^{-1}\) that is at the upper end of its uncertainty range based on previous studies (e.g., Yang et al., 2013).

Moreover, climate models contain many components that are generally developed independently. The way of implementing the coupling among different components can have a considerable impact on the model behavior (Gross et al., 2018; Donahue & Caldwell, 2018). In CAM5_CLUBB, CLUBB is called after ZM. The modified convective closure reserves a part of CAPE for the ensuing shallow convection. It is possible that switching the order of the calls between CLUBB and ZM may have a similar effect to revising the convective closure. To examine the model sensitivity to the order between physics, two additional experiments (i.e., CLB1st and CLB1st_REV; Table 1) are conducted that call CLUBB before ZM.
We also conduct two 1-year high-resolution (i.e., 0.63° longitude × 0.47° latitude) experiments, that is, F05 and F05_RE, which use the same physics as CNTL and REV, respectively. Only the June-July-August (JJA) results of these two experiments are evaluated.

3.2. Observational Data Sets
To evaluate the simulated diurnal variation of precipitation, we use the precipitation product from the Tropical Rainfall Measuring Mission (TRMM, Huffman et al., 2007) 3B42 version 7 (referred to as TRMM; available at ftp://disc2.nascom.nasa.gov/data/TRMM/Gridded) from 2001 to 2005. The TRMM data have a spatial resolution of 0.25° and a temporal resolution of 3 h. We also use the hourly rain gauge data acquired from the National Meteorological Information Center of China Meteorological Administration to assess the model performance over the East Asian summer monsoon region.

3.3. Analysis Method
Normalized precipitation diurnal cycle is derived, following Yuan (2013) and Zhang and Chen (2016):

\[ N(h) = \frac{P(h) - P_{\text{mean}}}{P_{\text{mean}}} \]

where \( P(h) \) and \( N(h) \) are the original and normalized hourly precipitation, respectively, and \( P_{\text{mean}} \) is the daily mean precipitation. The diurnal peak time of precipitation is defined as the hour when the daily maximum precipitation happens.

Individual rainfall events are identified for both observational and model results to determine the onset and end time of precipitation. First, we define non-precipitating hours as those with precipitation intensity below 0.2 mm h\(^{-1}\). Then, two or more consecutive non-precipitating hours are used to split the long-term precipitation time serials into different rainfall events. Note that the threshold of 0.2 mm h\(^{-1}\) is not used in the calculations of any other precipitation metrics.

Global models often underestimate the occurrence frequency of heavy precipitation. Here we evaluate the diurnal variation of the occurrence frequency for moderate-to-heavy precipitation in the model (i.e., >2.5 mm h\(^{-1}\); https://en.wikipedia.org/wiki/Rain#Intensity). Moderate and heavy precipitation are combined into one category because the occurrence frequency of heavy precipitation (i.e., >7.6 mm h\(^{-1}\)) is too small in all simulations.

4. Results
4.1. Precipitation Climatology
First, we briefly assess the simulated annual mean precipitation (2001–2005) in experiments with different convective closures (Figure 1). In observations (Figure 1a), we can observe large precipitation amounts over the tropical oceans such as the intertropical convergence zone (ITCZ), the Southern Pacific convergence zone, and the eastern equatorial Indian Ocean. Strong rainfall centers are also seen over the tropical land regions such as the Maritime Continent and the tropical areas of Africa and South America. The CNTL experiment (Figure 1b) with the default closure well reproduces the annual mean precipitation pattern shown in TRMM (Figure 1a). Changing the convective closure has only a very small impact on the simulated precipitation climatology (Figures 1b–1d). All three experiments simulate a double ITCZ with overestimated precipitation in the southeastern Pacific. Meanwhile, all experiments overestimate the precipitation over the western equatorial Indian Ocean and underestimate the precipitation over the East Asian monsoon region and North America.

4.2. Precipitation Diurnal Variation
4.2.1. Global Distribution of Precipitation Diurnal Feature
In Figure 2a, we present the spatial distribution of the diurnal peak time for the annual mean precipitation in TRMM, which shows a strong contrast between ocean and land regions. Precipitation generally peaks in nighttime or early morning over oceans but late afternoon over lands. Nighttime or early morning peaks of precipitation can also be found in some land regions, including the southern slope of the Tibet Plateau (TP), as well as areas to the east of large terrains such as TP and Rocky Mountains where precipitation exhibits strong propagation features in the diurnal peak time (Dai, 1999; Wu et al., 2018; Yu et al., 2007).
The simulated diurnal features of precipitation in experiments with different closures (Figures 2b–2d) are compared against the TRMM observation. With the default closure, the CNTL experiment (Figure 2b) can simulate the nighttime peak of precipitation over oceans but fails to reproduce the late afternoon peak of precipitation over many land regions. For example, CNTL produces a too early peak of precipitation over tropical lands in Africa and South America, with the peak time at around 12LST compared with 16–18LST in TRMM. Evident biases are also found over midlatitudes. CNTL can somehow reproduce the nighttime peak of precipitation over the southern and eastern flanks of TP but is unable to capture the nighttime peak of precipitation over North America.

