

1 **Modeling the diurnal cycle over tropical oceans using the weak**
2 **temperature gradient approximation**

3 SHARON L. SESSIONS, * LEAH A. LINDSEY,
CARLOS LÓPEZ CARRILLO, AND DAVID J. RAYMOND

Physics Department and Geophysical Research Center,

New Mexico Institute of Mining and Technology, Socorro, NM 87801

* *Corresponding author address:* Sharon L. Sessions, Department of Physics and Geophysical Research Center, New Mexico Institute of Mining and Technology, 801 Leroy Pl., Socorro, NM 87801.

E-mail: sessions@kestrel.nmt.edu

ABSTRACT

4
5 We investigate the extent to which precipitation over tropical oceans is modulated by the
6 diurnal variations in the thermodynamic environment. Tropical precipitation is modeled
7 using a cloud system resolving model with the large scale parameterized using the weak
8 temperature gradient (WTG) approximation. In WTG, convection responds to specified
9 potential temperature and humidity profiles. By imposing diurnal variations observed during
10 the 2001 EPIC field program to the reference profiles of potential temperature and mixing
11 ratio, we assess the extent to which convection responds to these changes and accounts
12 for the diurnal variability in precipitation observed during EPIC. Remarkably, the WTG
13 approximation is able to reproduce a precipitation maximum near the observed time, despite
14 an imperfect reproduction of the diurnal variability in saturation fraction. The ability of the
15 model to capture the diurnal variability relies heavily on a strict enforcement of the WTG
16 approximation and the lateral entrainment of moisture into the model domain resulting from
17 this enforcement.

1. Introduction

Understanding the diurnal variability in precipitation over tropical oceans remains an important and difficult problem. Observations show that the diurnal amplitude over oceans is weak compared to that over land, and that the peak in precipitation occurs in the early morning hours with a weaker afternoon peak in some ocean regions (Yang and Slingo 2001; Nesbitt and Zipser 2003). The weak afternoon peak is associated with an increase in absorption of shortwave radiation, either by the ocean surface (Chen and Houze 1997; Sui et al. 1997) or by clear sky water vapor (Takahashi 2012). It is often obliterated in disturbed environmental conditions, and is therefore only present in limited observations where afternoon convection is associated with small, unorganized systems (Nesbitt and Zipser 2003; Cifelli et al. 2008).

The origin of the predominant early-morning precipitation maximum is not as well understood. For ocean regions in the vicinity of land, there is a strong influence from the diurnal heating of the land itself. The land-based diurnal forcing may result from extended sea breezes (Gille et al. 2003; Takahashi 2012), or from longer-ranged propagation of gravity waves initiated from land-based convection (Mapes et al. 2003a,b; Warner et al. 2003; Yang and Slingo 2001; Jiang 2012).

Ocean regions which are far from land influence also exhibit an early morning rainfall maximum. The popular mechanisms explaining this peak all involve the interaction between radiation and convection. Some mechanisms suggest that convection increases as a result of thermal destabilization of upper clouds due to enhanced radiative cooling of cloud tops (Kraus 1963; Ramage 1971; Randall et al. 1991); others emphasize the role of cloud-free regions, stating that absorption of solar radiation by water vapor warms the clear-sky regions which inhibits convective growth by reducing convergence into cloudy regions during the day (Ruprect and Gray 1976a,b; Gray and Jacobson 1977). At least one numerical study concluded that the direct interaction between radiation and convection played the primary role in modulating diurnal precipitation, with the interaction between cloudy and cloud-free

45 regions playing a secondary role (Liu and Moncrieff 1998).

46 The timing and prominence of the rainfall maximum is influenced further by interactions
47 with large-scale tropical waves (Chen and Houze 1997; Sui et al. 1997), wind patterns (Pereira
48 and Rutledge 2006; Takahashi 2012), seasonality (Hendon and Woodberry 1993; Biasutti
49 et al. 2012), location (Hendon and Woodberry 1993; Kubota and Nitta 2001; Yang and
50 Slingo 2001; Nesbitt and Zipser 2003; Cifelli et al. 2008; Biasutti et al. 2012), and whether
51 the diurnally modulated convection is part of a large-scale organized system or not (Tripoli
52 1992; Sui et al. 1997; Kubota and Nitta 2001; Nesbitt and Zipser 2003; Cifelli et al. 2008).
53 An excellent review of the proposed mechanisms involved in modulating the diurnal cycle
54 over both land and oceans is presented by Yang and Smith (2006).

55 Understanding how these mechanisms influence convection is important for improving the
56 representation of the diurnal cycle in regional and global models (Dai and Trenberth 2004;
57 Wang et al. 2007) without the computational expense associated with super-parameterized
58 (Pritchard and Somerville 2009) or global cloud resolving models (Sato et al. 2009; Noda et al.
59 2012). One approach to this problem is to consider the following question: To what extent
60 is the diurnal convection over tropical oceans modulated by changes in the thermodynamic
61 environment?

62 Raymond and Sessions (2007) showed that modeled convection in the context of the
63 weak temperature gradient (WTG) approximation is sensitive to changes in the potential
64 temperature and moisture profiles representing the convective environment. They found that
65 both moister or more stable environments resulted in more extensive convection with higher
66 average precipitation rates compared to unperturbed conditions. They also found that the
67 more stable conditions produced more “bottom-heavy” convective mass flux profiles with
68 higher precipitation efficiencies.

69 Wang et al. (2013) recently performed WTG simulations with time-dependent reference
70 profiles generated from TOGA COARE (Tropical Ocean Global Atmosphere Program’s Cou-
71 pled Ocean Atmosphere Response Experiment) observations. Their results suggested that

72 the observed precipitation variability was influenced more by forcing from surface fluxes than
73 by changes in the potential temperature profiles. It is worth noting that their study excluded
74 lateral entrainment of moisture from outside the model domain which may be important in
75 WTG simulations.

76 The weak temperature gradient (WTG) approximation provides a unique tool for as-
77 sessing the relative importance of the thermodynamic environment in the diurnal forcing of
78 convection. The WTG approximation represents a parameterization of the large scale based
79 on approximate horizontal homogeneity of virtual temperature in the tropical atmosphere.
80 In WTG simulations, convection evolves to maintain a specified reference temperature which
81 represents the convective environment. If a particular forcing mechanism diurnally modu-
82 lates the thermodynamic environment in which the convection is evolving, and if the convec-
83 tion is sensitive to those changes, then the properties of WTG-simulated convection should
84 exhibit observed characteristics of the diurnal variability in convection. Thus, we expect good
85 representation of the observed characteristics if (1) the dominant diurnal forcing mechanism
86 manifests in the thermodynamic profiles, and (2) if the convection is sufficiently sensitive to
87 the thermodynamic environment.

88 Whether or not this approach is successful will provide valuable information for improving
89 the representation of the diurnal cycle in global models. In particular, identifying the specific
90 mechanisms may be unnecessary if it is sufficient to note that they act via the thermodynamic
91 environment. This would greatly reduce the factors that need to be accounted for in large
92 scale models, given the extreme space and time heterogeneity in the observed diurnal cycles
93 over tropical oceans. On the other hand, if convection simulated via the WTG approximation
94 fails to capture the diurnal variability, we can assume that the dominant mechanisms directly
95 modulate the convection, and do not act through the thermodynamic environment.

96 To demonstrate the application of the WTG approximation in diurnal forcing, we incor-
97 porate observational data taken during the 2001 field program, EPIC2001 (East Pacific In-
98 vestigation of Climate Processes in the Coupled Ocean-Atmosphere System; Raymond et al.

99 2004), into WTG simulations. In this region, it is believed that the dominant mechanism
100 for diurnal variations is modulation by gravity waves initiated from land based convection
101 (Cifelli et al. 2008; Mapes et al. 2003b; Takahashi 2012). This location is just within the
102 range of this effect (see, e.g., Cifelli et al. 2008; Takahashi 2012); however, it doesn't pre-
103 clude the influence of other mechanisms, including the dynamic radiation-convection effect
104 (Ruprect and Gray 1976a,b; Gray and Jacobson 1977) which results from an oscillation
105 between cloudy and adjacent cloud-free regions, or the static radiation-convection mecha-
106 nism (Kraus 1963; Ramage 1971; Randall et al. 1991), in which the nighttime convection is
107 enhanced by an increase in the radiative cooling of the cloud tops which thermally desta-
108 bilizes the upper cloud. While it is clear that the gravity wave mechanism would act via
109 the thermodynamic profiles, it is likely that these alternate mechanisms would also alter the
110 potential temperature profiles and thus affect the development of convection. The goal of
111 the present study is not to determine which of these is dominant, but rather to determine
112 the extent to which changes in the thermodynamic profiles—regardless of how those changes
113 occur—influence the diurnal modulation of convection over open oceans. While these mech-
114 anisms represent an explanation for the early morning precipitation maximum, some ocean
115 regions also exhibit a weak afternoon peak which results from the heating of the ocean sur-
116 face by solar insolation. Since an afternoon peak was not observed during the EPIC program
117 (Cifelli et al. 2008; Raymond et al. 2004), this mechanism is likely to be insignificant for this
118 work. Other mechanisms summarized in Yang and Smith (2006) cannot be distinguished in
119 the work presented here, for reasons that we discuss in section 1b.

120 In the following sections, we briefly describe the observational data used for this study, as
121 well as the essential ingredients for the particular implementation of WTG used in our cloud
122 system resolving model. Following that, section 2 gives results from our simulations and
123 compares those with corresponding observations. We discuss the significance of the results
124 and conclude in section 3.

