Understanding Biases in Simulating the Diurnal Cycle of Convection over the Western Coast of Sumatra: Comparison with Pre-YMC Observation Campaign

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ABSTRACT

Biases in simulating the diurnal cycle of convection near the western coast of the island of Sumatra have been investigated using the data from the pilot field campaign of the Years of the Maritime Continent (pre-YMC). The campaign was carried out at a sea [Research Vessel (R/V) *Mirai*] and a land (Bengkulu, Sumatra) site. Simulations are performed using a tropical configuration of the Met Office model at a grid resolution of 1.5 km in a limited-area mode. The focus of this study is to understand how biases in the input conditions from ECMWF high-resolution deterministic forecast affect the diurnal cycle. Modeled precipitation is found to be delayed and weak, with cold SST bias in the model as the key contributing factor affecting convection at both sites. Colder SST causes a delay in the trigger of convection at Bengkulu by delaying the onset of the local land breeze, which in turn delays the local convergence. The cold outflow from precipitation over the adjacent mountain is also found to be delayed in the model, contributing to the total delay. This delay in the evening convection at Bengkulu is shown to directly affect the timing of nighttime convection at *Mirai*. Weaker convection at Bengkulu is argued to be due to lower-tropospheric dry humidity bias in the model initial condition. Convection at *Mirai* is shown to be caused by the convergence of the cold outflow from Bengkulu with the prevailing landward wind over the sea. Both thermodynamic and dynamic conditions near the cold outflow front are found to be less favorable for intense convection in the simulation, the reason for which is argued to be a combination of the cold SST bias and a weaker cold outflow.

1. Introduction

Rainfall over the tropical Maritime Continent (MC) is dominated by convective activities with a distinct diurnal cycle that varies from place to place (Yang and Slingo 2001). While precipitation over land tends to reach its maximum late in the afternoon, precipitation over open sea reaches its maximum in the early morning. In the coastal regions, precipitation shows a complex behavior that is affected by both land and sea. As most

of the coastal areas in this region have significant topography, mountains also form an integral part of the land and sea interaction. The rainfall peaks over coastal regions tend to show a maximum over the mountains in the early afternoon, followed by another maximum at the mountain foot early in the evening that then migrates offshore in the night. It has been reported and discussed extensively in the literature using both models and observations. See for example, Houze et al. (1981), Yang and Slingo (2001), Neale and Slingo (2003), Mori et al. (2004), Sakurai et al. (2005), Qian (2008), Wu et al. (2009), Love et al. (2011), Hassim et al. (2016), Yokoi et al. (2017), and the references therein.

Although the focus of these studies has been on understanding the diurnal cycle of convection in the coastal MC, some of them have also commented on the

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biases in their simulations. For example, Neale and Slingo (2003) reported systematic dry rainfall bias over Maritime Continent in the global circulation model (GCM) simulation at a resolution of $2.5^\circ \times 3.75^\circ$ in latitude and longitude, respectively. They attributed these biases to the model's inability to capture the diurnal land–sea interaction in this region because of its coarse grid resolution. They further reported that even three-fold increase in the (horizontal) grid resolution did not make any difference. Qian (2008) performed 25-km regional model simulation and came to the same conclusion that the underestimation of precipitation over the MC in GCMs are the result of inadequate representation, or even absence, of islands and mountains because of their coarse resolution. More recently, Hassim et al. (2016) performed convection-permitting (4 km) simulation and found a reasonable agreement of the accumulated rainfall with Tropical Rainfall Measuring Mission (TRMM) regarding spatial patterns. The authors observed biases near the mountainous regions where the model was found to overestimate rainfall intensity. They also attributed the overestimation of rainfall to the model's inability to adequately represent entrainment at convection-permitting resolution.

