Improved Precipitation Diurnal Cycle in GFDL Climate Models with Non-Equilibrium Convection

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Key Points:

- A new convective closure is applied to GFDL's CMIP6 climate models AM4 (atmosphere-only) and CM4 (ocean-atmosphere coupled).
- The diurnal cycle of precipitation is significantly improved over land.
- The new closure does not significantly change many aspects of AM4 and CM4's mean state and variability aside from their diurnal precipitation cycles.

Abstract

Most global climate models with convective parameterization have trouble in simulating the observed diurnal cycle of convection. Maximum precipitation usually happens too early during summertime, especially over land. Observational analyses indicate that deep convection over land cannot keep pace with rapid variations in convective available potential energy, which is largely controlled by boundary-layer forcing. In this study, a new convective closure in which shallow and deep convection interact strongly, out of equilibrium, is implemented in atmosphere-only and ocean-atmosphere coupled models. The diurnal cycles of convection in both simulations are significantly improved with small changes to their mean states. The new closure shifts maximum precipitation over land later by about three hours. Compared to satellite observations, the diurnal phase biases are reduced by half. Shallow convection to some extent equilibrates rapid changes in the boundary layer at subdiurnal time scales. Relaxed quasi-equilibrium for convective available potential energy holds in significant measure as a result. Future model improvement will focus on the remaining biases in the diurnal cycle, which may be further reduced by including stochastic entrainment and cold pools.

Key Words: deep convection; shallow convection; diurnal cycle of precipitation.

Plain Language Summary

In this study, we tackled a common challenge in general circulation models concerning the timing of intense rainfall over land during summertime. Many models tend to predict the peak of precipitation too early in the day. To address this, our study introduced a new approach to simulate convection by accounting for the role of shallow convection in stabilizing rapid changes in the atmospheric boundary layer at shorter time scales. This approach delayed maximum precipitation over land by approximately three hours. This adjustment significantly improved the simulated precipitation, aligning them more closely with observations from satellite data. Overall, our research contributes to improving numerical models, bringing them closer to accurately simulating the intricate dynamics of convection and precipitation over land.

1 Introduction

Cumulus parameterization is a key component of general circulation models (GCMs). It connects the intensity and vertical structure of sub-grid scale convection with the grid column mean state. Arakawa and Schubert (1974) argued that convection is in quasi-equilibrium, as measured by cloud work function (CWF), closely related to convective available potential energy (CAPE), with forcing due to large-scale processes like advection and radiation. For time scales of order 1 day and beyond, quasi-equilibrium has proved valid based on both theoretical and experimental studies (Jones & Randall, 2011; Neelin & Yu, 1994; Yano & Plant, 2012). However, processes in the planetary boundary layer (PBL) usually exhibit shorter time scales than processes in the free troposphere (Donner & Phillips, 2003; Raymond & Herman, 2011; Zhang, 2002), challenging quasi-equilibrium on these shorter time scales.

Convective heat sources, moisture sinks, precipitation, and tracer transports are closely related to convective mass fluxes. The vertical profiles of the mass fluxes are typically determined by using plume models, with the base mass flux determined by a closure. Several convection parameterization schemes following the mass-flux framework have been developed to simulate "equilibrium convection" (Arakawa & Schubert, 1974; Donner et al., 2001; Emanuel, 1991; Kain & Fritsch, 1993; Tiedtke, 1989; Zhang & McFarlane, 1995). These schemes show varied abilities in simulating tropical waves and intraseasonal variability (Kim et al., 2011; Lin et al., 2006). The quasi-equilibrium assumption used in the European Centre for Medium-Range Weather Forecasts (ECMWF) Integrated Forecasting System (IFS) is found to reproduce reasonably well the observed mid-latitude synoptic variability, tropical wave spectra and intraseasonal variability (Bechtold et al., 2008; Hirons et al., 2013).

In contrast to equilibrium convection, non-equilibrium convection often holds at time scales of several hours (Davies et al., 2013; Jones & Randall, 2011; Yano & Plant, 2012). Non-equilibrium convection usually occurs in the presence of strong free troposphere forcing or surface heat fluxes. During summertime, surface heat fluxes over land are largely modulated by solar insolation. Variations in surface heat fluxes drive rapid changes in CAPE with which convection is out of equilibrium, as is the related strong diurnal cycle of precipitation over land. Overall, the diurnal cycle of convection over land is characterized by shallow convection in the morning with precipitation associated with deep convection peaking in the late afternoon to early evening (Dai et al., 1999; Tian et al., 2005). It is challenging for state-of-art GCMs to correctly simulate the observed diurnal cycle of precipitation over land. Model-simulated maximum precipitation over land usually happens too early and its phase closely follows the phase of surface fluxes that are largely determined by solar radiation (Dong et al., 2023a; Stratton & Stirling, 2012; Zhao et al., 2018b). While coarse-resolution GCMs struggle in simulating the observed diurnal cycle of convection, the observed diurnal cycle can be simulated reasonably in cloud resolving models with resolutions of order 2.5 km or higher (Petch et al., 2002; Stirling & Stratton, 2012). This study addresses convective parameterization in coarse-resolution GCMs.

Apart from resolution, previous studies have explored various strategies to improve the diurnal cycle of convection, including but not limited to entrainment rates (Del Genio & Wu, 2010; Piriou et al., 2007; Stratton & Stirling, 2012), a prognostic closure accounting for convective memory (Gerard et al., 2009; Pan & Randall, 1998), convective closures including convective inhibition and cold pools (Fletcher & Bretherton, 2010; Mapes, 2000; Rio et al., 2009), a convective triggering function (Xie et al., 2019), and humidity (Fuchs & Raymond, 2007). However, these approaches have not proved to be universally suited for GCMs. An alternate strategy is suggested by observational analyses showing that CAPE variability is controlled by boundary-layer variability, emphasized by Donner and Phillips (2003) and consistent with analyses by Zhang (2002, 2003). In light of this, Bechtold et al. (2014) proposed augmenting the relaxed quasi-equilibrium closure often used for deep convection by a non-equilibrium term accounting for the inability of deep convection to equilibrate CAPE changes resulting from rapid non-convective PBL processes. Improvements in the ECMWF IFS diurnal cycle using this closure encouraged us to explore non-equilibrium convection in the Geophysical Fluid Dynamics Laboratory (GFDL) climate models.

In this study, we show that including non-equilibrium, shallow convection in the closure for deep convection significantly improves the simulation of the diurnal cycle in GFDL AM4 (atmosphere) (Zhao et al., 2018a, 2018b) and CM4 (coupled ocean-atmosphere) (Held et al., 2019) models, while still leaving a notable bias. Section 2 describes the non-equilibrium closure for deep convection. Section 3 shows that changing the deep convection closure in AM4 and CM4 improves diurnal precipitation cycles while changing only slightly other key simulation characteristics. In Section 4 we summarize and suggest possible further ways to improve simulation of the diurnal cycle of convection. In the Appendix, we relate this closure based on relaxed quasi-equilibrium including both deep and shallow (non-equilibrium) convection to a non-equilibrium closure in which CAPE changes are controlled by non-convective PBL processes. We show that shallow convection responds strongly to changes produced by rapid non-convective PBL processes. In the limit where shallow convection equilibrates non-convective PBL processes, the two non-equilibrium closures are identical.

2 Method and Experiments

2.1 Non-Equilibrium Convection

An established method for parameterizing convection is balancing CAPE or CWF changes produced by deep convection with a relaxation of CAPE or CWF changes by non-convective processes, notably large-scale advection, surface fluxes, radiative cooling, and eddy diffusion (Moorthi & Suarez, 1992; Zhang & McFarlane, 1995). Several generations of GFDL climate models have adopted this approach (Anderson et al., 2004; Donner et al., 2011; Zhao et al., 2018a), and all have suffered from large biases in their diurnal cycles of precipitation, mostly over land but also to a lesser degree over ocean. Based on observational analysis showing this balance does not hold at sub-diurnal time scales (Donner & Phillips, 2003; Zhang, 2002, 2003), we generalize this balance to include non-equilibrium convection. The processes changing CAPE are:

$$\left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right) = \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,BL}} + \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,FT}} + \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{deep}} + \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{shal}} \quad (1)$$

Here, the subscript "nc" refers to all non-convective processes. "BL" refers to changes in the PBL, while changes in the overlying free troposphere are denoted by "FT". These tendencies are easily computed in a model using tendencies from the dynamical core and parameterizations for radiative transfer, surface fluxes, and sub-grid diffusion. The subscripts "deep" and "shal" refer to CAPE changes from deep and shallow convection, respectively.

Quasi-equilibrium (Arakawa & Schubert, 1974) has been proposed as a closure for convective mass fluxes, to which are related convective heat sources, moisture sinks, precipitation, and tracer transport. Quasi-equilibrium posits:

$$\left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right) \ll \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,BL}} + \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,FT}} \quad (2)$$

When (2) holds, convective CAPE tendencies balance non-convective CAPE tendencies. The convective mass fluxes are then obtained from the convective CAPE tendencies. Observations show that quasi-equilibrium is a reasonable approximation as diurnal time scales are approached but does not hold for sub-diurnal time scales (Donner and Phillips, 2003, Fig. 2). A recourse is to relax the non-convective terms in (1):

$$\left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,BL}} + \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{nc,FT}} = \frac{\mathbf{CAPE-CAPE_0}}{\boldsymbol{\tau}} \quad (3)$$

In Eq. (3), $\mathrm{CAPE_0}$ is a reference value toward which CAPE, having evolved by non-convective processes since its last adjustment due to convection (the beginning of a model time step in AM4), relaxes over an extended time τ. The value of τ is typically tuned to multiple hours in models. Based on field experiments, Donner and Phillips (2003) estimated $\mathrm{CAPE_0}$ to be 0 J $\mathrm{kg^{-1}}$, and τ to be 6.5 hours. Zhao et al. (2018b) tuned these to 10 J $\mathrm{kg^{-1}}$ and 8 hours, respectively, in AM4.