125 *a. Weak temperature gradient (WTG) approximation*

126 In this work, we use an updated version of the cloud system resolving model (CRM)
 127 described in Raymond and Zeng (2005). The model implements the weak temperature
 128 gradient approximation similar to that introduced by Sobel and Bretherton (2000). The
 129 basic idea is that buoyancy anomalies are rapidly redistributed throughout the tropical
 130 troposphere, resulting in a nearly horizontally homogeneous virtual temperature profile. In
 131 nature, this effect is achieved by gravity waves (Bretherton and Smolarkiewicz 1989; Mapes
 132 and Houze 1995). In the model, we accomplish this by generating a hypothetical vertical
 133 velocity, w_{wtg} (the weak temperature gradient vertical velocity), which counteracts the effects
 134 of diabatic heating. The WTG velocity obeys mass continuity independent of the velocity
 135 field in the model (see Raymond and Zeng 2005 or Sessions et al. 2010 for a thorough
 136 discussion of the implementation of WTG in the CRM).

137 The WTG vertical velocity enters in the governing equation for potential temperature,
 138 θ^1 :

$$\frac{\partial(\rho\theta)}{\partial t} + \nabla \cdot (\rho\mathbf{v}\theta + \mathbf{T}_\theta) \equiv \rho(S_\theta - E_\theta) \quad , \quad (1)$$

139 where ρ is the density, \mathbf{v} is the wind field computed explicitly by the model (which does
 140 not include the contribution from enforcement of WTG), \mathbf{T}_θ is the contribution due to
 141 unresolved eddy and viscous transport, S_θ is the diabatic source of potential temperature,
 142 and E_θ enforces the WTG approximation via a relaxation of θ to a reference profile θ_0 :

$$E_\theta = w_{wtg} \frac{\partial \bar{\theta}}{\partial z} = \sin(\pi z/h) \frac{\bar{\theta} - \theta_0(z)}{t_\theta} \quad . \quad (2)$$

143 Here, the overbar signifies a horizontal average over the model domain, h is the tropopause
 144 height, and t_θ is the time scale over which the domain averaged potential temperature profile
 145 relaxes to the reference profile. Practically speaking, t_θ is a measure of enforcement of WTG:

¹The weak temperature gradient approximation really applies to horizontal homogeneity of the virtual temperature. Our model doesn't distinguish between virtual and potential temperature so we enforce WTG via the potential temperature budget.

146 $t_\theta \rightarrow 0$ corresponds to strict WTG enforcement (as implemented in Sobel and Bretherton
147 2000), while $t_\theta \rightarrow \infty$ turns WTG-mode off and allows the domain to evolve to radiative
148 convective equilibrium (RCE). Physically, t_θ is believed to be associated with the time it
149 takes gravity waves to travel some characteristic distance in the model. In the work presented
150 here, we vary t_θ and examine its effect on the ability of the model to capture the diurnal
151 cycle.

152 To examine the diurnal cycle, we prescribe time-dependent perturbations to the reference
153 potential temperature profile, and we therefore modify the reference profile in equation (2) to
154 be time-dependent, $\theta_0(z, t)$. This is similar to the approach used by Wang et al. (2013), who
155 imposed the observed, time-dependent potential temperature profile from TOGA COARE
156 in the enforcement of WTG. There are several significant differences between their work and
157 the work presented here. The first is that they do not include a sinusoidal modulation of
158 the potential temperature profile that is given in equation (2). This essentially represents a
159 modulation of the gravity wave speed; the enforcement of WTG in our model is strongest
160 in the mid-troposphere and attenuates toward the tropopause and boundary layer. Both
161 here and in Wang et al. (2013), the enforcement of WTG in the boundary layer is linearly
162 interpolated to zero at the surface, since WTG is not a good approximation in the boundary
163 layer (Sobel and Bretherton 2000). Also, Wang et al. (2013) impose a relaxation time scale
164 of 4 hours. This is not fast enough to allow the convection to respond to diurnal variations
165 in the thermodynamic profiles, and thus we choose shorter relaxation times (see section 1c).

166 Probably the most significant difference between Wang et al. (2013) and the work pre-
167 sented here is the treatment of moisture. In both studies, moisture within the model domain
168 is advected vertically by the WTG vertical velocity (w_{wtg} in equation (2)); however, Wang
169 et al. (2013) do not in any way incorporate moisture outside the model domain into the
170 computational domain. There are three choices for how to incorporate environmental mois-
171 ture from outside the model domain into the domain. The first is via horizontal advection
172 by large scale circulations. The second is by specifying a separate moisture relaxation time

173 analogous to the potential temperature relaxation time given in equation (2). This was done
 174 in Sobel et al. (2007), and they found that relaxation to the reference profile has a significant
 175 impact on the ability of a model domain to sustain multiple equilibria². Alternatively, we
 176 adopt a third method which was originally implemented in Raymond and Zeng (2005). In
 177 this case, moisture is entrained laterally into the model domain by satisfying mass continuity
 178 in the WTG velocity field. The governing equation for total water mixing ratio, r_t , is given
 179 by:

$$\frac{\partial(\rho r_t)}{\partial t} + \nabla \cdot (\rho \mathbf{v} r_t + \mathbf{T}_r) \equiv \rho(S_r - E_r) \quad , \quad (3)$$

180 with \mathbf{T}_r the contribution from unresolved eddy and viscous transport, S_r the source of r_t due
 181 to precipitation and evaporation, and E_r represents both entrainment from the environment
 182 and vertical transport by large scale vertical motion:

$$E_r = \frac{(r_t - r_x)}{\bar{\rho}} \frac{\partial(\bar{\rho} w_{wtg})}{\partial z} + w_{wtg} \frac{\partial r_t}{\partial z} \quad . \quad (4)$$

183 Here,

$$r_x = \begin{cases} r_0(z, t) & \frac{\partial(\bar{\rho} w_{wtg})}{\partial z} > 0 \quad , \\ \bar{r}_t & \text{otherwise} \quad . \end{cases} \quad (5)$$

184 This definition ensures that outflowing air has a mixing ratio equal to the model domain
 185 while inflowing air has a mixing ratio equal to that of the reference profile, $r_0(z, t)$. In
 186 previous work, the reference moisture profile, r_0 , was time-independent; here we generalize
 187 the definition in anticipation of the diurnal variability of moisture observed during EPIC.

188 Another difference between the general procedure described in Wang et al. (2013) and
 189 the method here is the treatment of radiation. In order to avoid complications arising from
 190 cloud-radiation feedbacks, they prescribe a non-interactive, time-dependent radiative heating
 191 profile obtained from a simulation with imposed vertical motion (their control simulation).
 192 Our model uses interactive radiation computed from a toy radiation model (Raymond and
 193 Zeng 2000) which cools uniformly across the domain.

²Here, multiple equilibria refers to the ability for a model domain to maintain both a dry or a precipitating steady state with identical boundary conditions but different initial conditions. See also Sessions et al. (2010).

194 Finally, it is interesting to note that the reference profiles from TOGA COARE used in
195 Wang et al. (2013) represent profiles averaged over the entire Intensive Flux Array (IFA)
196 region and the results are compared against the budget-derived precipitation rate for the IFA
197 region. In the work described here, profiles are obtained from a source at a single location,
198 and we compare precipitation rates with observations from radar aboard the ship.

199 An important ingredient in the implementation of WTG is specification of the reference
200 profiles of potential temperature and mixing ratio (θ_0 and r_0 , respectively). We usually
201 take time and domain averages of a simulation run to radiative-convective equilibrium (i.e.,
202 $t_\theta \rightarrow \infty$ in equation (2)) to represent the environmental conditions outside the model domain.
203 In this work, we add observed diurnal anomalies to the RCE reference profiles to investigate
204 the response of modeled convection to diurnal variations in the temperature and moisture
205 profiles. The observed anomalies were constructed from the EPIC2001 field program, which
206 is described in the next section. Following that, we provide details of the model set-up and
207 describe the parameter space investigated in this study.

208 *b. EPIC2001 data*

209 The focus of the EPIC field program was to document and understand the mechanisms of
210 subseasonal variability in the East Pacific (see Raymond et al. 2004). The project lasted from
211 September 1 to October 10, 2001. The scope of the project included observations of a deep
212 layer of the atmosphere as well as upper layers of the ocean. The observations which con-
213 tribute to this study were all obtained from ship-based measurements from NOAA’s research
214 vessel *Ron H. Brown* (RHB). During the field program, radiosondes were launched every four
215 hours, which provide a time series of the thermodynamic environment at the ship location
216 (95°W , 10°N). Rainfall measurements were estimated from the radar aboard the RHB, and
217 are freely available from the CODIAC website (<http://data.eol.ucar.edu/codiac/>). We
218 choose the Z-R relation for precipitation calibrated from insitu data taken from NCAR’s
219 C130 measurements for comparison with observations.

220 We use the observational data in two ways: (1) to construct diurnal perturbations for
221 the reference profiles used in the WTG simulations, and (2) as a validation of model results.
222 The latter is discussed in section 2. To construct the diurnal perturbations, we started
223 with the time series taken from the JOSS/UCAR quality controlled soundings³. From these,
224 we derived a time series of potential temperature and mixing ratio profiles. Each day is
225 divided into four-hour time intervals, and each time interval is averaged over all days. In
226 this way, we construct thermodynamic profiles of a “typical day”. We note that during
227 EPIC, several easterly waves passed by and were observed from the RHB (Petersen et al.
228 2003; Raymond et al. 2004). The process of constructing a “typical day” effectively averages
229 out the thermodynamic conditions for the easterly waves, though these could contaminate
230 the time series compared to observations in undisturbed conditions. Given that some diurnal
231 cycle mechanisms operate differently in clear versus disturbed regions (Tripoli 1992; Sui et al.
232 1997; Nesbitt and Zipser 2003; Yang and Smith 2006; Cifelli et al. 2008), we have eliminated
233 our ability to isolate these effects in the WTG approximation. Nevertheless, this stands as
234 a first step toward understanding the role of the thermodynamics in the diurnal modulation
235 of convection. For the purpose of this work, we linearly interpolate the profiles to a regular
236 temporal grid with one-hour resolution.