The grid resolution is known to be critical in controlling the model biases, regardless of whether it is an underestimation (Neale and Slingo 2003; Qian 2008) or an overestimation (Hassim et al. 2016). However, biases in the initial and boundary conditions are important as well, more so for the regional simulations because of their dependency on the lateral boundary conditions (LBC). For example, Birch et al. (2016) performed regional climate simulations at grid resolutions of 12 and 4.5 km with parameterized and explicit convection, respectively, over the western MC and found that the variation of domain mean wind and equivalent potential temperature with Madden–Julian oscillation (MJO) phase are very similar in the two simulations. Based on this result, they concluded that the input from LBC have stronger control over the regional simulations than the way convection is represented in the model. Zhu et al. (2017) conducted coupled simulations by nudging model sea surface temperature (SST) to model climatologies at different time scales and found that the weakening in the air–sea coupling strength significantly degrades the MJO propagation. Their results clearly signify the role of SST in correctly simulating convection over the MC.

It is reasonable to expect the climate simulations to be more sensitive to the boundary conditions than the initial conditions. For the numerical weather prediction (NWP), on the other hand, the initial condition is as important as the boundary condition, which makes it more challenging to diagnose the reasons for a simulation bias. To our

knowledge, no study has looked at the role of these input conditions (initial and boundary conditions) for modulating the diurnal convection over the western MC from a NWP perspective. We believe that there is a strong need for such studies, particularly for the meteorological services in the MC region running regional models for weather prediction using input conditions from the big centers like the European Centre for Medium-Range Weather Forecasts (ECMWF), the Met Office (United Kingdom), and the National Oceanic and Atmospheric Administration (United States).

A novelty of the current study is that it utilizes results from a convection-permitting state-of-the-art NWP model to highlight the biases in the input conditions from the high-resolution (9 km) deterministic forecast from ECMWF. These biases were discovered by comparing the results from a tropical version of the Met Office Unified Model (UM) with the observed data collected during the pre-YMC campaign (Yokoi et al. 2017). Pre-YMC was the pilot study program of the two years (2017–19) international project Years of Maritime Continent (YMC).¹ It is stressed that the present article does not aim to highlight issues in the model physics at convection-permitting resolution nor does it intend to provide the final answers for the simulation biases. It only seeks to highlight possible biases in the input conditions for the interest of meteorological services in the region and to show how they affect the modeled precipitation.

In the following, observation and simulation details are provided in section 2. Section 3 is for the results and discussion in which we shall try to explain the precipitation biases in the simulation. We end with the conclusions in section 4.

2. Observation and simulation details

The pre-YMC intensive observation campaign was conducted near the western coast of the Indonesian island of Sumatra (Fig. 1b), from 22 November to 24 December of 2015, by the Japan Agency for Marine–Earth Science and Technology, the Indonesian Agency for the Assessment and Application of Technology, and the Indonesian Agency for Meteorology, Climatology and Geophysics. Two observation sites, one located on the coastal water [Research Vessel (R/V) *Mirai*] and another nearly 55 km away on the coastal city of Bengkulu, Sumatra, were selected to study rainfall migration. These two

¹ A draft science plan of this project can be accessed online at the following website: http://www.jamstec.go.jp/ymc/docs/YMC_SciencePlan_v2.pdf.

FIG. 1. (a) Contours of orography showing the simulation domain; (b) Zoomed view of the marked rectangular area in (a) showing the observation sites (R/V *Mirai* and Bengkulu) along the transect AB.

sites lie on a line roughly perpendicular to the coastline that runs approximately linearly from northwest to southeast. During the period, radiosondes were released every 3 h at both the sites. Basic surface meteorological measurements were also conducted. Yokoi et al. (2017) noted that the diurnal cycle in the period 23 November–12 December was more regular than in the later period of 14–24 December. They attributed this change in the diurnal behavior to the change in the large-scale wind conditions from easterly to westerly in the lower troposphere sometime between 12 and 13 December [see Fig. 8 of Yokoi et al. (2017)]. This change in the wind direction caused the wind to blow toward the land obscuring the offshore migration of precipitation (Ichikawa and Yasunari 2008). Based on their observation, and also to use the longest available period, we have restricted our study period to 23 November–12 December of 2015. Both observation and simulation data are composited over this period to create a mean diurnal cycle (referred to as the diurnal cycle henceforth). More details of the campaign can be found in Yokoi et al. (2017).