The relaxed CAPE tendency is then assumed to be in quasi-equilibrium with convective tendencies, implying:

$$\left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{deep}} = -\frac{\mathbf{CAPE-CAPE_0}}{\boldsymbol{\tau}} - \left(\frac{\partial \mathbf{CAPE}}{\partial \mathbf{t}}\right)_{\mathbf{shal}} \quad (4)$$

In GFDL AM2, AM3, and AM4, the contribution from non-equilibrium shallow convection has not been included in the determination of the deep convective CAPE tendency and mass fluxes. Rather, only deep convection has been assumed to be in relaxed equilibrium with non-convective tendencies. Shallow convective tendencies are likely to be non-equilibrium and tied closely to the evolution of the PBL. In GFDL AM3 and AM4, shallow convective base mass fluxes $\mathrm{M_{shal}}$ follow Bretherton et al. (2004):

$$\mathrm{M_{shal}} \propto \sqrt{c_1 \mathrm{TKE}}\, \exp\left(-\frac{c_2 \mathrm{CIN}}{\mathrm{TKE}}\right) \quad (5)$$

where TKE denotes PBL turbulent kinetic energy; CIN, convective inhibition; and c_1 and c_2 denote constants. $\left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{shal}}$ in Eq. (4) depends strongly on $\mathrm{M_{shal}}$. We hypothesize that non-equilibrium shallow convection and its diurnal cycle are important in the relaxed balance in Eq. (4). By removing enthalpy and moisture from the boundary layer, shallow convection generally reduces CAPE (Figs. A1a,b and A2a,b), though infrequent exceptions of small magnitude can occur, for example, for a PBL cold and dry relative to the free troposphere above.

2.2 Observations and Implementation in AM4 and CM4

For observations, we use the NASA Global Precipitation Measurement (GPM) Integrated Multisatellite Retrievals for GPM (IMERG) (Huffman et al., 2015) to evaluate the models' ability in simulating the diurnal cycle of precipitation. IMERG provides a global-gridded product with 0.1° horizontal resolution and 30-min frequency. For a direct comparison with AM4 and CM4 outputs, the IMERG product is re-gridded to 1.0° latitude × 1.25° longitude resolution. In addition, Global Precipitation Climatology Project (GPCP v3.2)(Huffman et al., 2022) is used to evaluate the mean precipitation climatology. The high-frequency IMERG observations are well-suited for evaluating the diurnal cycle, while the combination of surface and satellite observations in GPCP are better suited for evaluation of long-term means. Radiation at the top of the atmosphere from CERES EBAF Edition 4.1 (Loeb et al., 2018) is used to evaluate the models' energy budget. Monthly ERA5 data (Hersbach et al., 2020) is used for temperature comparisons .

AM4 with prescribed sea surface temperatures (SSTs) and sea ice, and pre-industrial (PI) CM4 ocean-atmosphere coupled model are used to explore the effects of non-equilibrium convection. Detailed descriptions of AM4 can be found in Zhao et al. (2018a, 2018b), and CM4 is documented in Held et al. (2019). In both AM4 and CM4, the atmosphere component has 33 vertical levels with a horizontal resolution of approximately 100 km. The PI configuration of CM4 is chosen because the PI period is in global energy balance (unlike present day), rendering an evaluation of PI model drift and energy cycle more straightforward than for an evolving historical simulation, which would, in any case, need to start from a realistic, balanced PI simulation.

We implement a non-equilibrium closure based on Eq. (4) including the CAPE tendency due to shallow convection and examine its impact on the diurnal cycle of precipitation in AM4 and CM4 in subsequent sections. Default configurations of the standard AM4 documented in Zhao et al. (2018a, 2018b) used a relaxed quasi-equilibrium closure where CWF changes induced by deep convection are balanced by a relaxation of CWF changes due to non-convective processes. Here, we implement relaxed quasi-equilibrium using CAPE instead of CWF (Eqs. (3) and (4)). Doing so facilitates computations of changes with respect to state changes, which are easily done for CAPE but require additional convective plume calculations for CWF. In this study, the CWF based quasi-equilibrium closure is referred to as standard AM4. The CAPE based non-equilibrium closure based on Eq. (4) including the CAPE tendency due to shallow convection is referred to as relax QE D+S. Two parameters are retuned in relax QE D+S from those in standard AM4. The retuning simulates Earth's energy imbalance in the relax QE D+S AMIP integration (2001-2014) to a value of 1.08 W/m^2, close to CERES EBAF (0.88 W/m^2). The retuned parameters (1) increase ice fall speeds by 10% and (2) change cloud-top entrainment in stratiform clouds to increase absorbed shortwave radiation (SWABS) at top of the atmosphere (TOA). For CM4, the PI-Control simulation uses the quasi-equilibrium closure based on CWF. The PI simulation using the CAPE based non-equilibrium closure based on Eq. (4), with the same two retuned parameters, is referred to as PI relax QE D+S. Table 1 lists the experiments and observational products.

Bechtold et al. (2014) provide an alternate interpretation for non-equilibrium convection. Its incorporation in the ECMWF IFS improved its diurnal precipitation cycle. The Appendix shows the approach here, when implemented in AM4, is closely related.

Table 1 Experiments and observations.

Data Source	**Experiment Name**	**Resolution**	**Time**
AM4	Standard AM4	1.0° lat x 1.25° lon	Prescribed SSTs (Rayner et al., 2003) from 1979-2014
	relax QE D+S		
CM4	PI-Control		200 years
	PI relax QE D+S		
Observation	IMERG	0.1° original and re-gridded to 1.0° lat, 1.25° lon	2001-2014
	GPCP v3.2	2.5° lat x 2.5° lon	2001-2014
	CERES EBAF Edition 4.1	1.0° lat x 1.0° lon	2001-2014
	ERA5	0.25° lat x 0.25° lon	2001-2014

3. Results

3.1 Diurnal Precipitation Cycle in AM4

To quantify the diurnal amplitude and phase of precipitation, the model outputs and satellite observations are first composited in hourly bins. A Fourier analysis is then applied to the hourly-binned data to get its first harmonic. Figure 1 compares the diurnal amplitude of boreal summer precipitation between IMERG, standard AM4 and relax QE D+S. The observed spatial distribution of diurnal amplitude shows notable peaks over central America, northern Amazon, central Africa, south Asia, and Maritime Continent. Overall, the pattern of the diurnal amplitude is similar for standard AM4 and relax QE D+S, indicating that the modified closure has little impact on the diurnal amplitude of precipitation.

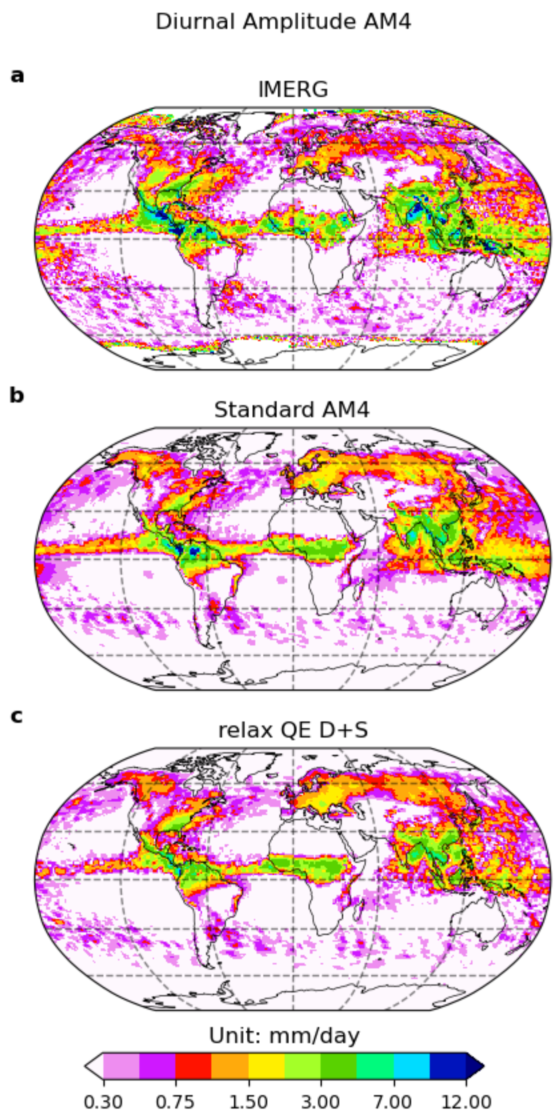

Figure 1 Diurnal amplitude of precipitation (unit: mm day^{-1}) during boreal summer (June-August) for (a) IMERG, (b) Standard AM4, and (c) relax QE D+S.

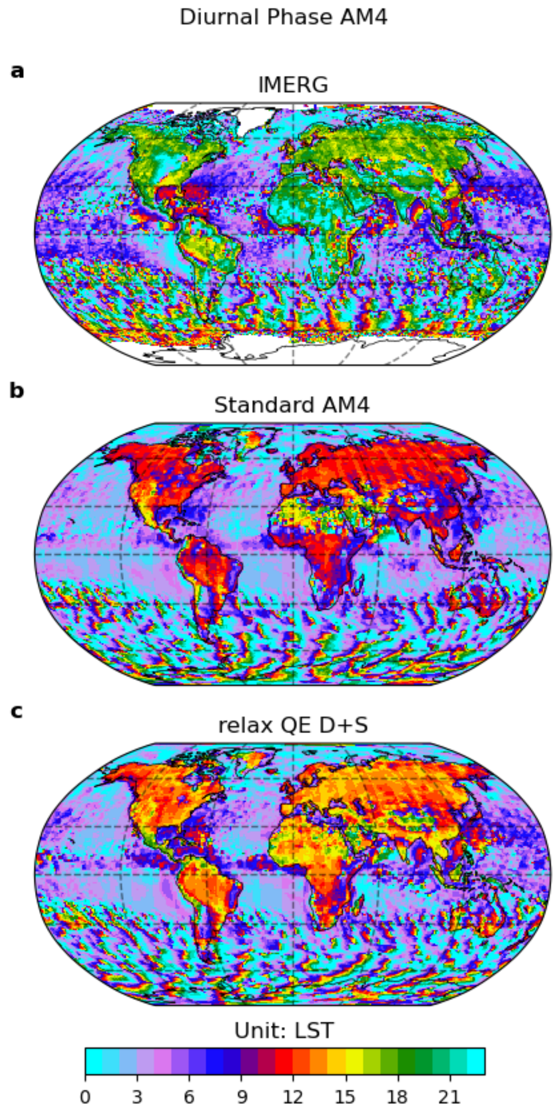

Figure 2 Diurnal phase (unit: LST hour) during boreal summer (June-August) for (a) IMERG, (b) Standard AM4, and (c) relax QE D+S.