237 Radiative convective equilibrium represents conditions in the model’s native environ-
238 ment. Thus, rather than directly imposing the thermodynamic profiles observed in EPIC,
239 we add the diurnal anomalies to the model’s RCE profile (see figures 1 and 2). In order to
240 construct statistically significant results, we repeat the simulation with diurnal anomalies for
241 25 consecutive days. For comparing the model results with observations, we keep data from
242 days 5-25, and composite all of the days for the “typical model day” with 1 hour resolution.

³Soundings recorded measurements of dew point, relative humidity, temperature, pressure, and horizontal wind velocity with 2 s vertical resolution.

243 *c. Experimental setup*

244 In order to understand the diurnal cycle in the context of the weak temperature gradi-
245 ent approximation, it is important to note that the only diurnal modulation occurs in the
246 reference potential temperature and moisture profiles. These are assumed to represent the
247 conditions immediately outside the model domain. We impose no diurnal forcing in the sur-
248 face fluxes or in radiative cooling. This is an important point given the results from Wang
249 et al. (2013) which suggest that these are both important factors in modulating precipitation
250 variability during TOGA COARE.

251 As mentioned in section 1a, we use a version of the model described in Raymond and
252 Zeng (2000), which implements the WTG approximation. All simulations are run with 2-
253 dimensional domains. The vertical dimension is 20 km, with a tropopause height of 15 km.
254 The WTG approximation is enforced in the altitude range between 1 km and 15 km. The
255 WTG vertical velocity is linearly interpolated to zero below 1 km. The vertical resolution is
256 250 m. The horizontal domain is doubly periodic and ranges in size from 100 to 400 km, with
257 one kilometer resolution. Sessions et al. (2010) found that the existence of multiple equilibria
258 in WTG simulations was sensitive to domain size, so we are investigating the extent to which
259 domain size affects characteristics of convection with diurnally modulated reference profiles.

260 For each domain size used, we ran the model for 50 days in non-WTG mode to construct
261 the RCE reference profile which serves as the baseline for diurnal anomalies. RCE was
262 calculated for a surface wind speed of 5 m s^{-1} over an ocean with a sea surface temperature
263 (SST) of 303 K. Figure 1 compares the RCE profiles of potential temperature and mixing
264 ratio for the 200 km domain with the observed mean profiles. The differences between the
265 observed and RCE profiles for all domain sizes are shown in figure 2. The model RCE states
266 are 1-2 K warmer through most of the troposphere, but cooler above 10 km compared to
267 observed conditions; they are dryer in the 2 km layer just above the boundary layer, and
268 moister aloft. Also note that the 400 km domain has the largest differences from the observed
269 potential temperature profile (differences between 100 and 200 km domains are negligible in

270 figure 2a), while the 100 km domain is the driest in the mid-troposphere compared to the
271 other RCE states and the observations. For this reason, we choose to perform most of our
272 sensitivity experiments on a 200 km domain.

273 For each set of RCE reference profiles, we impose the diurnal anomalies derived from
274 the EPIC field program. These are shown in figure 3 (local time, LT). Note that in the
275 early morning hours (0000-0400 LT), the lowest 5 km are moist and cool relative to the daily
276 mean. Both of these would be expected to produce heavier precipitation, according to results
277 from Raymond and Sessions (2007). As the day progresses, the lower troposphere dries and
278 becomes more unstable, which is expected to decrease precipitation efficiency. Based on the
279 observed diurnal anomalies and the results from Raymond and Sessions (2007), we would
280 expect the precipitation maximum to occur between 0-4 LT, with an afternoon minimum.
281 Note the significant anomalies in potential temperature near the tropopause throughout
282 most of the day. While the enforcement of the WTG approximation will certainly respond
283 to those anomalies, the gravity waves enforcing WTG attenuate at altitudes approaching the
284 tropopause as a result of the sinusoidal modulation in equation (2). While this helps damp
285 the influence of these anomalies, we note that care should be taken in interpreting model
286 results at high altitudes.

287 In order to procure a large enough sample for statistical averaging, we impose the diurnal
288 anomalies shown in figure 3, linearly interpolated to every hour, for 25 consecutive simulation
289 days. The model diurnal cycle is constructed from the average of each hour for the last 20
290 days of the 25 day simulations. In order to assess the variability in the model results, we
291 run a few simulations for 45 days and compare averages from two different 20 day segments.

292 In addition to varying the domain size, we also considered the effect of additional constant
293 surface fluxes by increasing the surface wind speed relative to RCE conditions. A diurnal
294 cycle was not imposed in surface wind speed, SSTs, or in the radiation scheme (diurnal
295 variations are imposed only in the reference profiles of potential temperature and mixing
296 ratio). Though the increase in SST from solar insolation likely contributes to the minor

297 afternoon peak (Chen and Houze 1997; Sui et al. 1997; Yang and Smith 2006), the afternoon
298 peak is not observed in the EPIC region (Cifelli et al. 2008; Raymond et al. 2004). This
299 mechanism also tends to be more prevalent in undisturbed or clear regions (Nesbitt and
300 Zipser 2003; Cifelli et al. 2008), and the passing easterly waves during EPIC would have made
301 it difficult to capture this effect. Furthermore, Cifelli et al. (2008) showed that the diurnal
302 variability in latent heat flux, SST, and surface wind speed was small during EPIC. Thus,
303 we justify the neglect of diurnal forcing in surface fluxes both because we expect this to be a
304 small contribution to the diurnal cycle in precipitation, and because our primary goal is to
305 determine the extent to which convection is forced by diurnal changes in the thermodynamic
306 environment. Our results will be particularly interesting in light of the Wang et al. (2013)
307 conclusions that the intraseasonal variability in TOGA COARE is largely a result of surface
308 forcing.

309 Finally, we also investigate how the degree to which the WTG approximation is strictly
310 enforced affects the model’s ability to generate a diurnal cycle. To do this, we vary the
311 potential temperature relaxation time, t_θ in equation (2). We do not expect to detect
312 diurnal variations for large t_θ since the convection will respond on a time scale longer than
313 the time scale of changes in the perturbations. As t_θ becomes smaller than the time scale
314 of diurnal variability, the modeled convection responds much faster to those changes and we
315 expect to generate a diurnal cycle with which we can compare to observations.

316 2. Results

317 Our primary goal is to compare the observed diurnal variability with simulations having
318 diurnal forcing imposed in the thermodynamic environment and enforced via the WTG
319 approximation. The most significant comparison is in the diurnal cycle of precipitation. For
320 this, we use the median radar-derived rain rate over a domain which extends 100 km in all
321 directions from the *Ron H. Brown*. The Z-R relation used is from the Baumgardner C-130

322 insitu data (Cifelli et al. 2002). We also use this data to compare the fraction of the model
 323 domain that is precipitating to the observed rain fraction.

324 In addition to precipitation rate and rain fraction, we compare several other variables
 325 which can easily be calculated from the sonde data used for the diurnal forcing in the WTG
 326 simulations. These include a measure of the atmospheric instability, saturation fraction,
 327 deep convective inhibition, the vertical distribution of moisture, and mean boundary layer
 328 mixing ratio.

329 Atmospheric instability is diagnosed from the saturated moist entropy. We define an
 330 instability index according to

$$\Delta s^* = s_{low}^* - s_{mid}^* \quad , \quad (6)$$

331 where s_{low}^* is the saturated moist entropy averaged over the 1-3 km layer and s_{mid}^* is the sat-
 332 urated moist entropy averaged over the 5-7 km layer. If the environment is saturated, larger
 333 Δs^* corresponds to greater instability which promotes higher precipitation rates, according
 334 to Raymond and Sessions (2007).

335 The saturation fraction is defined as the ratio of precipitable water to saturated pre-
 336 cipitable water. As in Raymond et al. (2011), we approximate the moist entropy by $s \approx$
 337 $s_d + Lr_v/T_R$, where s_d is the dry entropy, L is the (constant) latent heat of condensation, r_v
 338 is the water vapor mixing ratio, and T_R is a constant reference temperature. Using this, we
 339 can approximate the saturation fraction by

$$S \approx \frac{\int_0^h \rho(s - s_d) dz}{\int_0^h \rho(s^* - s_d) dz} \quad , \quad (7)$$

340 where the integrals are taken from the surface to the tropopause height, h .

341 We also compare the deep convective inhibition (DCIN; Raymond et al. 2003), which is
 342 defined as

$$DCIN = s_t^* - s_b \quad , \quad (8)$$

343 where s_t^* is the vertical average of saturated moist entropy over the height range 2000-2500
 344 m; it is the threshold entropy for convection. The boundary layer entropy, s_b , is defined as

345 the vertical average of moist entropy over the height range 0-1750 m.

346 Cifelli et al. (2008) showed that diurnal variability in the mean boundary layer mixing
347 ratio also exhibited a significant diurnal amplitude. Given that our model does not ade-
348 quately resolve the boundary layer, and that WTG does not apply in this layer, we would
349 not expect good agreement with observations. Nevertheless, we calculate the mean boundary
350 layer mixing ratio in the lowest kilometer and compare with observations.

351 Nesbitt and Zipser (2003) and Biasutti et al. (2012) analyzed satellite data and concluded
352 that the diurnal cycle in this region is a result of more frequent convective events rather than
353 more intense events. This is consistent with the Cifelli et al. (2008) observation that there
354 is a diurnal cycle in the fraction of the region that is precipitating (rain fraction). To see if
355 our model qualitatively captures these observations, we compare the fraction of the model
356 domain which is precipitating to the reported fractional area of precipitation in Cifelli et al.
357 (2008). For this purpose, a grid point is considered precipitating if it has a precipitation rate
358 of at least 1 mm hour^{-1} .

359 We begin the data analysis with a comparison between observations and the results from
360 selected WTG simulations.

361 *a. Comparison with EPIC observations*

362 In comparing the WTG simulations with observations, we would expect the best results
363 with a strict enforcement of WTG. Figure 4 compares the observed values of rain rate, insta-
364 bility index, saturation fraction, DCIN, mean boundary layer mixing ratio, and rain fraction
365 to select WTG simulations. The simulations shown correspond to strict enforcement of
366 WTG ($t_\theta = 6.7 \text{ s}$ in equation (2), which is just larger than the 5 second time step implemented
367 in the model) on 100, 200, and 400 km domains, and a slightly relaxed enforcement ($t_\theta = 67$
368 s) of the WTG approximation on a 200 km domain.