Simulations are performed using a tropical version (SINGV) of the Met Office model, UM, run in a limited-area mode at a grid resolution of 1.5 km. SINGV is called a tropical version because its science configuration is different from the one used in the United Kingdom. The main differences are in the cloud scheme and the treatment of the boundary layer parameterization. While the U.K. version uses the diagnostic cloud scheme of Smith (1990) and a value of 0.2 for the turbulence coefficient in the Smagorinsky turbulence scheme, the prognostic cloud scheme of Wilson et al. (2008) is used in

SINGV and the turbulence coefficient set to 0.5 in the Smagorinsky scheme. Furthermore, the U.K. version uses stochastic perturbations for temperature and specific humidity in the boundary layer, which in SINGV has been deactivated. Simulations are initialized using the ECMWF analysis every day at 0000 UTC. The ECMWF forecasts are used as lateral boundary condition every 3 h. Soil moisture and SST are also initialized using ECMWF analysis. While the soil moisture is prognosed by the land surface scheme, SST is kept constant for the entire forecast. Each forecast length is 36 h of which the first 12 h are discarded due to spinup. Figure 1 shows the simulation domain and the two observation sites. Figure 1b also indicates nine transects parallel to the middle transect, which passes through *Mirai* and Bengkulu. These transects, which are about 2 km apart from each other, are used to average modeled data and GPM rainfall spatially as discussed in section 2b.

The reason for choosing such a domain in which the observation sites are somewhat close to the southern boundary is that it is the standard domain used at the Centre for Climate Research Singapore (CCRS) for NWP applications. Comparison with the high-resolution pre-YMC campaign data provides us with a unique opportunity to validate the model and understand its biases. However, to disregard lateral boundaries as a potential cause of the seen biases, complete simulation on a domain with an extended southern boundary (by 1°) was also performed. The differences in the diurnal cycle of standard surface variables in these two domains were found to be small (not shown) suggesting that the domain boundary probably has no effect on the biases discussed in the article.

FIG. 2. Time–distance plot of hourly mean precipitation rate, averaged across the transects in Fig. 1b, from (a) GPM and (b) UM for the entire simulation period (22 Nov–12 Dec 2015). The distance (in km) is measured positive from SW toward NE. The solid and dashed vertical lines indicate the locations of *Mirai* and Bengkulu, respectively, and the horizontal dashed lines indicate 1900 LT on each day.

a. Overview of the simulation period

The spatial pattern of modeled precipitation is assessed with the Global Precipitation Measurement (GPM) (Huffman 2017) rainfall product. Figure 2 shows the time–distance plot of hourly rainfall averaged over all transects in Fig. 1b. Distance is increasing from southwest (SW) to northeast (NE) and vertical solid and dashed lines mark the locations of *Mirai* and Bengkulu, respectively. The simulation data are remapped onto the GPM grid for comparison. The diurnal cycle of rainfall is more evident in GPM than in the model. Out of 14 days, on which the observed rainfall is found to migrate offshore from Bengkulu, the model has captured 11 of them. Offshore propagation speed is also seen to be reasonably captured by the model. Heavy rainfall ($>20 \text{ mm h}^{-1}$) at Bengkulu appears to originate over the mountain ranges, which then propagates over to the sea in both observation and model. This suggests that most of the heavy rainfall events at *Mirai* trigger due to propagating convective systems that developed over the mountains, rather than

from the systems initiating over the sea. Similar finding was reported in Hassim et al. (2016).

The diurnal cycle of hourly rainfall in Fig. 3 shows a delay at Bengkulu, which is not apparent in Fig. 2. While precipitation at Bengkulu is seen to peak at 1900 LT in GPM, it is found to peak a few hours later in the model. The figure also shows a line connecting the centroids of high precipitation (yellow patch) between Bengkulu and *Mirai* to identify the direction and speed of offshore propagation. These are marked GG' in Fig. 3a and MM' in Fig. 3b. It is clear from the figure that although the modeled precipitation at Bengkulu is delayed by a few hours, once generated, the convective system propagates offshore nearly at the same speed (3.5 m s^{-1}) as the observation (3.8 m s^{-1}).