The diurnal phase (local solar time, LST) is shown in Figure 2. For IMERG, maximum precipitation over land occurs mostly from late afternoon to early evening. Over tropical oceans, maximum precipitation predominantly occurs in early morning, although certain regions, such as the Gulf of Mexico and coastal regions off central America and south Asia, exhibit peak precipitation in late morning to local noon (Figure 2a). The largest phase errors in standard AM4 are generally over land. Maximum precipitation over land in standard AM4 occurs around local noon (Figure 2b), which is too early compared to IMERG and is a known issue in AM4 (Zhao et al., 2018a). In relax QE D+S, the diurnal cycle over land is delayed by about three hours relative to standard AM4 (Figure 2c), which means the phase biases between standard AM4 and IMERG are reduced by half. The significant improvement indicates that shallow convection strongly mediates rapid variations in CAPE at subdiurnal time scales consistent with observational analyses shown in Donner and Phillips (2003). The retuning of relax QE D+S discussed in Section 2 does not change Figure *2* noticeably (not shown).

Next, we focus on the diurnal cycle over the contiguous United States during boreal summer (CONUS; Figure 3). The observed diurnal amplitude shows peaks eastward of the Rocky Mountains area and over the southeast CONUS and the corresponding coastal regions (Figure 3a). Standard AM4 reasonably reproduces the diurnal amplitude over the southeast CONUS but underestimates the diurnal amplitude east of the Rocky Mountains (Figure 3b). Compared to standard AM4, relax QE D+S shows a similar spatial pattern of the diurnal amplitude (Figure 3c), which again indicates that the revised closure has little impact on the spatial pattern of the diurnal amplitude.

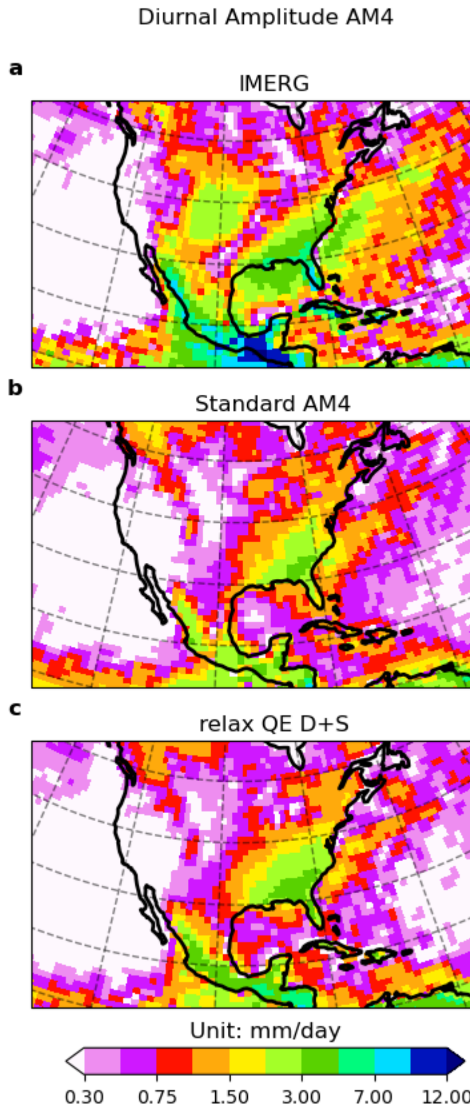

Figure 3 Same as Figure 1 but over the contiguous United States.

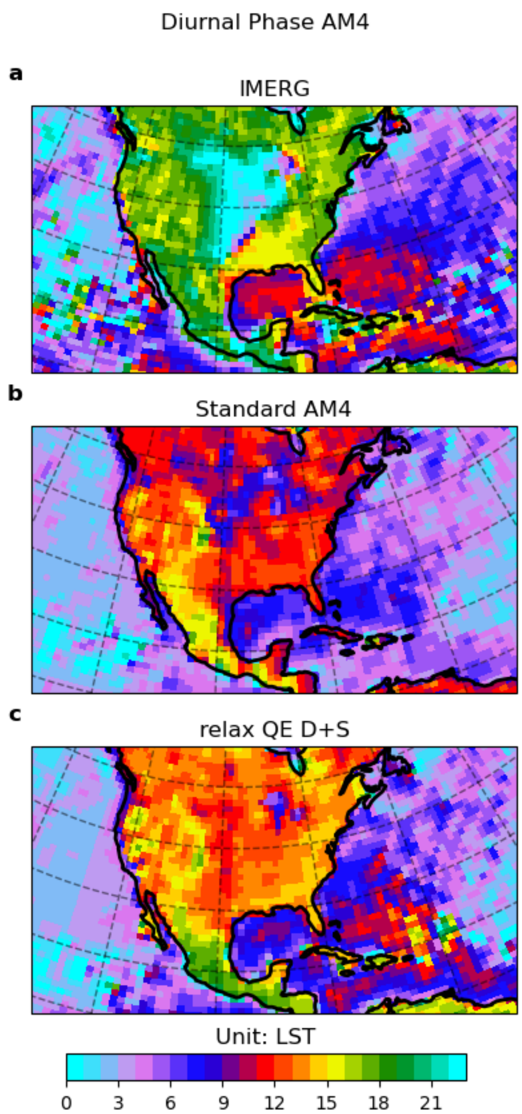

Figure 4 Same as Figure 2 but over the contiguous United States.

For the diurnal phase, the observed spatial pattern in IMERG is characterized predominantly by late afternoon convection over the western and eastern United States, and nighttime convection over Central Plains due to mesoscale convective systems (Figure 4a), which is consistent with previous results (Bechtold et al., 2014; Dai et al., 1999; Dong et al., 2023b; Tian et al., 2005; Watters et al., 2021). The diurnal phase in standard AM4 is significantly biased from IMERG with maximum precipitation occurring around local noon in most parts of CONUS (Figure 4b). By including the modulation of CAPE by shallow convection in the closure for deep convection, relax QE D+S shifts the diurnal cycle later by about three hours. The diurnal phase is also improved over the surrounding ocean, especially the Gulf of

Mexico and coastal regions off the southeast CONUS (Figure 4c). Bechtold et al. (2014) showed similar improvement of the diurnal cycle, globally and over CONUS, with non-equilibrium convection implemented in the ECMWF IFS (their Figures 2 and 4), also with small changes in CONUS amplitudes (their Figure 3). This suggests the closure based on Eq. (4) may be related to the non-equilibrium closure implemented in Bechtold et al. (2014). The Appendix discusses this in detail.

The improved diurnal cycle is further confirmed by domain-averaged precipitation without projection onto a single harmonic. The overall improvement over land areas holds globally (Figure 5a) and at regional scales (Figure 5b-f). Compared to standard AM4, relax QE D+S weakens convective activity around local noon time but enhances it during late afternoon and early evening. The extent to which the diurnal cycle improves varies with location. For example, the diurnal cycle over the western United States shows less improvement than over the eastern United States. One possible reason is that AM4 with coarse horizontal resolutions is not able to simulate well mesoscale connective systems east of the Rocky Mountains indicated by Figure 3. The improved diurnal cycles over Asia, Europe, and Africa are consistent and comparable to those reported by Bechtold et al. (2014) (their Figure 5). Although the diurnal cycle of convection in relax QE D+S is significantly improved compared to the observations, the late-night precipitation in relax QE D+S is still underestimated. As mentioned by Bechtold et al. (2014), the late-night precipitation deficit may relate to the missing representation of surface cold pools and upper-level mesoscale lifting.