369 Not all observed features in the diurnal variability are reproduced in the WTG simula-
370 tions; however, the model does an excellent job in capturing the early morning precipitation

371 peak with a mid-afternoon/early-evening minimum. The 200 km domain with strict en-
372 forcement of WTG (black short-dashed line in figure 4a) shows an earlier peak at 0400 LT
373 compared to the EPIC observations or the other two simulations shown. However, the exact
374 timing and magnitude of the early morning peak is quite variable, even within a single model
375 run. Figure 5 shows a comparison between two different 20 day segments in the 45 day run
376 for the 200 and 400 km domains with strict enforcement of WTG. The saturation fraction
377 and instability index exhibit no change in the timing of the diurnal variations, while the
378 timing of the precipitation maximum varies up to 3 hours. Similar variability is exhibited
379 with a 100 km domain (not shown). While this figure provides a sense of the magnitude of
380 the noise in these simulations, it nevertheless maintains a clear diurnal cycle which agrees
381 well with observations.

382 All simulations in figures 4 and 5 show considerably reduced precipitation rates in the
383 afternoon compared to observations. One may hypothesize that this is a result of excluding
384 the diurnal variability in surface fluxes which result from SST and wind speed variability.
385 We do not think this is the case here because the diurnal variability in these quantities so
386 small (0.5 K and 0.7 m s^{-1} , resp. Cifelli et al. 2008) that they are insufficient to increase
387 the precipitation rate by 5 mm day^{-1} in our model (compare precipitation rates for wind
388 speeds of 5 and 10 m s^{-1} in figure 8). Instead, we suspect that the dramatic reduction
389 in precipitation rate in the late afternoon compared to the peak value in early morning is
390 more likely a result of the two-dimensionality in the model domains. Wang and Sobel (2011)
391 compared WTG simulations between two- and three-dimensional (2D and 3D, respectively)
392 CRM domains. They found that 2D domains had lower values of gross moist stability (GMS)
393 which resulted in larger precipitation rates compared to corresponding 3D runs. Raymond
394 and Sessions (2007) demonstrated that lower GMS is associated with increased stability, so
395 we interpret these results as an enhancement of the precipitation response to instability via
396 GMS in 2D compared to 3D (Wang and Sobel 2011). This effect also seems to apply to
397 smaller domain sizes (see figure 4 and section 2c). Thus, a more stable atmosphere would

398 produce more precipitation while a more unstable atmosphere would correspond to smaller
399 precipitation rates with the effect exaggerated in 2D.

400 The WTG simulations in general do an excellent job of capturing the diurnal variability
401 in atmospheric instability and DCIN (figures 4b,d). This is perhaps not surprising since the
402 simulations shown represent strict enforcements of the WTG approximation, which means
403 that we expect potential temperature anomalies in the model to replicate observed diurnal
404 anomalies (this is the forcing imposed after all). Figure 6 shows excellent agreement between
405 the observed diurnal anomalies in potential temperature from the EPIC soundings and from
406 the strict enforcement (i.e., $t_\theta = 6.7$ s) of WTG on the 400 km domain. Since DCIN
407 is calculated from the entropy profiles (which are related to potential temperature), we
408 expect these to follow the observed diurnal tendencies, and figure 4d shows this is indeed
409 the case. Note that there is an offset between the observed and simulated instability index
410 and DCIN. This is likely due to the differences in the mean thermodynamic profiles in the
411 model environment compared to the real environment.

412 The most significant difference between the model and observed values occurs with the
413 saturation fraction, as shown in figure 4c. In this case, the model captures the general trend,
414 with the highest simulated values near the highest observed values, but it underestimates
415 the saturation fraction in the early morning hours, and does not capture the late afternoon
416 increase at all. We can understand these differences by comparing the vertical distributions
417 of moisture in the model with those from the EPIC observations. The left panel of figure 7
418 shows the diurnal mixing ratio anomalies from the RHB soundings. These were added to the
419 reference profile to represent the environmental moisture surrounding the model domain (r_0
420 in equation (5)). The right panel of figure (7) shows the diurnal variations in mixing ratio
421 calculated by the model. While the model captures the timing in the diurnal variability, all
422 of the variability is in the lowest few kilometers of the model domain; it completely misses
423 the variations in the free troposphere. The lack of a positive moisture anomaly in the 1-5 km
424 layer in the early morning explains the model's underestimation of the saturation fraction at

425 this time. Similarly, the dry anomaly in the lowest model layer extends later in the afternoon
426 than in observations, which explains in part why the afternoon peak is not seen in the model.
427 A thorough analysis of the how the model is distributing moisture in the troposphere will
428 be investigated in future work.

429 Despite the limitations of the model to accurately reproduce the free tropospheric mois-
430 ture, the diurnal cycle in boundary layer moisture seems to qualitatively agree with observa-
431 tions. We can see that in the lowest layers in the mixing ratio shown in figure 7, and in the
432 mean boundary layer mixing ratio shown in figure 4e. The latter also approximately agrees
433 with the results in figure 5 of Cifelli et al. (2008).

434 Analysis of three years (1997-2000) of TRMM satellite data by Nesbitt and Zipser (2003)
435 found that the peak in diurnal rainfall variability was almost exclusively a result of an increase
436 in the number of systems, not in the intensity of the systems. A very high resolution analysis
437 of the TRMM data between 1998 and 2007 by Biasutti et al. (2012) also attributed the peak
438 in diurnal variability to an increase in frequency of rainfall, not intensity. As a quick check
439 to see if the WTG simulations capture this tendency, we can look at the diurnal variability
440 in rain fraction in the model domain. We define the rain fraction to be the fraction of the
441 domain having a rainfall rate greater than 1 mm hr^{-1} . Figure 4f compares the rain fraction in
442 the WTG simulations to the rain fraction observed during EPIC. The rain fraction increases
443 proportionally to the rainfall, which indicates there is a larger fraction of the domain that
444 is precipitating, rather than the same fraction with a higher intensity. This is qualitatively
445 consistent with observations by Nesbitt and Zipser (2003) and Biasutti et al. (2012) and also
446 agrees with the diurnal variability in rain area of mesoscale convective systems reported by
447 Cifelli et al. (2008, see their figure 12), and simulated on a global CRM (Noda et al. 2012).
448 Furthermore, it is notable that the rain fraction data derived from radar is independent of
449 the sounding data used in the WTG simulations. Thus, it provides additional validation for
450 investigating the diurnal variability in the context of the WTG approximation.

451 The results in this section are actually quite remarkable, and they suggest that enforcing

452 the WTG approximation on diurnal timescales reproduces observed variability to a much
453 better degree than might be expected. Though it is not surprising that the model repro-
454 duces the potential temperature variability and by extension the instability and DCIN, is it
455 surprising that it gets the approximate timing in precipitation maximum correct, and it does
456 a decent job on mean boundary layer mixing ratio and rain fraction. The main deficiency
457 is that the model fails to capture the variability in the vertical profiles of moisture, and
458 consequently some features in the saturation fraction. Despite this, the model still does a
459 good job in representing the diurnal variability in the precipitation rate.

460 *b. Sensitivity to WTG relaxation time*

461 Here, we examine the sensitivity of the modeled diurnal cycle on the WTG relaxation
462 time, t_θ , in equation (2). These experiments are performed using a 200 km domain, with
463 surface wind speeds equal to the RCE wind speed ($v_y = 5 \text{ m s}^{-1}$) to see the effect of diurnal
464 variations in reference profiles only. We repeat these experiments with stronger surface wind
465 speeds ($v_y = 10 \text{ m s}^{-1}$) to examine the extent to which surface fluxes enhance or diminish the
466 diurnal variability. Figure 8 shows the modeled diurnal cycle in precipitation rate, saturation
467 fraction and instability index for the different relaxation time scales for surface wind speeds
468 of 5 m s^{-1} (left panels) and 10 m s^{-1} (right panels). Observed values are shown in blue.

469 As seen in figure 8, the diurnal amplitude diminishes rapidly with even a slight increase
470 in the relaxation time scale. It is virtually absent in all observables for $t_\theta \geq 1$ hour, though
471 prominent features are all retained for $t_\theta \sim 10$ minutes, regardless of the the imposed surfaces
472 fluxes (which are modulated by a constant surface wind speed in this case). The reason that
473 the diurnal variability vanishes for longer relaxation times is because the reference profile
474 is changing faster than the model has time to adjust to those changes. This suggests that
475 convection must respond rapidly to diurnal variations in the thermodynamic environment
476 for this to be a viable mechanism in the diurnal cycle.

477 Increasing the imposed wind speed, and hence surface fluxes, enhances the diurnal cycle

478 in precipitation for strict enforcement of WTG ($t_\theta \leq 11$ minutes). The additional moisture
479 (comparing middle panels in figure 8) in the early morning hours contributes to the larger
480 precipitation maximum in the early morning as well as a slight increase in the afternoon
481 precipitation rate compared to lower surface wind speeds. The dramatic increase in precipi-
482 tation rate in the early morning hours compared to the slight increase in the late afternoon
483 for a proportional increase in saturation fraction is likely a result of the sensitive dependence
484 of precipitation rate on saturation fraction (Bretherton et al. 2004; Raymond et al. 2007) as
485 well as the increase in precipitation efficiency due to a more stable environment (Raymond
486 and Sessions 2007).

487 Examining the instability index in these experiments is a simple way to diagnose the
488 enforcement of WTG. It explains why the diurnal variability based on forcing via the ther-
489 modynamic profiles vanishes with a weaker enforcement of WTG. Once the diurnal cycle in
490 the instability index vanishes, the lateral entrainment of environmental moisture becomes
491 uniform and the diurnal signal vanishes in both saturation fraction and precipitation rate.