The simulated diurnal variation of precipitation exhibits a strong sensitivity to convective closure, particularly over lands. Although in some regions the diurnal peak of precipitation still occurs too early, using the revised closure considerably delays the diurnal peak over many areas of South America and Africa (Figures 2c and 2d), showing better agreement with TRMM (Figure 2a). The improved diurnal peaks are also seen when the calculations are based on the first diurnal harmonic of precipitation.

Figure 1. Spatial distributions of annual (2001–2005) mean precipitation from (a) TRMM and simulations of (b) CNTL, (c) REV, and (d) REV2.
Figure S1). The revised closure also better simulates the diurnal peak of precipitation over East Asia to some extent. However, it is unable to capture the nighttime peak of precipitation over North America. The revised closure slightly improves the diurnal peaks for precipitation over oceans, such as over the ITCZ where the peak time is delayed from around 05LST to 07LST. The improvements are less evident over ocean than over land mainly due to the weak diurnal variation of the reserved CAPE over ocean because shallow convections are relatively statistically steady there (Brown et al., 2002). The results of REV2 (Figure 2d) are generally similar to those of REV because they produce very similar CAPE tendency of...
shallow convection (Figure S2), indicating that in the ZM scheme, CAPE tendency is dominated by the changes of moisture and temperature in the updraft source layer. In the rest of the paper, we mainly compare the results of CNTL and REV.

**Figure 3.** Diurnal cycle of precipitation (horizontal axis) and its seasonal variation (vertical axis) over southern Amazon (70°W–50°W, 15°S–0°) from (a) TRMM and simulations of (b) CNTL and (c) REV.

shallow convection (Figure S2), indicating that in the ZM scheme, CAPE tendency is dominated by the changes of moisture and temperature in the updraft source layer. In the rest of the paper, we mainly compare the results of CNTL and REV.
The seasonal variations in the precipitation diurnal feature over southern Amazon (70°W–50°W, 15°S–0°) are further analyzed given that the precipitation simulations there are greatly improved (Figure 3). Observations indicate that precipitation generally peaks at 17LST throughout the year (Figure 3a), which is consistent with previous findings (e.g., Zhang et al., 2017). The CNTL experiment underestimates the total precipitation amount during both wet and dry seasons, and the diurnal peak of precipitation usually occurs too early (Figure 3b) compared with observations. During wet season, both the magnitude and the diurnal feature of precipitation are better simulated in the REV experiment with the revised closure (Figure 3c). However, the revised closure has little effect on the simulations of the precipitation diurnal feature during dry season. It still overestimates the precipitation amounts at around noon, indicating that we may have underestimated the effects of shallow convection or neglected some other effects that also consume CAPE at this time. Model results show that the moistening of the upper part of the boundary layer lags that in the lower part, implying that the convective boundary layer does not reach its equilibrium state simultaneously and its upper part can gradually acquire energy accumulated in the lower part that is "potentially available" for the deep convective updrafts in ZM. Our preliminary results suggest that reserving some portion of CAPE for the mixing within the boundary layer (or heating and moistening in its upper part) with an appropriate adjustment time scale could further improve the simulated precipitation diurnal cycle. However, whether this treatment is physically justifiable needs further examination.