In Fig. 3, the model overestimates rainfall at Bengkulu, which is opposite to what is found in section 3 when the same is compared to the rain gauge measurement in Fig. 4a. The diurnal cycle of hourly precipitation from GPM at the point closest to the gauges are also shown in Figs. 4a,b for reference. It is worth noting in Figs. 4a,b

FIG. 3. As in Fig. 2, but for the mean diurnal cycle of precipitation rate. The black oblique lines marked GG' and MM' indicate the direction and speed of propagation of precipitation offshore. Mountain peak adjacent to Bengkulu is marked by the bold vertical line.

that precipitation from GPM and gauge differ significantly for short lived (<2 h) events, over both Bengkulu and *Mirai*. Phases of the longer-lived precipitating events, such as the diurnal convection, are nearly the same in both GPM and gauge but the magnitude differ. While the gauge reports lack spatial representativeness, which is where satellite products are very useful, they are considered to be the most reliable source of point measurement (Huffman et al. 1997). Therefore, only in situ gauge measurements from the pre-YMC campaign are used in this article to assess the high-resolution modeling biases.

b. Processing of data

Care needs to be taken when comparing the model data with the observation, especially for such a short

simulation period. As the surface data in the observation were recorded every minute, an hourly running mean is applied to the dataset to filter spurious signals, (except for the data used in Fig. 12). (The data used in Fig. 12 is filtered by applying a 10-min running mean, the reason for which is explained in the corresponding section.) No filtering is applied to the radiosonde and SST data recorded every 3 h.

Surface simulation data are averaged over and stored every 10 min whereas the vertical profiles are averaged over and stored every hour. As the model data are compared with the point measurements, it is further averaged in a circular domain of radius 10 km about the observation sites to account for slight displacement of a rainfall event. Furthermore, a diameter of 20 km amounts to 13 grid points in one direction, which is

FIG. 4. Mean diurnal cycle of observed (rain gauge and GPM) and modeled precipitation rate at (a) Bengkulu, (b) *Mirai*, and (c) sea surface temperature at *Mirai*. Point closest to the rain gauges are used to calculate diurnal cycle from GPM data.

roughly the effective wavelength resolved in a typical NWP model (Skamarock 2004). When plotting figures along the transects shown in Fig. 1b, spatial averaging over the transects is performed. Note that these transects extend to about 18 km along the coastlines, which is similar to the diameter of the circle used for averaging model data about the observation sites. This is to make sure that the model data are consistently used in the figures.

3. Results and discussion

a. Diurnal cycle

To look deeper into the simulation biases, time series of the diurnal cycle of precipitation at both sites and the diurnal cycle of SST at *Mirai* are shown in Fig. 4. We start with SST in Fig. 4c, which clearly shows that the model SST has a cold bias of around 0.9 K. Furthermore, both model and observation show strong convective precipitation at Bengkulu (Fig. 4a), which starts at 1600 local time (LT) and peaks around 1900 LT in the observation. The modeled peak is delayed roughly by 3 h and is weaker in magnitude. Similarly, precipitation peak at *Mirai* (Fig. 4b) is observed around 2230 LT, which appears to be delayed and weaker in UM. Unlike in the observation, modeled precipitation shows two peaks at *Mirai*. One at 0100 LT and the other at 0400 LT. Following the line MM' in Fig. 3b, we deduce that it is the peak around 0100 LT which corresponds to the diurnally migrating convective system from land to the sea. As the delay at *Mirai* is roughly the same as at Bengkulu and also because the propagation speed (see Fig. 3) in model and GPM are nearly the same, we further conclude that precipitation at *Mirai* is delayed primarily because of the delay in the trigger of convection at Bengkulu.

In summary, the major biases noted in the diurnal cycle of simulated precipitation and the questions that we shall try to address in the remaining article are as following:

- Why is the peak in model convection delayed at Bengkulu?
- What causes weaker convection at Bengkulu in the model?
- What causes weaker convection at *Mirai* in the model?