492 *c. Effect of domain size*

493 Figure 4 shows the effect of domain sizes varying from 100 km - 400 km on the model's
494 ability to reproduce the observed diurnal variations. With strict enforcement of WTG, the
495 instability index closely resembles the observed values for all domain sizes. There are slightly
496 lower values for the 400 km domain compared to the 100 and 200 km domains, which is a
497 result of the slightly warmer free troposphere in RCE for the 400 km domain compared to the
498 other two (see figure 2). Also, we can see that the smaller the domain, the higher the mean
499 saturation fraction, which is also a result of the moister free troposphere for successively
500 smaller domains in the unperturbed RCE profiles (figure 2). It is interesting to see how
501 these variations affect the domain mean precipitation rates for the different domain sizes.
502 Probably the most significant difference is the magnitude of precipitation rate in the 100
503 km domain compared to the 200 and 400 km domains. The peak precipitation rate for the

504 100 km domain is 60 mm day^{-1} (peak not shown) near 0300 LT, whereas the peak rates for
505 the 200 and 400 km domains are much closer to the observed 16 mm day^{-1} . This is likely a
506 result of a combination of the moister reference environment and the exaggerated increase
507 in precipitation efficiency for more stable environments (see the discussion in section 2a).

508 **3. Discussion and conclusions**

509 The goal of this work is to determine to what extent the diurnal variability in convection
510 over open oceans is modulated by changes in the thermodynamic environment. We per-
511 formed a series of numerical experiments which incorporated diurnal anomalies observed in
512 the vertical profiles of potential temperature and mixing ratio taken from radiosonde data
513 during the EPIC2001 field program. The limited domain simulations implemented the weak
514 temperature gradient approximation, which parameterizes the large scale environment by
515 enforcing the potential temperature profile in the model to relax to the reference profile rep-
516 resenting the environment outside the model domain. This enforcement generates a vertical
517 velocity (the weak temperature gradient vertical velocity), which vertically advects moisture
518 and, via mass continuity, results in lateral entrainment of moisture from the environment
519 outside the domain.

520 There are several proposed mechanisms which explain the diurnal variability in precipi-
521 tation over open oceans, and in particular the early morning rainfall peak. The EPIC region
522 is just within the boundaries where gravity waves from land-based convection can modulate
523 the convection, and this is believed to be an important mechanism in this location. Other
524 potential mechanisms may be classified as interactions between radiation and convection,
525 as explained in section 1. The work presented here does not aim to determine which of
526 the possible mechanisms are responsible, only whether or not the convection responds suffi-
527 ciently fast to changes in the thermodynamic environment so that the principal features in
528 diurnal variability are reproduced in WTG simulations. Thus, we expect good results if (1)

529 the mechanisms governing the diurnal variability manifest in the thermodynamic environ-
530 ment, and (2) if the convection is sufficiently sensitive to the thermodynamic environment.
531 While not all of the proposed mechanisms would be expected to manifest in the thermody-
532 namic environment (see Yang and Smith 2006), it is likely that the greatest contributions are
533 from those that do (propagating gravity waves and radiation-convection interactions would
534 certainly modify the local temperature profiles).

535 In order to assess the success of this approach, we compared the modeled diurnal variabil-
536 ity in several observable quantities with measurements from the EPIC field program. While
537 most of the comparisons were able to reproduce the general trends in the daily cycle, this
538 might be expected by the design of the project. In particular, we imposed diurnal variations
539 from the thermodynamic profiles taken from radiosonde measurements, and a significant
540 number of our comparisons were against variables also measured in the soundings. Thus,
541 the most significant comparison is between the model results and a source that is indepen-
542 dent of the sounding data. For this purpose we use the radar-derived precipitation rate
543 and rain fraction. With a strong enough enforcement of WTG, our model reproduces the
544 observed early morning precipitation maximum and the corresponding peak in rain fraction.
545 The diurnal variability in rain fraction indicates that a larger fraction of the model domain
546 is precipitating rather than the same fraction with a higher intensity, consistent with obser-
547 vations. In this case, the modeled diurnal variations in precipitation are much stronger than
548 observed, though they vanish as the enforcement of the WTG approximation weakens (i.e.,
549 as the potential temperature relaxation time scale approaches the time scale of the imposed
550 changes, about 1 hour).

551 There are at least two significant results from this work. The first is, based on the ability
552 of the model to reproduce significant features in the diurnal variability of convection over
553 open oceans, we conclude that the diurnal variability is largely modulated by changes in the
554 thermodynamic environment. The second is that WTG is a valid approach to understanding
555 mechanisms controlling tropical convection. Wang et al. (2013) demonstrated one way to

556 incorporate observations into WTG simulations to investigate dominant mechanisms in the
557 evolution of the Madden Julian Observation. This work represents another example of
558 incorporating observations to investigate a phenomenon on a completely different time scale
559 and under different environmental conditions. The general idea of integrating observations
560 in WTG simulations is a promising opportunity to make significant gains not only in our
561 understanding of the convective response to changes in the environment, but to help identify
562 mechanisms which dominate the convective evolution in a variety of different atmospheric
563 conditions.

564 *Acknowledgments.*

565 We would like to thank Shuguang Wang and Adam Sobel for useful discussions, and
566 Paquita Zuidema and Robert Cifelli for assistance with EPIC data, which is provided by
567 NCAR/EOL under sponsorship of the National Science Foundation (<http://data.eol.ucar.edu/>).
568 Many of the simulations were performed on The New Mexico Computing Applications Cen-
569 ter supercomputer, ENCANTO, and on the supercomputer EXEMPLAR, located on the
570 New Mexico Tech campus. This work was supported by U.S. National Science Foundation
571 Grants AGS-1056254 and AGS-1021049.

REFERENCES

- 574 Biasutti, M., S. E. Yuter, C. D. Burleyson, and A. H. Sobel, 2012: Very high resolution
575 rainfall patterns measured by TRMM precipitation radar: seasonal and diurnal cycles.
576 *Climate Dyn.*, **39**, 239–258, doi:10.1007/s00382-011-1146-6.
- 577 Bretherton, C. S. and P. K. Smolarkiewicz, 1989: Gravity waves, compensating subsidence
578 and detrainment around cumulus clouds. *J. Atmos. Sci.*, **46**, 740–759.
- 579 Bretherton, C. S., et al., 2004: The EPIC 2001 stratocumulus study. *Bull. Am. Meteor. Soc.*,
580 **85**, 967–977.
- 581 Chen, S. S. and R. A. Houze, 1997: Diurnal variation and life-cycle of deep convective
582 systems over the tropical Pacific warm pool. *Q. J. Roy. Meteor. Soc.*, **123**, 357–388.
- 583 Cifelli, R., D. Baumgardner, W. A. Petersen, S. A. Rutledge, C. Williams, P. Johnston, and
584 K. Gage, 2002: Comparison Z-R relationships in EPIC-2001. *Eos Trans. AGU Fall Meet.*
585 *Suppl.*, San Francisco, CA, AGU, Vol. 83(47), Abstract A22A–0053.
- 586 Cifelli, R., S. W. Nesbitt, S. A. Rutledge, W. A. Petersen, and S. Yuter, 2008: Diurnal
587 characteristics of precipitation features over the tropical east Pacific: A comparison of the
588 EPIC and TEPPS regions. *J. Climate*, **21**, 4068–4085, doi:10.1175/2007JCLI2020.1.
- 589 Dai, A. and K. E. Trenberth, 2004: The diurnal cycle and its depiction in the Community
590 Climate System Model. *J. Climate*, **17**, 930–951.
- 591 Gille, S. T., S. G. L. Smith, and S. M. Lee, 2003: Measuring the sea breeze from Quikscat
592 scatterometry. *Geophys. Res. Lett.*, **30**, 1114, doi:10.1029/2002GL016230.
- 593 Gray, W. M. and R. W. Jacobson, 1977: Diurnal variation of deep cumulus convection. *Mon.*
594 *Wea. Rev.*, **105**, 1171–1188.

- 595 Hendon, H. H. and K. Woodberry, 1993: The diurnal cycle of tropical convection. *J. Geophys.*
596 *Res.*, **98**, 16 623–16 637.
- 597 Jiang, Q., 2012: On offshore propagating diurnal waves. *J. Atmos. Sci.*, **69**, 1562–1581,
598 doi:10.1175/JAS-D-11-0220.1.
- 599 Kraus, E. B., 1963: The diurnal precipitation change over the sea. *J. Atmos. Sci.*, **20**,
600 546–551.
- 601 Kubota, H. and T. Nitta, 2001: Diurnal variations of tropical convection observed during
602 the TOGA-COARE. *J. Meteor. Soc. Japan*, **79**, 815–830.
- 603 Liu, C. and M. W. Moncrieff, 1998: A numerical study of the diurnal cycle of tropical oceanic
604 convection. *J. Atmos. Sci.*, **55**, 2329–2344.
- 605 Mapes, B. E. and R. A. Houze, 1995: Diabatic divergence profiles in western Pacific mesoscale
606 convective systems. *J. Atmos. Sci.*, **52**, 1807–1828.
- 607 Mapes, B. E., T. T. Warner, and M. Xu, 2003a: Diurnal patterns of rainfall in northwestern
608 South America. Part I: Observations and context. *Mon. Wea. Rev.*, **131**, 799–812.
- 609 Mapes, B. E., T. T. Warner, and M. Xu, 2003b: Diurnal patterns of rainfall in northwestern
610 South America. Part III: Diurnal gravity waves and nocturnal convection offshore. *Mon.*
611 *Wea. Rev.*, **131**, 830–844.
- 612 Nesbitt, S. W. and E. J. Zipser, 2003: The diurnal cycle of rainfall and convective intensity
613 according to three years of TRMM measurements. *J. Climate*, **16**, 1456–1475.
- 614 Noda, A. T., K. Oouchi, M. Satoh, and H. Tomita, 2012: Quantitative assessment of diurnal
615 variation of tropical convection simulated by a global nonhydrostatic model without
616 cumulus parameterization. *J. Climate*, **25**, 5119–5134, doi:10.1175/JCLI-D-11-00295.1.