Figure 4. Diurnal peak times (LST) for JJA (2001–2005) mean precipitation from observations of (a) TRMM and (b) rain gauge station, as well as simulations of (c) CNTL and (d) REV over land areas of the Asian monsoon region. The black boxes in (e) denote different subregions of China: (1) southern China, (2) middle Yangtze River Valley (YRV), (3) region between Yangtze and Yellow Rivers (YZ-YR), and (4) Sichuan Basin. Areas with mean precipitation magnitude below 1 mm day$^{-1}$ are masked out.
4.2.2. Precipitation Diurnal Variation Associated With Asian Summer Monsoon

Over the Asian monsoon region, a large portion of annual rainfall is contributed by summertime (i.e., JJA) precipitation associated with the Asian summer monsoon. Figure 4 shows the diurnal peak time of summertime precipitation over the Asian monsoon region. In TRMM (Figure 4a), the diurnal feature of precipitation mainly shows a late afternoon peak over most lands of the monsoon region but with evident region-to-region variations due to the strong heterogeneity in the underlying surface type and terrain (Yu et al., 2007; Zhou

Figure 5. Diurnal variations of JJA (2001–2005) mean precipitation (normalized by daily mean value) from observations (i.e., TRMM and rain gauge station) and simulations (i.e., CNTL and REV) over different subregions of China (see Figure 4e).

Figure 6. Diurnal variations of the occurrence frequency of precipitation onset in JJA (2001–2005) from observation (i.e., rain gauge station) and simulations (i.e., CNTL and REV) over different subregions of China (see Figure 4e).
Areas with nighttime precipitation peak are mostly located over the southern and eastern flanks of TP. In addition to the TRMM data, hourly rain gauge records over mainland China are also utilized for model evaluation here. We can see that the spatial distribution of the precipitation diurnal peak time from the hourly rain gauge data (Figure 4b) is overall consistent with the 3-hourly TRMM data (Figure 4a).

Figure 7. Diurnal variations of the occurrence frequency of moderate-to-heavy precipitation (>2.6 mm h$^{-1}$) in JJA (2001–2005) from observation (i.e., rain gauge station) and simulations (i.e., CNTL and REV) over different subregions of China (see Figure 4e).

Figure 8. Diurnal and zonal distributions of JJA (2001–2005) mean precipitation (normalized by daily mean value) averaged within 28°–35°N from observations of (a) TRMM and (b) rain gauge station, as well as simulations of (c) CNTL and (d) REV.
However, for the rain gauge data, areas featuring nighttime precipitation peak are much broader, particularly for the region to the east of TP and the region between Yangtze and Yellow Rivers (YZ-YR; see Figure 4e for location).

Similar to the annual mean results (Figure 2), the summertime diurnal feature of precipitation is sensitive to convective closure over the monsoon region. The CNTL experiment (Figure 4c) well captures the diurnal peak time for precipitation over India, Indochina Peninsula, and TP. However, biases are apparent over the broad eastern China (i.e., east of 103°E), with precipitation mainly peaking at 12LST that is too early compared with observations. The CNTL experiment also fails to capture the nighttime peak of precipitation over YZ-YR. This problem can be alleviated to some extent via changes in other model configurations as discussed later. Similar results can be found if the diurnal peaks are defined based on the first diurnal harmonic of precipitation (Figure S3).

Figure 5 presents the simulated diurnal variations of normalized precipitation (see section 3.3) over four different subregions of China, that is, southern China, the middle Yangtze River Valley (YRV), YZ-YR, and Sichuan Basin (see Figure 4e for locations), which are selected based on their precipitation diurnal features (e.g., Yang et al., 2018; Zhou et al., 2008). In southern China (Figure 5a), both the rain gauge (black line) and

![Figure 9. Diurnal peak times (LST) for JJA (2000) mean precipitation from simulations of (a) F05 and (b) F05_REV over land areas of the Asian monsoon region. Areas with mean precipitation magnitude below 1 mm day$^{-1}$ are masked out.](image-url)
the TRMM (black dot) data sets show a dominant late afternoon peak at around 17LST. The CNTL simulated precipitation (red line) peaks at 15LST, which is shifted to 17LST when using the revised closure (blue line). In the middle YRV (Figure 5b), the observed precipitation has two diurnal peaks respectively at 14LST and 05LST as indicated by previous studies (e.g., Zhou et al., 2008). The two experiments well capture the daytime peak of precipitation but fail to simulate the nighttime peak. In YZ-YR (Figure 5c), the two observational data sets show different diurnal features of precipitation, with rain gauge data having two peaks in late afternoon (around 17LST) and early morning (around 06LST) but only one dominant late afternoon peak in TRMM. Previous studies (e.g., Zhou et al., 2008) have indicated that TRMM may have difficulty in detecting the morning peak of precipitation in this region. The CNTL experiment simulates a strong daytime peak of precipitation, but the peak time is 3 h earlier than in observation. The early morning peak of precipitation is not apparent in CNTL. In contrast, the revised closure can produce two diurnal peaks, but the nighttime peak is too weak compared with the daytime peak and cannot be seen in the spatial distribution (Figure 4d). In the Sichuan Basin (Figure 5d), observations indicate a dominant nighttime peak of precipitation. The CNTL experiment fails to capture this feature and produces more precipitation in daytime than in nighttime. When using the revised closure, the simulated precipitation diurnal cycle is remarkably improved. Overall, the revised closure tends to delay the daytime peak of precipitation and produce more nighttime precipitation. However, apparent biases in the nighttime peak of precipitation still exist, partly due to the restriction of air parcel originating from within the boundary layer as assumed in ZM (Xie et al., 2019). It is also possible that these biases come about because the model cannot reasonably represent the self-organization behavior of convection (Feng et al., 2015; Mapes & Neale, 2011; Rio et al., 2019).