The delay in the diurnal peak at *Mirai* is not listed as a major bias because it is directly related to the delay in the peak at Bengkulu, as discussed in the previous paragraph.

b. Local wind conditions

As our arguments in the section below draw a lot from the findings of Mori et al. (2004), it is important to note the differences in the local wind conditions in the

present case and that of Mori et al. (2004). While the local wind at the sea site in Mori et al. (2004) were reported to be westerly in the lower 6 km and easterly above that, during the pre-YMC period it is found to be easterly above 1 km and variable to westerly (late in the afternoon) below 1 km. It can be seen in Fig. 5, which shows the time–height contours of zonal wind and horizontal velocity vectors at *Mirai*, for both UM and observation. The figure also shows that the mean wind conditions and its diurnal variation are reasonably captured by the model, although it has an overestimation in the easterly around noon between 6 and 9 km.

c. Why is the peak in model convection delayed at Bengkulu?

To answer the question, we first need to understand why convection peaks at 1900 LT at Bengkulu, which is situated at the foot of a mountain. Two main ingredients for convection to take place are positive buoyancy above the condensation level and low-level convergence to overcome any available negative buoyancy. Figure 6 shows contours of buoyancy ($B = T_{\text{vparcel}} - T_{\text{venvironment}}$) of an undiluted parcel lifted from the surface layer in UM and observation. Both model and observation have negative buoyancy throughout the day in the lower 1 km, except between 1100 and 1500 LT in the observation, and layers of positive buoyancy above that. If the trigger of convection was to be dictated by parcel buoyancy, then the diurnal peak should have taken place between 1100 and 1500 LT when the negative buoyancy is the lowest in both model and observation. As this is not the case, we infer that the trigger of convection at Bengkulu is dictated by convergence of the buoyancy driven circulations instead.

Convergence over coastal Sumatra has been discussed by Mori et al. (2004) and Wu et al. (2009). They both suggest that the main trigger for convection over the coastal mountains is provided by the sea breeze when it meets the mountain ranges. A precipitation peak over the mountains is generally observed in the afternoon which then slowly dissipates. Mori et al. (2004) argued that the secondary precipitation peak at the mountain foot is a result of local convergence between background westerly (in their case) and the land breeze. Wu et al. (2009) does not discuss the rainfall at the mountain foot but does highlight the role of cold surface outflow (referred to as cold outflow henceforth), resulting from the dissipating convection over the mountains, in triggering convection just offshore.

To investigate land–sea interaction in details, we look at the horizontal wind anomaly projected along the transects in Fig. 1b. Figure 7 shows the vertical profiles of horizontal wind anomaly, together with their daily

FIG. 5. Time–height contour of the mean diurnal cycle of zonal wind and horizontal wind vectors at *Mirai* in (left) simulation and (right) observation.

means, for both sites. Note that positive wind speed implies landward direction in the figure. Here, wind anomalies are calculated with respect to its daily mean at the respective sites. Profiles are shown around the time convection is observed (i.e., at 1600, 1900, and 2200 LT). It is noted that the observed along-transect wind at *Mirai* (blue curves in Fig. 7b) is landward (sea breeze) in the lower 300 m until 2200 LT, at which time it reverses its direction toward the sea. As the mean along-transect wind (see the inset in Fig. 7b) is also landward in the lower 300 m, we conclude that the total landward wind at *Mirai* is a combination of sea breeze and the background wind before 2200 LT. The observed along-transect wind at Bengkulu (blue curves in Fig. 7a)

is also landward in the lower 300 m at 1600 LT, but it changes sign somewhere between 1600 and 1900 LT to become seaward (land breeze) and stays like that afterward. Although the mean wind at Bengkulu (see the inset in Fig. 7a) is opposing the land breeze, its magnitude is relatively small in the lower 300 m to affect the direction of the total wind. Therefore, it is the convergence of landward wind from *Mirai* with the local land breeze at Bengkulu around 1900 LT that triggers convection at Bengkulu. UM (red curves in Fig. 7) shows similar land–sea interaction but instead of starting around 1900 LT, as in the observation, the land breeze at Bengkulu in UM starts around 2200 LT causing a delay in the peak of convection.

FIG. 6. Undiluted buoyancy (in K) of a parcel lifted from the surface layer ($z = 18$ m) in (left) simulation and (right) observation at Bengkulu.