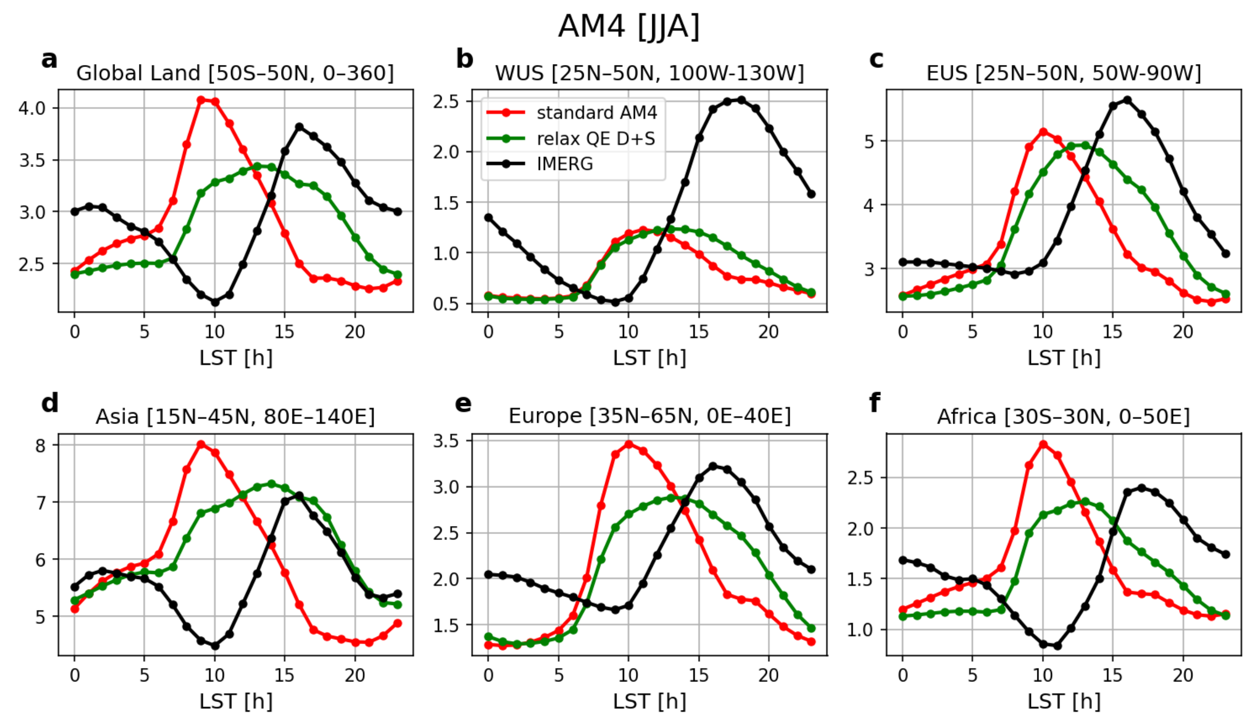

Figure 5 Domain averaged hourly land precipitation (units: mm day^{-1}) averaged from June to August for Standard AM4 (red lines), relax QE D+S (green lines), and IMERG (black lines) over (a) Global Land, (b) Western United States, (c) Eastern United States, (d) Asia, (e) Europe, and (f) Africa.

The total precipitation in AM4 consists of contributions from convective and large-scale parameterizations. We decompose the total precipitation into the convective and large-scale components (Figure 6). Overall, the convective precipitation peaks a few hours later in relax QE D+S than in standard AM4. The large-scale component is only slightly affected by relax QE D+S. The improved diurnal cycle of total precipitation shown in Figure 5 is mostly contributed by a better representation of the convective precipitation.

The phase changes in Figures 1-5 are strongly dominated by including the $\left(\frac{\partial CAPE}{\partial t}\right)_{shal}$ term in Eq. (4). There are only small contributions to the phase changes from using relaxed CAPE instead of relaxed CWF in Eqs. (3) and (4) and from tuning for relax QE D+S.

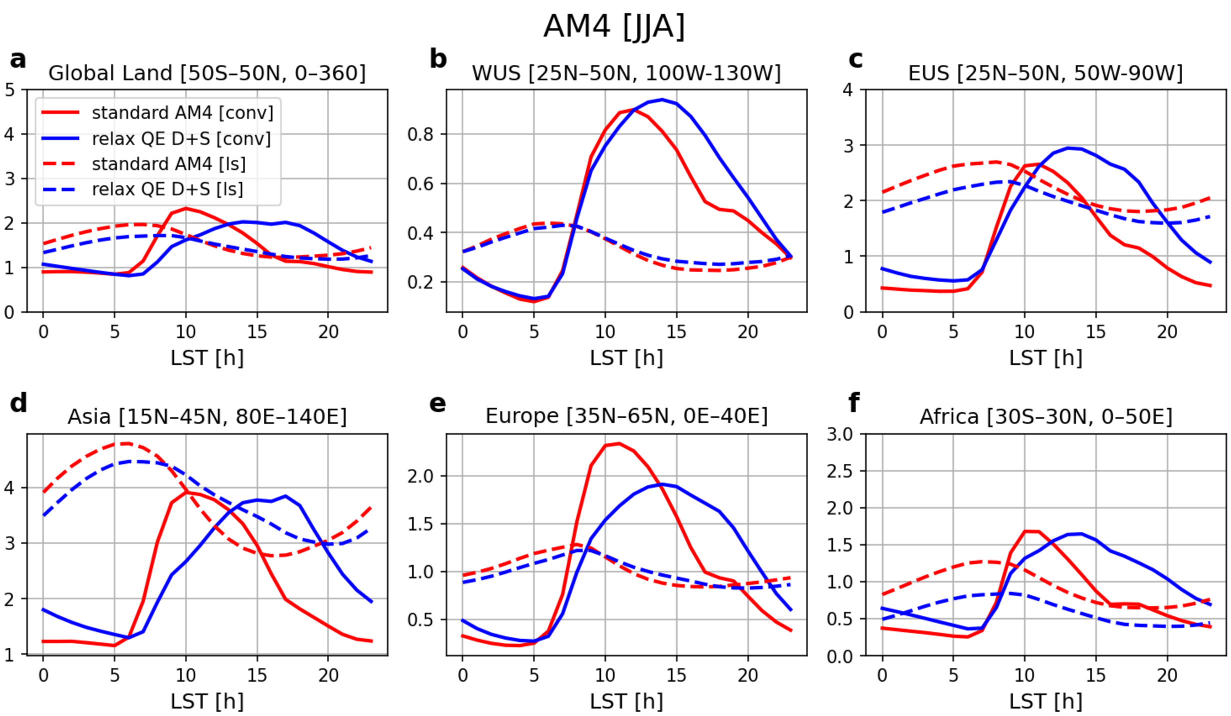

Figure 6 As in Figure 5 except showing a decomposition of the total precipitation into parameterized convection (solid lines) and large scale (dash lines) over the same regions.

3.2 Mean State and Tropical Variability in AM4

Figure 7 compares zonal mean temperature between ERA5, standard AM4 and relax QE D+S. Overall, both standard AM4 and relax QE D+S show colder troposphere but warmer stratosphere than ERA5 (Figure 7b and Figure 7c). In the tropics, standard AM4 is mostly warmer in the troposphere but colder in the stratosphere than relax QE D+S (Figure 7d). Temperature changes from replacing CWF in Standard AM4 with CAPE in the relaxation terms in Eqs. (3) and (4) are similar in pattern, larger in magnitude, but opposite in sign to those in Figure 7d. Temperature changes associated with the $\left(\frac{\partial CAPE}{\partial t}\right)_{shal}$ term in Eq. (4) largely oppose those from replacing CWF with CAPE in the relaxation, as do tuning changes to obtain relax QE D+S. (These intermediate changes are not shown.)

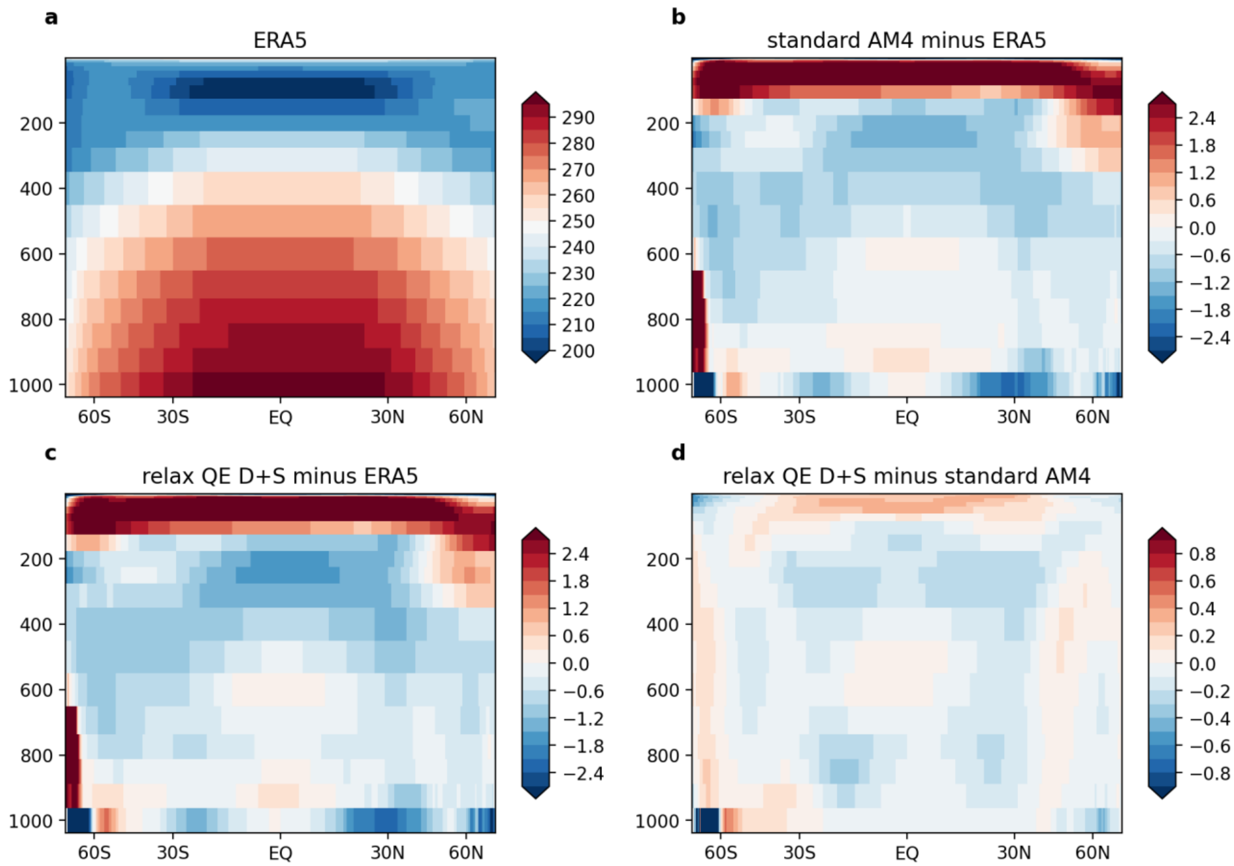

Figure 7 2001-2014 annual, zonal mean air temperature (units: K) for (a) ERA5, (b) standard AM4 minus ERA5, (c) relax QE D+S minus ERA5, and (d) relax QE D+S minus standard AM4.