- 617 Pereira, L. G. and S. A. Rutledge, 2006: Diurnal cycle of shallow and deep convection for
618 a tropical land and an ocean environment and its relationship to synoptic wind regimes.
619 *Mon. Wea. Rev.*, **134**, 2688–2701.
- 620 Petersen, W. A., R. Cifelli, D. J. Boccippio, S. A. Rutledge, and C. Fairall, 2003: Convection
621 and easterly wave structures observed in the eastern Pacific warm pool during EPIC-2001.
622 *J. Atmos. Sci.*, **60**, 1754–1773.
- 623 Pritchard, M. S. and R. C. J. Somerville, 2009: Assessing the diurnal cycle of precipitation
624 in a multi-scale climate model. *J. Adv. Model. Earth Syst.*, **1**, 12, doi:10.3894/JAMES.
625 2009.1.12.
- 626 Ramage, C. S., 1971: *Monsoon Meteorology*. Academic Press, 295 pp.
- 627 Randall, D. A., Harshvardhan, and D. A. Dazlich, 1991: Diurnal variability of the hydrologic
628 cycle in a general circulation model. *J. Atmos. Sci.*, **48**, 40–62.
- 629 Raymond, D. J., G. B. Raga, C. S. Bretherton, J. Molinari, C. López-Carrillo, and Ž. Fuchs,
630 2003: Convective forcing in the intertropical convergence zone of the eastern Pacific. *J.*
631 *Atmos. Sci.*, **60**, 2064–2082.
- 632 Raymond, D. J. and S. L. Sessions, 2007: Evolution of convection during tropical cyclogen-
633 esis. *Geophys. Res. Lett.*, **34**, L06 811, doi:10.1029/2006GL028607.
- 634 Raymond, D. J., S. L. Sessions, and C. L. Carrillo, 2011: Thermodynamics of tropi-
635 cal cyclogenesis in the northwest Pacific. *J. Geophys. Res.*, **116**, D18 101, doi:10.1029:
636 2011JD015624.
- 637 Raymond, D. J., S. L. Sessions, and Z. Fuchs, 2007: A theory for the spinup of tropical
638 depressions. *Q. J. Roy. Meteor. Soc.*, **133**, 1743–1754.
- 639 Raymond, D. J. and X. Zeng, 2000: Instability and large scale circulations in a two-column
640 model of the tropical troposphere. *Quart. J. Roy. Meteor. Soc.*, **126**, 3117–3135.

- 641 Raymond, D. J. and X. Zeng, 2005: Modelling tropical atmospheric convection in the context
642 of the weak temperature gradient approximation. *Quart. J. Roy. Meteor. Soc.*, **131**, 1301–
643 1320.
- 644 Raymond, D. J., et al., 2004: EPIC2001 and the coupled ocean-atmosphere system of the
645 tropical east Pacific. *Bull. Amer. Meteor. Soc.*, **85**, 1341–1354.
- 646 Ruprect, E. and W. M. Gray, 1976a: Analysis of satellite-observed cloud clusters. Part I:
647 Wind and dynamic fields. *Tellus*, **28**, 391–413.
- 648 Ruprect, E. and W. M. Gray, 1976b: Analysis of satellite-observed cloud clusters. Part II:
649 Thermal, moisture and precipitation. *Tellus*, **28**, 414–426.
- 650 Sato, T., H. Miura, M. Satoh, Y. N. Takayabu, and Y. Wang, 2009: Diurnal cycle of precipita-
651 tion in the tropics simulated in a global cloud-resolving model. *J. Climate*, **22**, 4809–4826,
652 doi:10.1175/2009JCLI2890.1.
- 653 Sessions, S. L., S. Sugaya, D. J. Raymond, and A. H. Sobel, 2010: Multiple equilibria in a
654 cloud resolving model using the weak temperature gradient approximation. *J. Geophys.*
655 *Res.*, **115**, D12 110, doi:10.1029/2009JD013376.
- 656 Sobel, A. H., G. Bellon, and J. Bacmeister, 2007: Multiple equilibria in a single-column model
657 of the tropical atmosphere. *Geophys. Res. Lett.*, **34**, L22 804, doi:10.1029/2007GL031320.
- 658 Sobel, A. H. and C. S. Bretherton, 2000: Modeling tropical precipitation in a single column.
659 *J. Climate*, **13**, 4378–4392.
- 660 Sui, C.-H., K.-M. Lau, Y. N. Takayabu, and D. A. Short, 1997: Diurnal variations in tropical
661 oceanic cumulus convection during TOGA COARE. *J. Atmos. Sci.*, **54**, 639–655.
- 662 Takahashi, K., 2012: Thermotidal and land-heating forcing of the diurnal cycle of oceanic
663 surface winds in the eastern tropical Pacific. *Geophys. Research Lett.*, **39**, L04 805, doi:
664 10.1029/2011GL050692.

- 665 Tripoli, G. J., 1992: An explicit three-dimensional nonhydrostatic simulation of a tropical
666 cyclone. *Meteor. Atmos. Phys.*, **49**, 229–254.
- 667 Wang, S. and A. H. Sobel, 2011: Response of convection to relative sea surface temperature:
668 cloud-resolving simulations in two and three dimensions. *J. Geophys. Res.*, **116**, D11 119,
669 doi:10.1029/2010JD015347.
- 670 Wang, S., A. H. Sobel, and Z. Kuang, 2013: Cloud-resolving simulation of TOGA-COARE
671 using parameterized large scale dynamics. *J. Geophys. Res.*, submitted.
- 672 Wang, Y., L. Zhou, and K. Hamilton, 2007: Effect of convective entrainment/detrainment of
673 the simulation of the tropical precipitation diurnal cycle. *Mon. Wea. Rev.*, **135**, 567–585,
674 doi:10.1175/MWR3308.1.
- 675 Warner, T. T., B. E. Mapes, and M. Xu, 2003: Diurnal patterns of rainfall in northwestern
676 South America, Part II: Model simulations. *Mon. Wea. Rev.*, **131**, 813–829.
- 677 Yang, G.-Y. and J. Slingo, 2001: The diurnal cycle in the tropics. *Mon. Wea. Rev.*, **129**,
678 784–801.
- 679 Yang, S. and E. A. Smith, 2006: Mechanisms for diurnal variability of global tropical rainfall
680 observed from TRMM. *J. Climate*, **19**, 5190–5226.

681 List of Figures

- 682 1 Time-mean potential temperature (left) and mixing ratio (right) profiles. The
683 blue lines are the observed profiles from EPIC, dashed black lines are RCE
684 profiles on a 200 km domain. The RCE profiles are the unperturbed reference
685 profiles for the WTG simulations. 31
- 686 2 Deviation of the RCE reference profiles from mean observations for domain
687 sizes of 100, 200, and 400 km. For all domain sizes, the RCE profiles are 1-2
688 K warmer through most of the troposphere compared to observed conditions.
689 The RCE profiles were also moister aloft and drier in the 2 km layer just above
690 the boundary layer. 32
- 691 3 Observed mean diurnal anomalies in potential temperature (red) and mixing
692 ratio (blue). Hours shown are in local time (LT). 33
- 693 4 Comparison between simulated diurnal cycle WTG simulations and observa-
694 tions: (A) precipitation rate, (B) instability index, (C) saturation fraction,
695 (D) DCIN, (E) mean boundary layer mixing ratio, and (F) the fraction of the
696 domain having a precipitation rate of at least 1 mm hour^{-1} . The solid blue
697 line denotes observations from EPIC, the black dashed lines are from 200 km
698 domains (short dashes for $t_\theta = 6.7 \text{ s}$; long dashes for $t_\theta = 67 \text{ s}$), the gray line
699 is for the 400 km domain with $t_\theta = 6.7 \text{ s}$. 34
- 700 5 Comparison between simulated diurnal cycle for composites of two different
701 20 day segments in single simulations with strict enforcement of WTG. Black
702 short dashed lines and green long dashed lines are from 200 km and 400 km
703 domains, respectively. 35

- 704 6 The left panel shows the diurnal anomalies in potential temperature from
705 the EPIC soundings. These were imposed in the reference profiles for the
706 WTG simulations. The right panel shows the simulated potential temperature
707 anomalies for a strict enforcement of WTG (i.e., $t_\theta = 6.7$ s) on a 400 km
708 domain. 36
- 709 7 The left panel shows the diurnal anomalies in mixing ratio from the EPIC
710 soundings. These are the anomalies applied to $r_0(z, t)$ in equation 5 for the
711 WTG simulations. The right panel shows the simulated mixing ratio anoma-
712 lies for a strict enforcement of WTG (i.e., $t_\theta = 6.7$ s) on a 400 km domain. 37
- 713 8 Simulated diurnal cycle in precipitation rate (top), saturation fraction (mid-
714 dle), and instability index (bottom) for relaxation time scales ranging from
715 6.7 seconds to 1.85 hours. The left panels correspond to imposed surface wind
716 speed of 5 m s^{-1} ; the right panels have imposed surface wind speed of 10 m
717 s^{-1} . Observed values from the RHB are shown in blue for comparison. 38

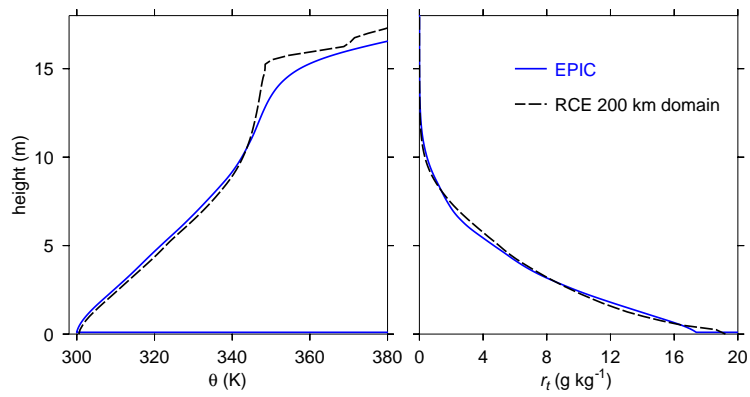


FIG. 1. Time-mean potential temperature (left) and mixing ratio (right) profiles. The blue lines are the observed profiles from EPIC, dashed black lines are RCE profiles on a 200 km domain. The RCE profiles are the unperturbed reference profiles for the WTG simulations.