The diurnal variations for the precipitation onset (see section 3.3) over different subregions of China are given in Figure 6. In observations, the precipitation onset mainly occurs at 14LST over southern China (Figure 6a) and 16LST over YZ-YR (Figure 6c). By contrast, there are two diurnal peaks at 14LST and 00LST over the Middle YRV (Figure 6b) and Sichuan Basin (Figure 6d). In CNTL, precipitation often develops too early (red line), which is alleviated to some degree in the REV experiment (blue line). However, both the default and the revised closures fail to capture the onset time of precipitation at night in the Sichuan Basin (Figure 6d). The occurrence frequencies for precipitation with different intensities are also sensitive to convective closure. The default closure often underestimates the occurrence frequency of moderate-to-heavy precipitation (Figure 7), while the revised closure performs better in this regard in terms of both mean magnitude and diurnal variations, particularly over southern China and the Sichuan Basin (Figures 7a and

Figure 10. Same as Figure 5 except that the simulations are from F05 (red lines) and F05_REV (blue lines) for the year of 2000.
7d). As a result, the revised closure can generally better capture the probability density functions of precipitation based on hourly values (Figure S4).

We further evaluate the propagation feature of the precipitation diurnal signal downstream of TP (Figure 8), which is a prominent phenomenon over East Asia (e.g., Yu et al., 2007). In observations (Figures 8a and 8b), precipitation over TP (i.e., west of 100°E) features a strong diurnal peak at around 20LST. To the east of TP (i.e., 100°E–110°E), there is a gradual shift in the peak time of precipitation from 00LST to 06LST in both TRMM and rain gauge data although some inconsistencies exist between the two data sets. Both the CNTL and the REV experiment fail to capture the propagation signal in the precipitation peak time downstream of TP (Figures 8c and 8d).

The improvements in the simulated precipitation diurnal features are also seen and even more evident when using the revised closure in high-resolution applications (Figure 9). With the default closure (i.e., F05; Figure 9a), the precipitation diurnal features from one-summer high-resolution results are generally consistent with the five-summer mean results in the low-resolution simulations (Figures 4c). Compared with the

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**Figure 11.** Diurnal pressure distributions of annual (2001–2005) mean (a, c, e) temperature and (b, d, f) moisture tendencies induced by the combined effects of moist processes and subgrid transport from (a, b) CNTL, (c, d) REV, and (e, f) their differences averaged over southern Amazon (70°W–50°W, 15°S–0°).
F05 experiment, the F05_REV simulated precipitation diurnal variations (Figure 9b) show better agreement with observation (Figures 4a and 4b), such as the late afternoon peak of precipitation over southern China and the nighttime peak over the Sichuan Basin and YZ–YR that are not seen in the low-resolution simulations (Figure 4d). Figure 10 presents the simulated diurnal variations of precipitation over the same four sub-regions as in Figure 5. Compared with the low-resolution simulations, the improvements of using the revised closure are more evident in the high-resolution simulations probably because the dependence on model resolution is larger for resolved (i.e., large-scale) precipitation than convective precipitation in CAM5, which potentially increases the precipitation response to model resolution when resolved and convective precipitation are respectively enhanced and suppressed with the revised closure (Yang et al., 2013).