Outgoing longwave radiation (OLR) in standard AM4 and relax QE D+S are overall similar to CERES-EBAF observations with some regional differences (Figure 8). Compared to standard AM4, relax QE D+S shows enhanced OLR over tropical ocean but reduced OLR over land, although the global mean OLR in relax QE D+S is slightly greater than that in standard AM4 (Figure 8d). The absorbed shortwave radiation (SWABS) patterns are also similar between standard AM4 and relax QE D+S but relax QE D+S also has slightly greater global mean SWABS than standard AM4 (Figure 9). As a result, the energy imbalance at TOA does not change much between standard AM4 and relaxed QE D+S. In addition, standard AM4 and relax QE D+S show similar spatial patterns of mean precipitation, with global mean precipitation in both experiments greater than GPCP (Figure 10). The precipitation intensity distribution is almost identical between standard AM4 and relax QE D+S in the tropics, northern hemisphere mid-latitudes, and southern hemisphere mid-latitudes (Figure 11). The precipitation intensity distribution also remains nearly unchanged between standard AM4 and relax QE D+S when land

and ocean are considered separately (not shown), indicating that the new non-equilibrium closure does not affect the overall precipitation statistics. Moreover, the tropical wave spectra analyses show only small changes between standard AM4 and relax QE D+S (Figure 12).

We also conducted idealized experiments with uniform SST warming by 2K to examine the 'Cess sensitivity' (Cess et al., 1990), which is computed by examining the change in net TOA flux in response to uniform 2K ocean warming. We found that relax QE D+S yields a Cess sensitivity of 0.54-0.55 K W^{-1} m^2, which is very similar to the value (0.56-0.57 K W^{-1} m^2) reported by Zhao et al. (2018a), indicating that the implementation of the revised closure has little impact on the model's global mean radiative response to uniform SST warming.

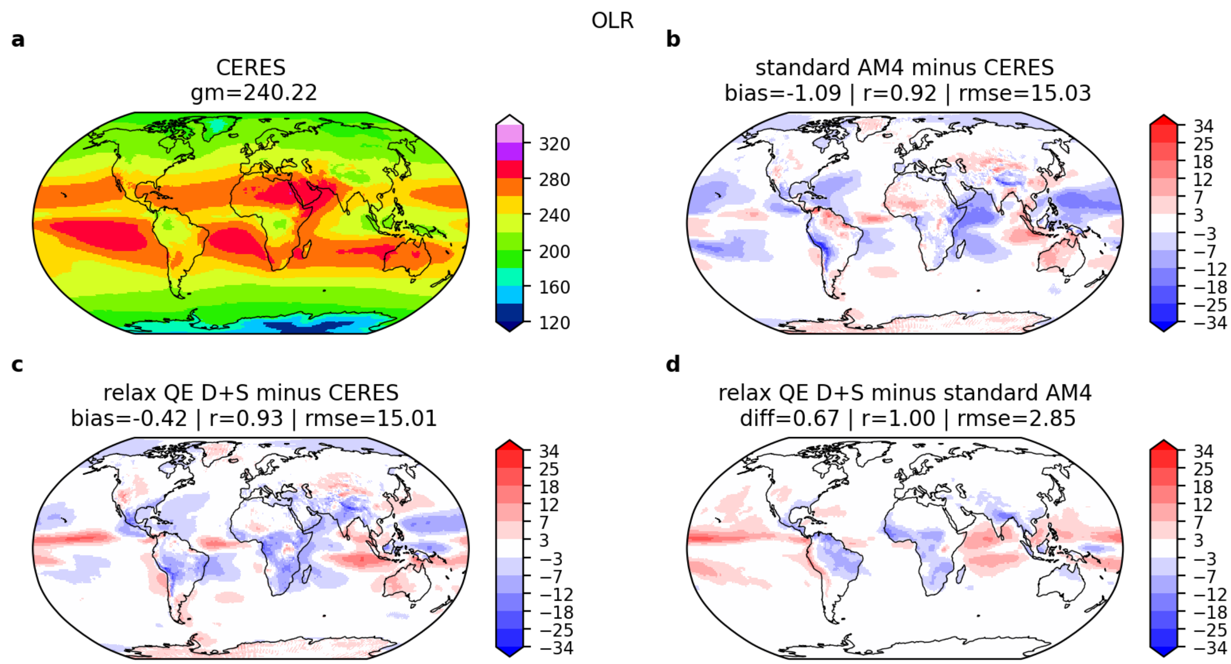

Figure 8 2001-2014 annual mean OLR (units: W m^{-2}) for (a) CERES EBAF, (b) standard AM4 minus CERES EBAF, (c) relax QE D+S minus CERES EBAF, and (d) relax QE D+S minus standard AM4. Bias (or difference), correlation coefficient, and root mean square error are listed.

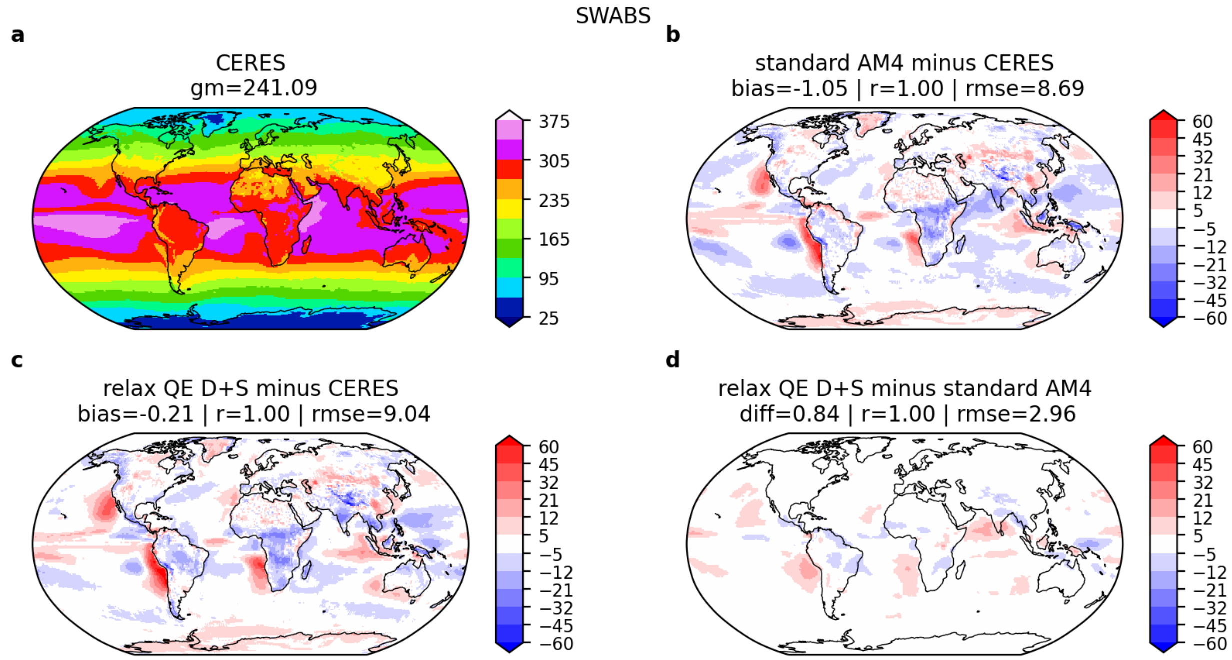

Figure 9 Same as Figure 8 but for SW radiation at TOA (units: W m^{-2}).

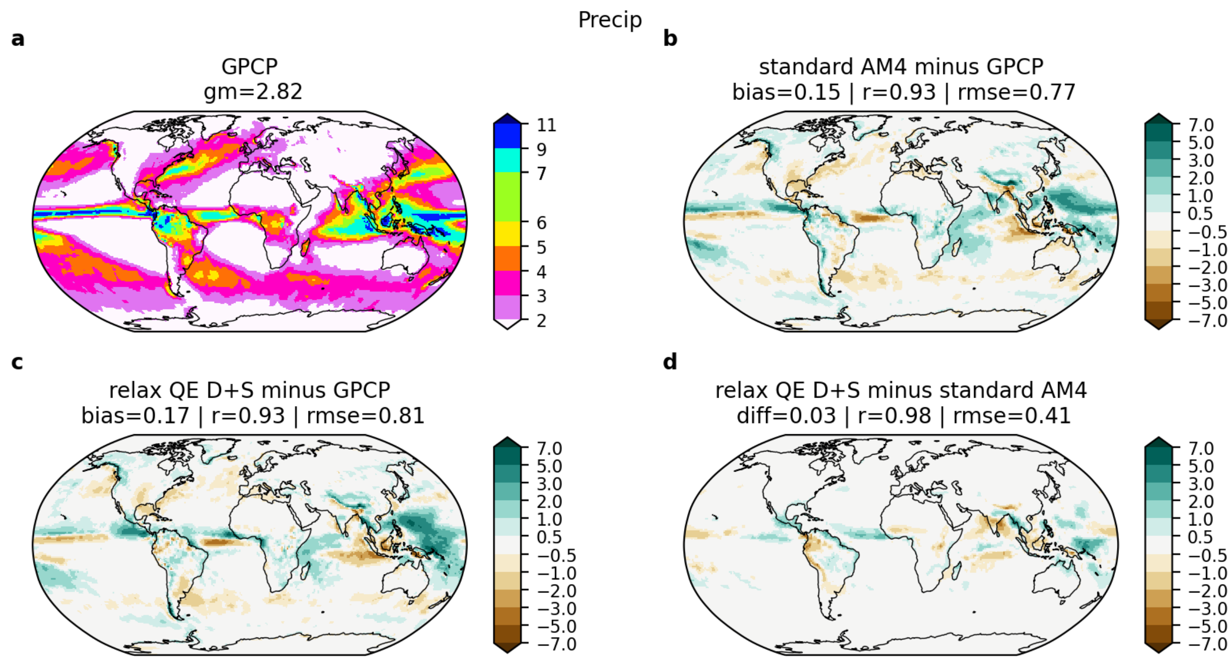

Figure 10 2001-2014 annual mean precipitation (units: mm day^{-1}) for (a) GPCP, (b) relax QE D+S minus GPCP, (c) relax QE D+S minus GPCP, and (d) relax QE D+S minus standard AM4. Bias (or difference), correlation coefficient, and root mean square error are listed.