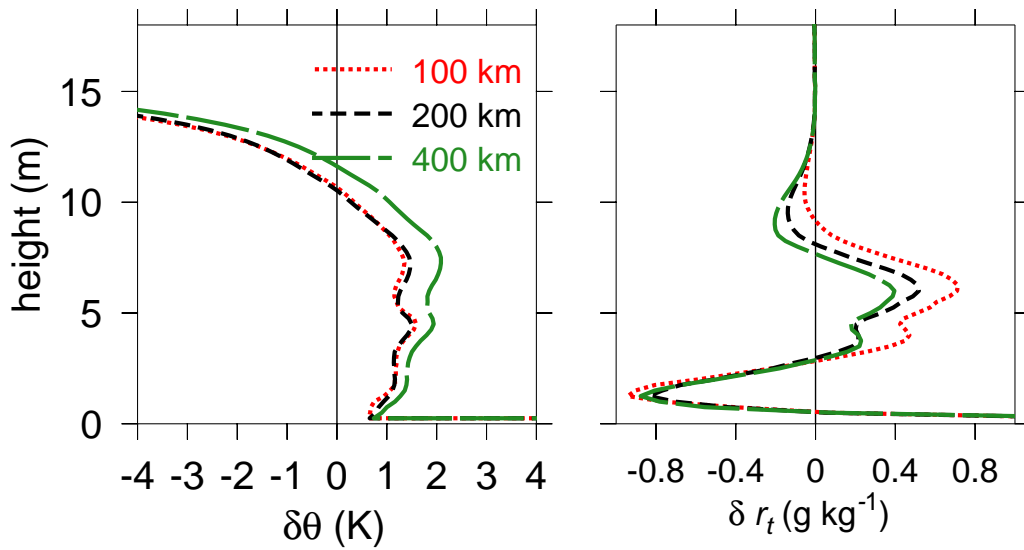


FIG. 2. Deviation of the RCE reference profiles from mean observations for domain sizes of 100, 200, and 400 km. For all domain sizes, the RCE profiles are 1-2 K warmer through most of the troposphere compared to observed conditions. The RCE profiles were also moister aloft and drier in the 2 km layer just above the boundary layer.

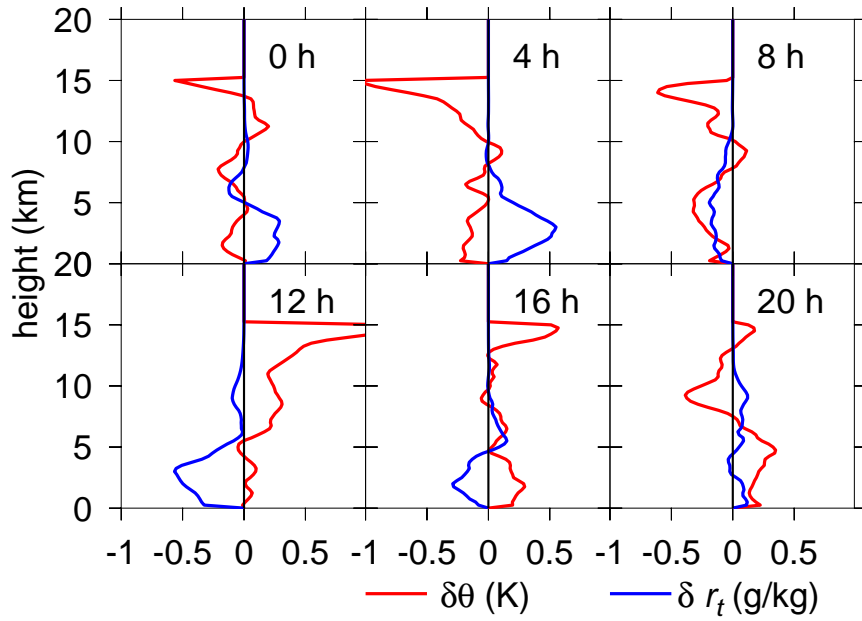


FIG. 3. Observed mean diurnal anomalies in potential temperature (red) and mixing ratio (blue). Hours shown are in local time (LT).

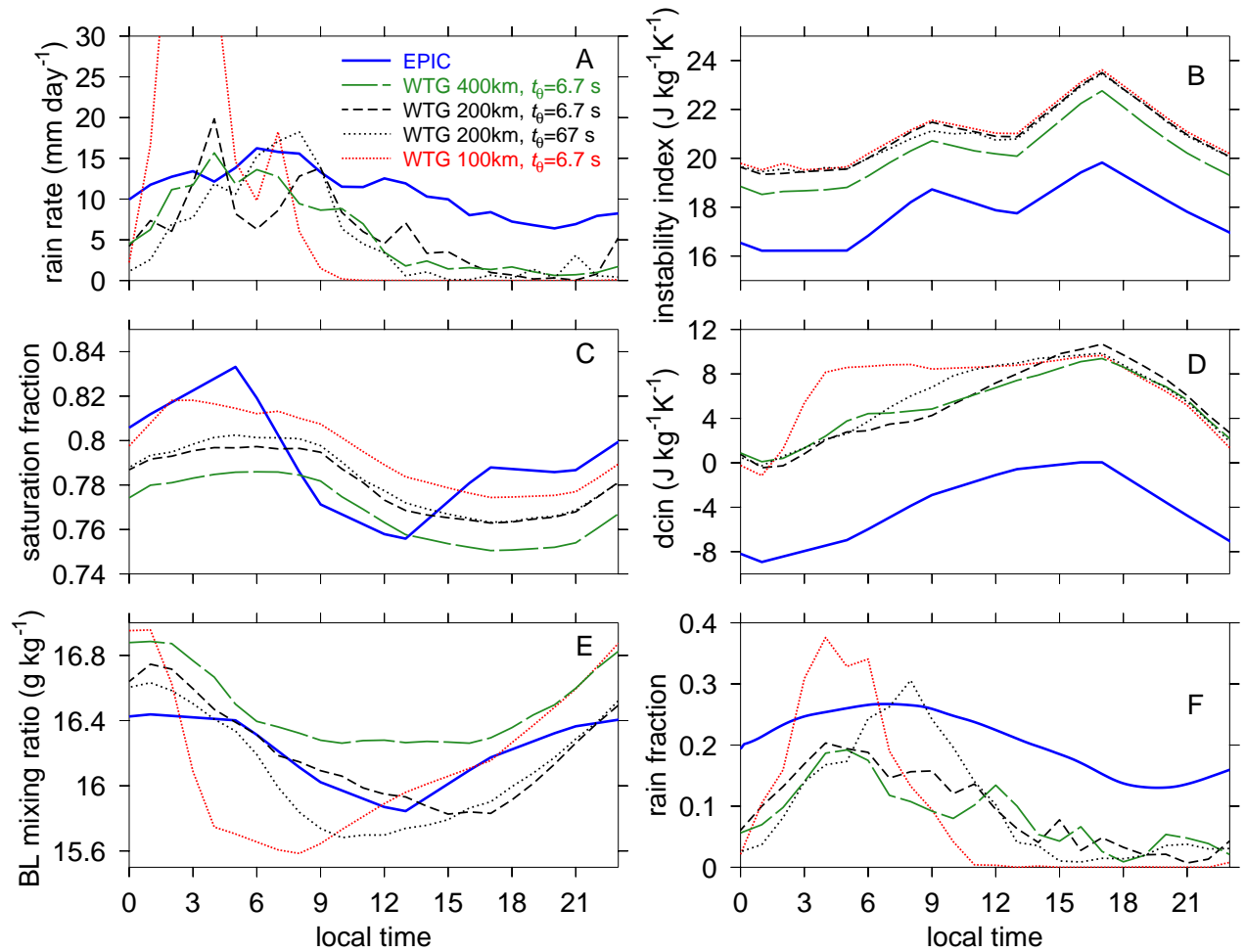


FIG. 4. Comparison between simulated diurnal cycle WTG simulations and observations: (A) precipitation rate, (B) instability index, (C) saturation fraction, (D) DCIN, (E) mean boundary layer mixing ratio, and (F) the fraction of the domain having a precipitation rate of at least 1 mm hour^{-1} . The solid blue line denotes observations from EPIC, the black dashed lines are from 200 km domains (short dashes for $t_\theta = 6.7 \text{ s}$; long dashes for $t_\theta = 67 \text{ s}$), the gray line is for the 400 km domain with $t_\theta = 6.7 \text{ s}$.