4.3. Diurnal Variations in Heating and Moistening Profiles

We further analyze the impacts of convective closure on the heating and moistening profiles to better understand the responses in the simulated precipitation diurnal variations. We first show in Figure 11 the diurnal pressure cross sections of the annual mean temperature and moisture tendencies due to moist processes and
subgrid transports (i.e., transports of moisture and heat by subgrid air motions such as turbulence and convection) averaged over southern Amazon. In both the CNTL and the REV experiments (Figures 11a and 11c), the low-level (i.e., below 800 hPa) heating due to moist processes and subgrid transports is stronger during daytime than during nighttime. Differently, the heating rate in the upper level (i.e., above 600 hPa) shows a relatively weak diurnal variation. At night, the strongest heating is located at around 500 hPa, accompanied by a thin layer featuring cooling just above the surface. Compared with CNTL, the REV experiment simulates a stronger heating below 600 hPa (Figure 11e) from 10 to 17LST. After 18LST, the midtroposphere heating is stronger in the REV experiment, corresponding to the intensified deep convective updraft mass fluxes (Figure 12e). Meanwhile, the low-level cooling is also enhanced due to evaporation of rainwater associating with the stronger deep convective downdraft (Figure 12f). This indicates that precipitation processes are intensified during this period, corresponding to the delay of the diurnal peak of precipitation (Figures 2 and 3). The diurnal variations of moistening profiles (Figures 11b and 11d) can be related to that of heating profiles. After 08LST, intensified surface evaporation and boundary layer mixing tend to moisten the boundary layer. Then (i.e., after 12LST), water vapor is further

Figure 13. Same as Figure 11 but for JJA mean results averaged over the broad region of eastern China (103°–120°E, 22°–40°N).
transported from the boundary layer to the low troposphere due to the enhancements of both shallow and deep convection. Compared with CNTL (Figure 11f), the REV experiment allows the atmosphere from the boundary layer top to 600 hPa be sufficiently moistened before 17LST because of the suppressed deep convection (based on Equation 4), which is favorable for the development of convection after onset (Figures 12e and 12f) and the occurrence of nighttime precipitation as indicated by the enhanced drying tendency above 900 hPa and moistening tendency below after 17LST (Figure 11f).

Figure 13 presents diurnal pressure cross sections of the JJA temperature and moisture tendencies due to moisture processes and subgrid transports averaged over the broad eastern China (103°–120°E, 22°–40°N).

We can see that the responses of heating and moistening profiles are generally similar to those based on the annual mean results over southern Amazon (Figure 11).

4.4. Impacts of Closure Versus Parameter Tuning

The revised closure may just simply act as a way to reduce the intensity of deep convection similar to parameter tuning. To check on this, Figure 14 presents the spatial distribution of the peak time of annual mean precipitation in experiments using the default and revised closures but applying a stronger entrainment rate for deep convection (ENT and ENT_REV; see Table 1) to effectively reduce the intensity of deep convection (e.g., Yang et al., 2013). It shows that increasing the entrainment rate has little impact on the simulated peak time of precipitation (Figure 14a), whereas using the revised closure together with the stronger entrainment rate (Figure 14b) evidently improves the simulations over most tropical land areas.

Similar results can be found over the East Asian monsoon region (Figure 15), where the simulated precipitation peak time exhibits a weak sensitivity to the perturbation in the deep convection entrainment rate (Figure 15a vs. Figure 4c). Using the revised closure together with the stronger entrainment rate improves the simulations of the precipitation peak time and better captures the nighttime peaks of precipitation over YZ-YR (Figure 15b), which is not the case for the experiment using the revised closure or using the stronger entrainment rate alone, indicating the necessity of both parameter tuning and making structural improvements to the parameterizations (Qian et al., 2018; Yang et al., 2019).

Figure 14. Spatial distributions of diurnal peak times (LST) for annual (2001–2005) mean precipitation from simulations of (a) ENT and (b) ENT_REV. Areas with mean precipitation magnitude below 2 mm day$^{-1}$ are masked out.
4.5. Sensitivity to the Order of Calls Between Shallow and Deep Convection

In CAM5_CLUBB, CLUBB is called after ZM when the total CAPE has been largely consumed by deep convection. Donahue and Caldwell (2018) have emphasized that the positioning of shallow convection plays an important role on the model solution. The revised convective closure improves the simulated precipitation diurnal variation by reserving a portion of CAPE for CLUBB. One might expect that switching the order between the calls of CLUBB and ZM may also be able to delay the onset of deep convection.