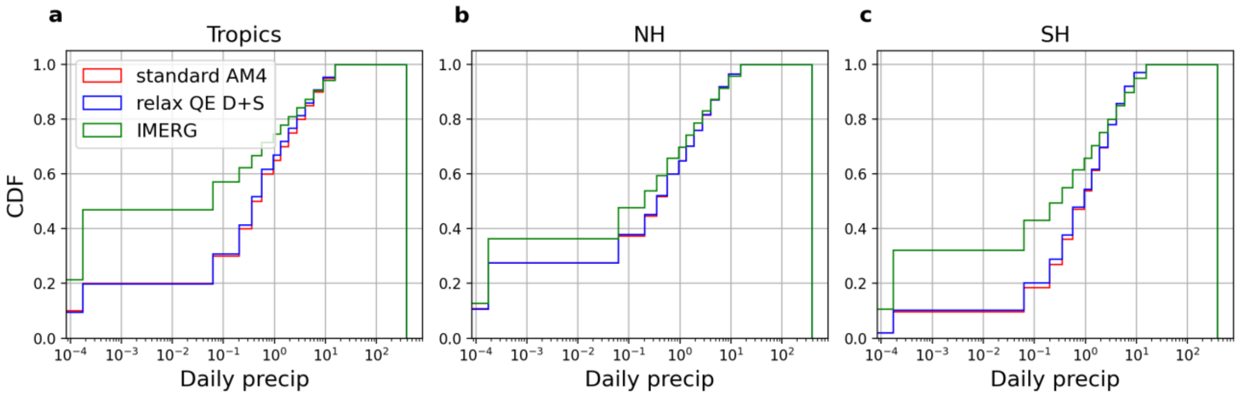

Figure 11 Cumulative distribution of precipitation over (a) Tropics (30°S-30°N), (b) Northern Hemisphere Mid-latitude (30°N-60°N), and (c) Southern Hemisphere Mid-latitude (60°S-30°S). Red curves for standard AM4, Blue for relax QE D+S, and Green for IMERG.

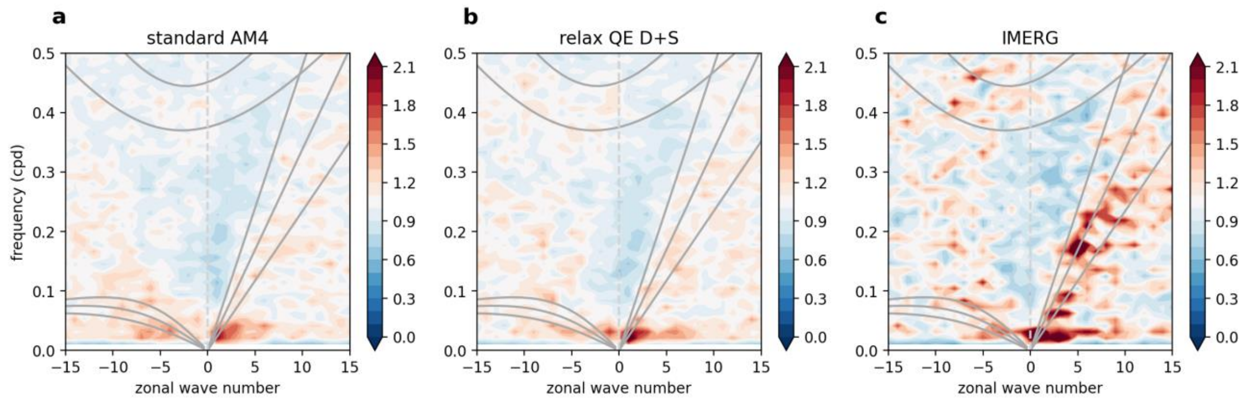

Figure 12 Wave spectra from (a) standard AM4, (b) relax QE D+S and (c) IMERG.

3.3 Non-Equilibrium Convection in CM4

Model configurations that are apparently successful in uncoupled simulations do not always withstand coupling without deterioration of major simulation characteristics. In this section, we briefly compare results between PI-Control and PI relax QE D+S using CM4. PI-Control shows stable integrations with a reasonable TOA energy balance reported by Held et al. (2019). Noting that the 2001-2014 CERES-EBAF TOA SWABS is 241.0 W m^{-2} and OLR is 240.2 W m^{-2}, PI relax QE D+S SWABS and OLR may be closer to their PI values than in PI-Control (Figure *13*).

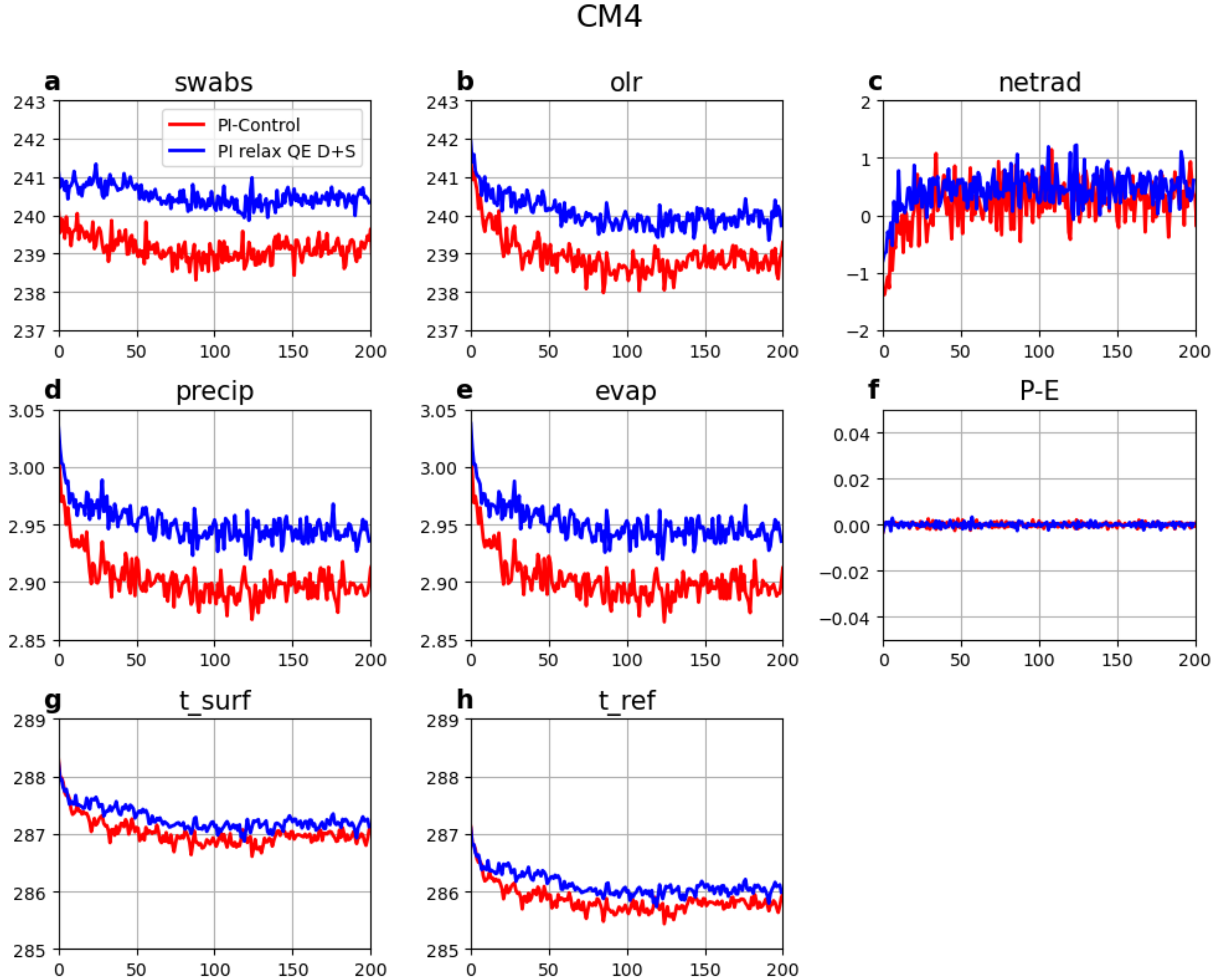

Figure 13 Global, annual-mean (a) SWABS (i.e. SW downward at TOA minus SW upward at TOA; W m^{-2}); (b) OLR (W m^{-2}), (c) net radiation (W m^{-2}), (d) precipitation (mm day^{-1}), (e) evaporation (mm day^{-1}), (f) precipitation minus evaporation (mm day^{-1}), (g) surface temperature (K), and (h) surface air temperature (i.e. temperature at 2m; K). Red solid lines are for PI-Control, while blue solid curves are for PI relax QE D+S.

Figure 14 shows that the improvements in the diurnal cycle of precipitation with relax QE D+S in AM4 are also evident in CM4. Biases in the diurnal precipitation cycle are reduced by several hours over most land areas, with improvements over many ocean regions as well.