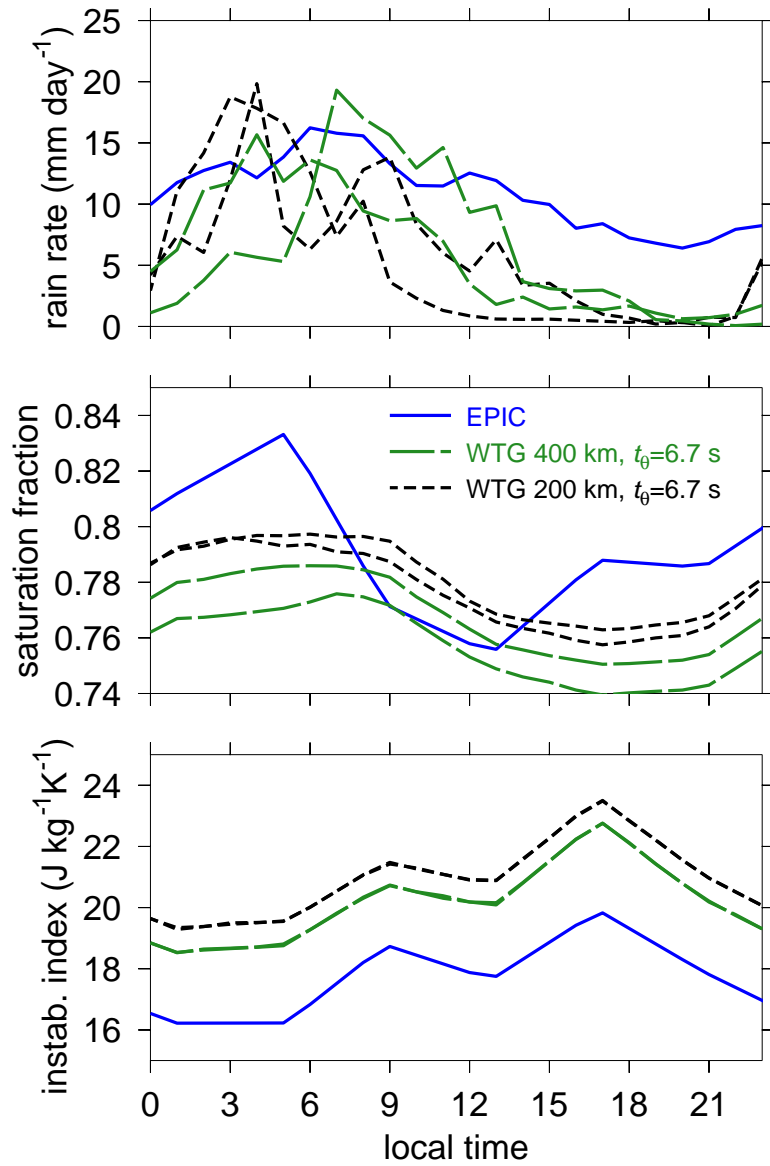


FIG. 5. Comparison between simulated diurnal cycle for composites of two different 20 day segments in single simulations with strict enforcement of WTG. Black short dashed lines and green long dashed lines are from 200 km and 400 km domains, respectively.

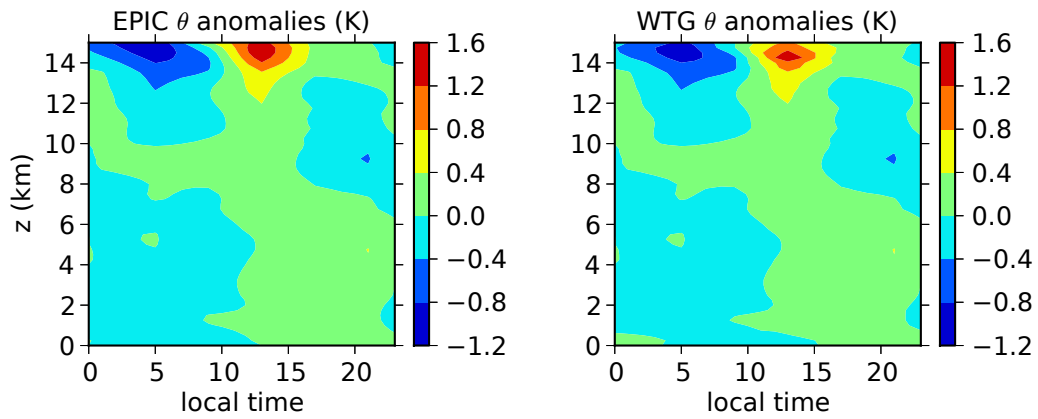


FIG. 6. The left panel shows the diurnal anomalies in potential temperature from the EPIC soundings. These were imposed in the reference profiles for the WTG simulations. The right panel shows the simulated potential temperature anomalies for a strict enforcement of WTG (i.e., $t_\theta = 6.7$ s) on a 400 km domain.

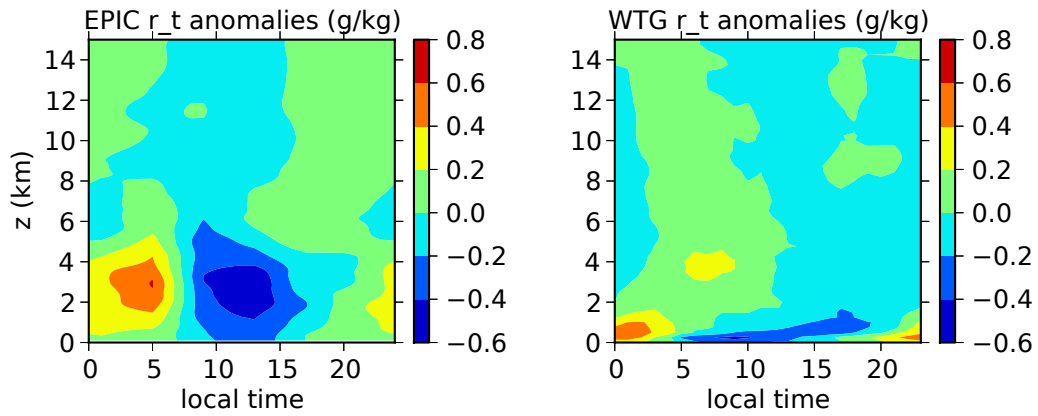


FIG. 7. The left panel shows the diurnal anomalies in mixing ratio from the EPIC soundings. These are the anomalies applied to $r_0(z, t)$ in equation 5 for the WTG simulations. The right panel shows the simulated mixing ratio anomalies for a strict enforcement of WTG (i.e., $t_\theta = 6.7$ s) on a 400 km domain.

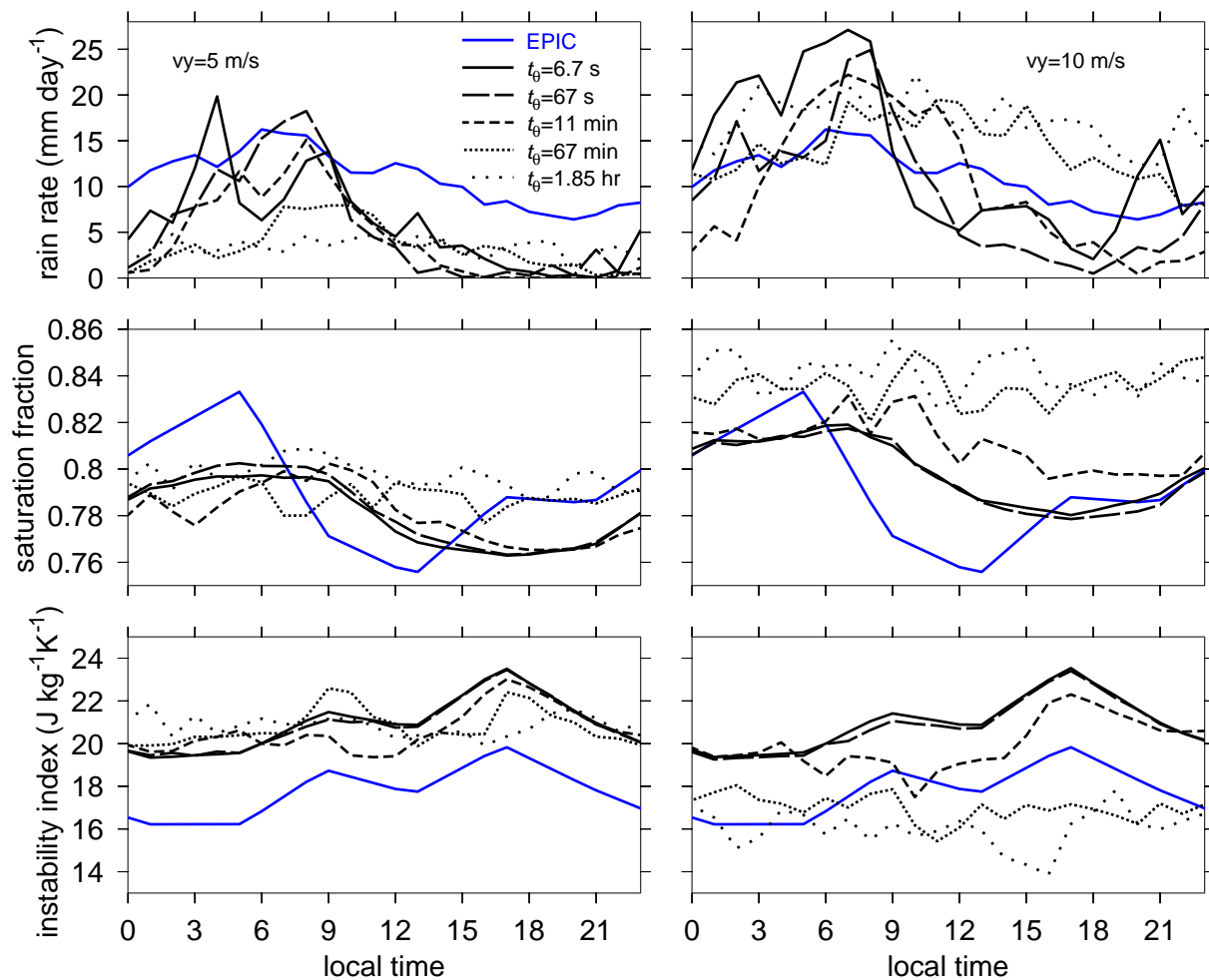


FIG. 8. Simulated diurnal cycle in precipitation rate (top), saturation fraction (middle), and instability index (bottom) for relaxation time scales ranging from 6.7 seconds to 1.85 hours. The left panels correspond to imposed surface wind speed of 5 m s^{-1} ; the right panels have imposed surface wind speed of 10 m s^{-1} . Observed values from the RHB are shown in blue for comparison.