Figure 16a presents the spatial distribution of the peak time of annual mean precipitation simulated in the experiment with CLUBB called before ZM (i.e., CLB1st; see Table 1). It shows that the results from CLB1st are overall consistent with that in CNTL. Two reasons are possibly responsible for these results. First, the revised closure reserves a considerably large portion of CAPE for maintaining shallow convection for an additional 1 h (i.e., the adjustment time scale of deep convection), which is different from the case of switching the order between CLUBB and ZM. Second, all physical processes in CAM5 are coupled together via the sequential update splitting approach (Donahue & Caldwell, 2018), in which the model state is updated after a single process before it is used as the input to the next process. Given the unified treatments in CLUBB, the coupling strategy between boundary layer turbulence and shallow convection is more like the parallel splitting approach, indicating that the two processes are fed the same model state though they are tightly connected. This means that most of the updated CAPE accumulated from surface fluxes might go to deep convection before shallow convection can sufficiently respond, even when CLUBB is

Figure 15. Diurnal peak times (LST) for JJA (2001–2005) mean precipitation from simulations of (a) ENT and (b) ENT_REV over land areas of the Asian monsoon region. Areas with mean precipitation magnitude below 1 mm day$^{-1}$ are masked out.

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called before ZM. When the revised closure is used together with calling CLUBB first (i.e., CLB1st_REV), the precipitation peaks are evidently delayed over most tropical land areas (Figure 16b), indicating that the benefits of using the revised closure is transferable across model configurations with different physics orders.

5. Summary and Discussion

In this study, we proposed to modify the ZM deep convective closure via utilizing the information of shallow convection from CLUBB, in which boundary layer mixing and shallow convection are treated in a unified approach. Similar to the concept of Bechtold et al. (2014), the revised closure reserves a portion of CAPE for shallow convection within the adjustment time scale of deep convection. Simulations using both the default and revised convective closures were conducted at two different horizontal resolutions to examine the impacts of convective closure on the simulated precipitation diurnal feature. We first briefly analyzed the precipitation diurnal variation in a global context and then focused on the results over the Asian summer monsoon region.

Our results indicated that the precipitation diurnal features simulated in the low-resolution (2.5° longitude × 1.9° latitude) experiments are very sensitive to convective closure, particularly over lands. Compared with the default closure, the revised closure evidently improves the simulated precipitation diurnal variation. For example, precipitation simulated by the default closure usually peaks at 12LST over the tropical areas of Africa and South America, which is about 3–4 h earlier than in observation. The revised closure delays the daytime peak of precipitation, agreeing more with observation. Similarly, the default closure often produces a too early peak for the daytime precipitation over most land areas in the East Asian summer monsoon region and fails to capture the nighttime peak of precipitation over areas to the east of TP (e.g., the Sichuan Basin) and YZ–YR (i.e., region between Yangtze and Yellow Rivers). By contrast, the revised closure delays the daytime peak of precipitation but still fails to well capture the nighttime peaks there. The revised closure is also found to improve the precipitation diurnal variation over East Asia in high-resolution simulations. Notably, the revised closure is able to simulate the nighttime peaks over the Sichuan Basin and YZ–YR that are not seen in the low-resolution simulations. Further analyses indicate that the diurnal features of
precipitation cannot be improved via tuning the entrainment rate for deep convection or switching the order of calls between shallow and deep convection alone.

Overall, the revised closure better captures the daytime peak of precipitation and produces more reasonable nighttime precipitation because it suppresses deep convection until the lower troposphere is sufficiently moistened by shallow convection, which helps delay the development of deep convection. However, apparent biases still exist in the simulation of the nighttime peak of precipitation in many regions, which is partly due to the restriction of air parcel lifting level within the boundary layer as assumed in ZM (e.g., Xie et al., 2019).

It should be noted that the improved precipitation diurnal feature with the revised closure might be caused by the compensation of biases from other model components. For example, the parameterized shallow convection might be too weak in CLUBB and the revised closure can lead to enhanced shallow convection by suppressing deep convection. In that case, our findings can be useful to detect the origins of such bias compensation. Moreover, as discussed earlier, the better precipitation diurnal cycle can be achieved via various approaches. For example, Rio et al. (2009) formulated the triggering and closure of deep convection based on parcel’s kinetic energy inside thermals. In our formulation, deep convection is suppressed when the CAPE consumption by shallow convection is strong. Then, after the low troposphere has been sufficiently moistened by shallow convection, air parcel with less dilution from ambient environment (i.e., the lower troposphere) can penetrate to higher altitudes with higher kinetic energy. Thus, the revised closure proposed here might be physically equivalent to some other methods although they use different types of constraints. In the future, we are planning to link the mass flux of deep convection to the statistics of subgrid vertical motion in CLUBB to further improve the consistency between shallow and deep convection.

**Data Availability Statement**

The model outputs and observational data used in this study are publicly available from Zenodo (https://doi.org/10.5281/zenodo.3735572).

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