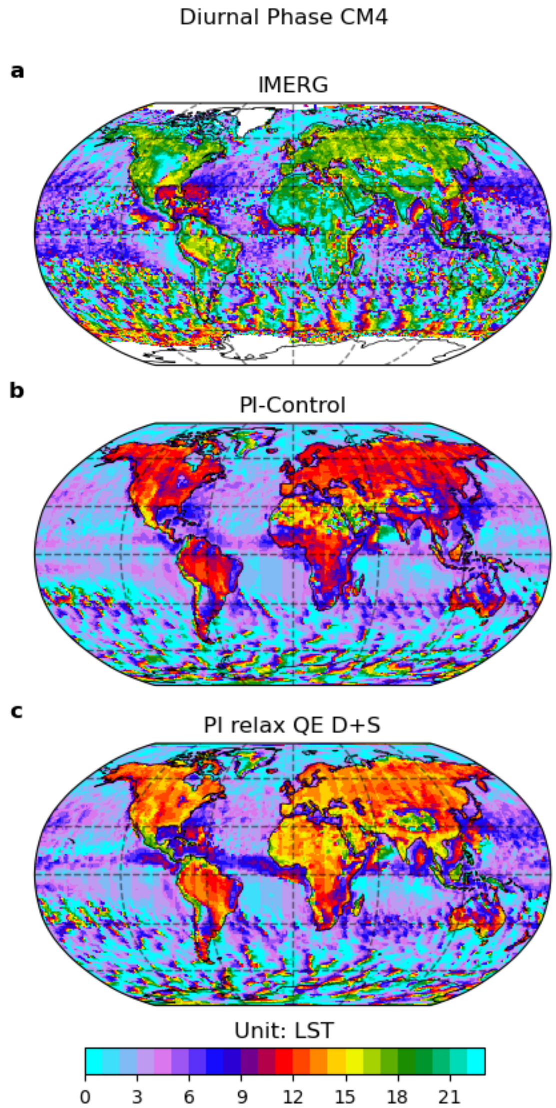

Figure 14 Same as Figure 2 but for the ocean-atmosphere coupled simulations using CM4.

4 Summary

In this study, a new convective closure on the mass flux for deep convection is implemented in GFDL climate models AM4 (atmosphere-only simulations) and CM4 (ocean-atmosphere coupled simulations). The new convective closure is CAPE based and assumes a relaxed quasi-equilibrium between deep convection, shallow convection and non-convective processes (large-scale advection, surface fluxes, radiative transfer, eddy diffusion).

Over land, this new closure improves the diurnal cycle of precipitation with the diurnal phase biases reduced by half. The diurnal cycle of precipitation over ocean is also improved to some extent. This indicates that shallow convection is key to delaying deep convection over land where boundary layer properties (to which shallow convection responds) change rapidly and plays a nonnegligible role in the presence of relatively slow-varying surface conditions like over some ocean regions.

Comparisons of the performance in AM4 and CM4 with the new convective closure against IMERG observations indicate that there is still room for improvement in simulating the diurnal cycle of precipitation during boreal summer. Future work may incorporate stochastic entrainment and effects from cold cools. One potential approach for parameterizing cold pools is to make the entrainment rate dependent on the mass flux from downdrafts and surface heat flux as illustrated in Suselj et al. (2019).

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Data Availability Statement

The standard AM4 and CM4 code can be found at https://data1.gfdl.noaa.gov/nomads/forms/am4.0/ and https://zenodo.org/records/3339397. The code for non-equilibrium convection can be found at https://zenodo.org/records/10709307. IMERG datasets can be downloaded from https://disc.gsfc.nasa.gov/datasets/GPM_3IMERGHH_07/summary?keywords=%22IMERG%20 final%. GPCP data can be found at https://disc.gsfc.nasa.gov/datasets/GPCPMON_3.2/summary. CERES EBAF data can be found at https://ceres.larc.nasa.gov/data/. ERA5 data can be found at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels?tab=overview.

Appendix

Donner and Phillips (2003) (their Figs. 1 and 2) show that CAPE is not in quasi-equilibrium at sub-diurnal time scales but instead changes following temperature and moisture changes in the planetary boundary layer:

$$\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}} \cong (\partial_{\mathrm{t}}\mathrm{CAPE})_{\mathrm{BL}} \leq \left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{nc}} \quad \text{(A1)}$$

$(\partial_t \mathrm{CAPE})_{\mathrm{BL}}$is the total change in CAPE due to changes in PBL temperature and moisture by all physical processes, including convection. In (A1), the evolution of CAPE is controlled primarily by the PBL (Donner and Phillips (2003), their Fig. 1).

The failure of quasi-equilibrium,

$$\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}} \ll \left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{nc}} \text{ (A2)}$$

to hold at sub-diurnal time scales has been compensated for by relaxing non-convective CAPE changes toward a threshold CAPE over a time of multiple hours, Eq. (3). Donner and Phillips (2003) (their Fig. 5) show that convective heat sources obtained by requiring convective tendencies to balance relaxed non-convective tendencies still agree rather poorly with convective heat sources diagnosed from observations. Accordingly, relaxed quasi-equilibrium for deep convection is augmented by a non-equilibrium term:

$$\left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{deep}} = -\frac{\mathrm{CAPE}-\mathrm{CAPE}_0}{\tau} + \left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{non-equil}} \text{ (A3)}$$

Analysis of non-convective tendencies of temperature and water vapor mixing ratio in Donner and Phillips (2003) (their Fig. 7) suggests:

$$\left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{non-equil}} = \alpha \left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{nc,BL}} \text{ (A4)}$$

where α is a proportionality constant between 0 and 1, tending towards 1 with increasing non-convective PBL CAPE tendencies. (A4) is very similar to Eq. (14) in Bechtold et al. (2014).

Note that, for $\alpha = 1$, the non-equilibrium convective closures Eq. (4), (A3) and (A4) are identical if shallow CAPE tendencies are equal to the negative of the CAPE tendencies from non-convective PBL processes. Physically, this corresponds to shallow convection equilibrating non-convective PBL CAPE changes. In relax QE D+S, this holds to a limited degree. Figures A1 and A2 show global patterns of instantaneous CAPE tendencies from shallow convection, non-convective PBL processes, and the sum of these tendencies composited four times daily. Non-convective CAPE tendencies are generally positive, driven by surface heat and moisture fluxes, which respond strongly over land to the diurnal cycle in surface shortwave absorbed radiation. Less frequently, this tendency can be negative and of smaller magnitude, for example, during nocturnal PBL cooling, e.g., Figure A1c over Mexico. CAPE tendencies from shallow convection do not fully equilibrate non-convective PBL CAPE changes, and this is especially true over land when surface heating is most pronounced during the diurnal cycle (Figs. A1e,f and A2e,f).

A CAPE based non-equilibrium closure based on Eq. (A3) and (A4) with $\alpha = 1$ is referred to as D PBL CAPE Control. Figures A3a-c show the base mass fluxes from a relax QE D+S integration. The base mass fluxes are also calculated from a diagnostic call to D PBL CAPE Control closure in this integration. Figures A3d-f show the differences between the base mass fluxes diagnosed from the D PBL CAPE Control closure and those from relax QE D+S. In areas where relax QE D+S base mass fluxes are large and shallow CAPE tendencies do not equilibrate non-convective PBL CAPE tendencies in Figures A1 and A2, the diagnosed D PBL CAPE Control base mass fluxes are less in Figure A3. In regions with smaller relax QE D+S base mass fluxes, the D PBL CAPE Control mass fluxes can be larger. These regions do not support intense convection, and a wider range non-convective PBL CAPE tendencies likely occur. The sum $\left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{shal}} + \left(\frac{\partial \mathrm{CAPE}}{\partial \mathrm{t}}\right)_{\mathrm{nc,BL}}$ in these regions is often small (Figures. A1e,f and A2e,f).

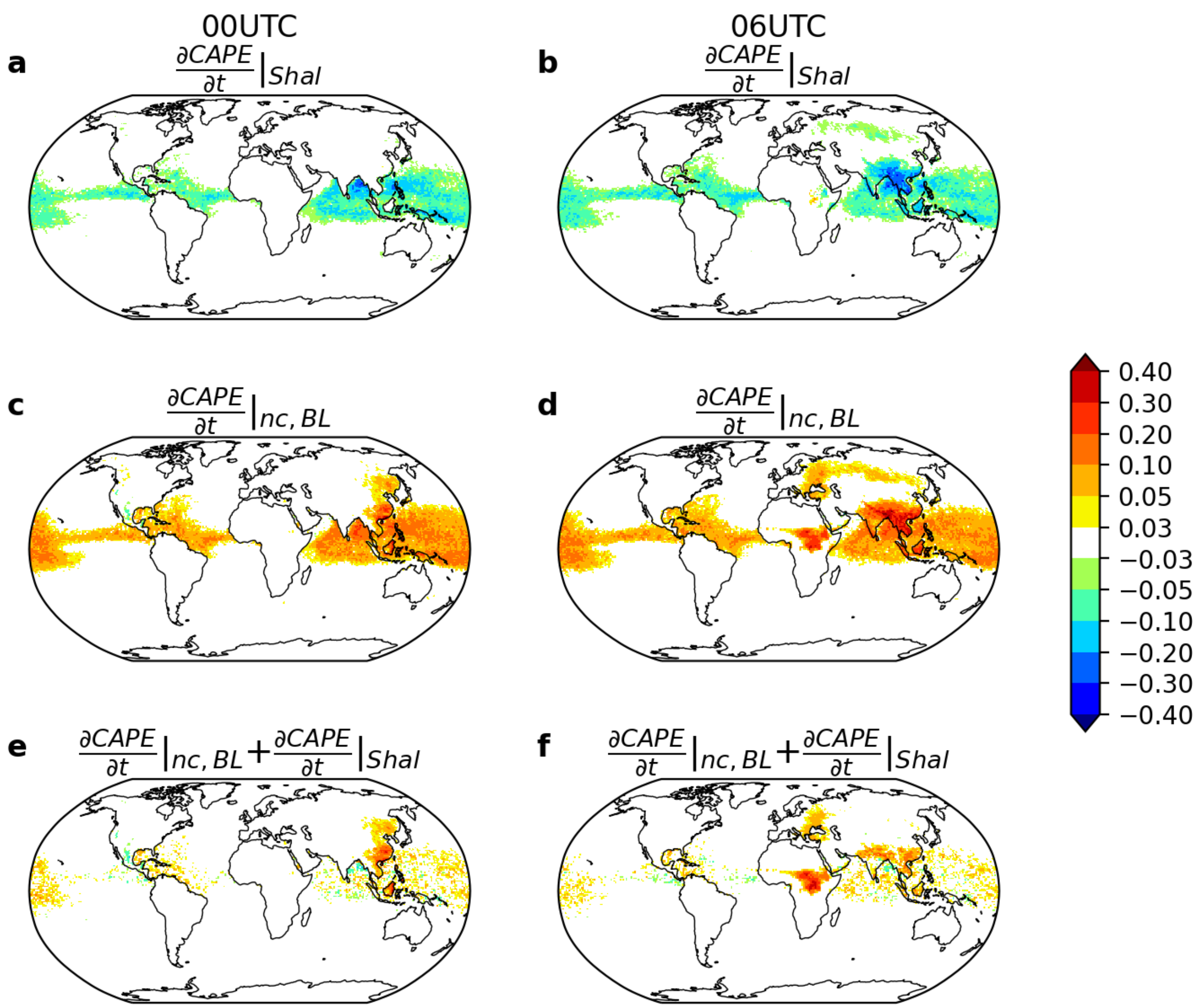

Figure A1 CAPE tendency due to shallow convection (panels a and b), CAPE tendency due to non-convective PBL processes (panels c and d), and the sum of them (panels e and f). The units for CAPE tendency are J kg^{-1} sec^{-1}. The left column is at 00 UTC and the right column is at 06 UTC.

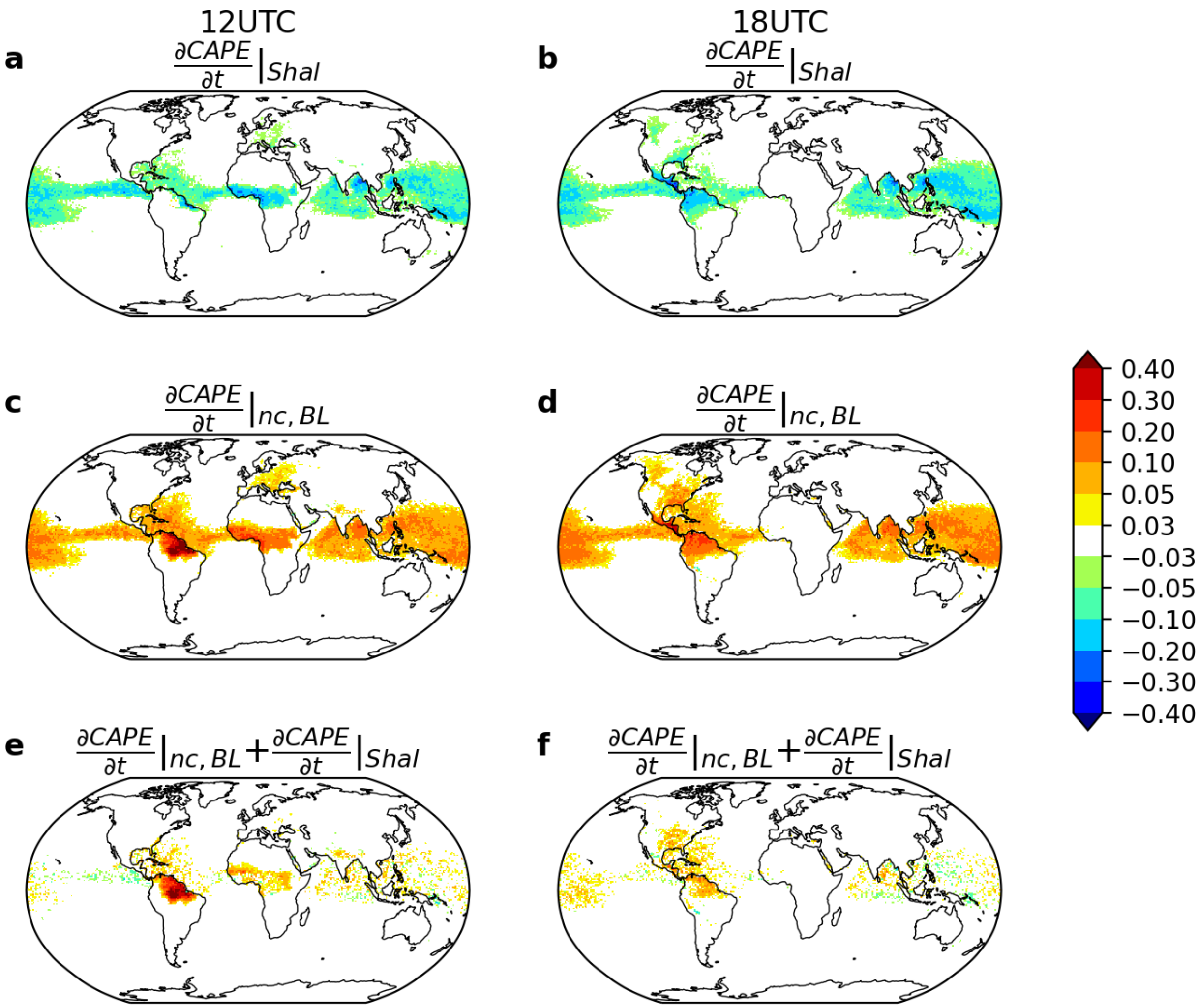

Figure A2 Same as Figure A1 but the left column is at 12 UTC and the right column is at 18 UTC. The units for CAPE tendency are J kg^{-1} sec^{-1}.

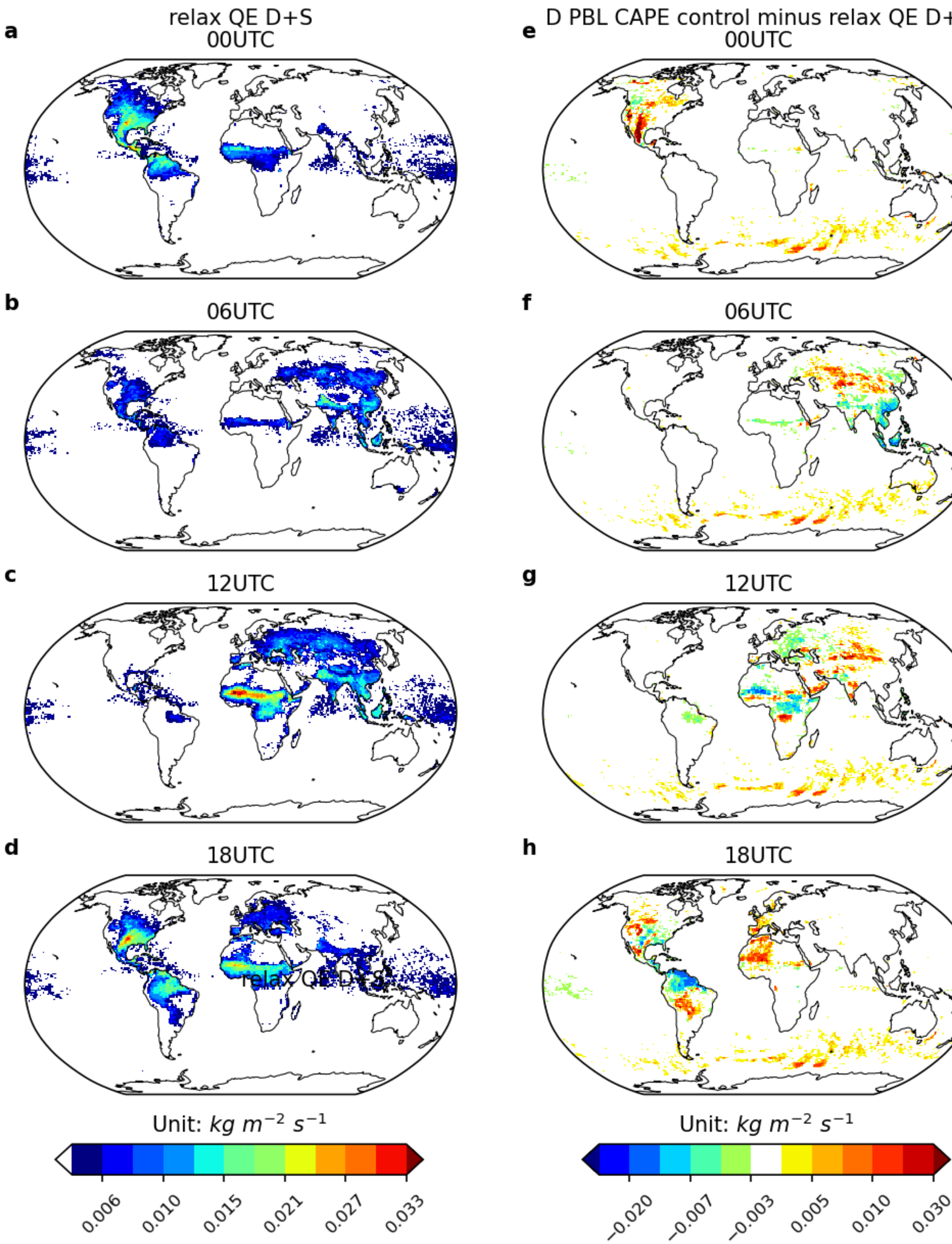

Figure A3 Cloud base mass flux computed from relax QE D+S (panels a-d), and the difference of cloud base mass flux between D PBL CAPE control and relax QE D+S (panels e-h) composited at 00, 06, 12 and 18 UTC. The units for cloud base mass flux are kg m^{-2} sec^{-1}.

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