FIG. 7. Vertical profiles of anomaly of horizontal wind projected along the transects in Fig. 1b at indicated local times for (a) Bengkulu and (b) *Mirai*. Anomaly is calculated with respect to the daily mean shown in the inset.

The next obvious question is—Why is the local land breeze delayed in the model? To answer that, we first try to understand how the land breeze is triggered at Bengkulu. Figure 8 shows the cross-sectional view of temperature anomaly contours and velocity vectors, overlaid on the topography, along the transects in Fig. 1b. Temperature anomaly is calculated with respect to the daily mean at each point (both horizontal and vertical) in the plane of the transect. The solid and dashed vertical lines in the figure indicate the locations of *Mirai* and Bengkulu, respectively. Precipitation (in mm h^{-1}) along the transect is also shown for assistance. The first panel shows that precipitation peaks over the mountains at 1630 LT² with no sign of a cold outflow

² Time stamp in the model output represents the center of the averaging window. That is, 1630 represents averaging between 1600 and 1700 LT. However, whenever model and observation are compared, model outputs are linearly interpolated to the observation nominal time.

FIG. 8. Contours of mean diurnal cycle of modeled temperature anomaly and velocity vectors, overlaid on the topography, along the transects in Fig. 1b is shown at the indicated local times. Anomaly is calculated with respect to daily mean at each point in the transect plane. The solid and the dashed line indicate the locations of *Mirai* and Bengkulu, respectively. Rain rate (rr) along the transect is shown in the bottom panel of each plot. The distance (in km) is measured positive from SW toward NE.

and the wind in the lower 300 m mostly landward. By 1930 LT, the convective system has moved seaward with the corresponding precipitation peak lying just behind Bengkulu. Downdrafts from this moving convective system have given rise to a shallow cold patch of surface flow over the western slope of the mountain and at Bengkulu. By 2230 LT, the cold outflow has moved little ahead of Bengkulu together with the precipitation peak. The local wind direction has also reversed to become seaward by this time and is seen to be moving along with the cold outflow. This suggests that the cold outflow is key for the generation of a local land breeze at Bengkulu because it reduces the temperature locally, which makes the conditions more conducive to develop a land breeze.

The preceding discussion implies that the land breeze at Bengkulu is delayed either because of a weaker land–sea temperature contrast or because of a delay in arrival

FIG. 9. (a) Mean diurnal cycle of precipitation rate, and (b) as in Fig. 7, but for the increased SST ($SST_{\text{control}} + 2 \text{ K}$) case at Bengkulu.

of the cold outflow at Bengkulu. The former can happen due to the cold SST bias in the model (see Fig. 4c) and the latter due to a delay in precipitation over the adjacent mountain. Focusing on precipitation over the mountain (see the bold solid line) in Fig. 3a, it is noted that GPM has two peaks: the first starts around 1400 LT in the afternoon and the second around 1700 LT in the evening. While the afternoon precipitation in UM appears to start early by an hour, the evening precipitation, which produces the cold outflow at Bengkulu, is found to start late by an hour. As the total delay at Bengkulu is about 3 h, we infer that the primary reason of a delay in the onset of the local land breeze at Bengkulu in UM is the weaker land–sea temperature contrast due to its lower SST.

To support our argument, another simulation is performed for the same period by artificially increasing the model SST by 2 K in the domain. The resulting diurnal cycle of precipitation and the vertical profile of along-transect wind at Bengkulu are shown in Figs. 9a and 9b, respectively. The along-transect wind in the lower 300 m, which is landward at 1900 LT in the control run, is found to be seaward in the increased SST case. This early onset of the local land breeze due to increased SST has led to an early convergence, which in turn has resulted in improvement in the phase of the peak of convection as seen in Fig. 9a. The early onset of the land breeze is accompanied by a timely onset of afternoon and evening precipitation over the mountain, and an improvement in the propagation speed (3.8 m s^{-1}) of the offshore migrating rainfall system, as shown in Fig. 10.

It is to be noted that the increased SST case is an artificial construct as the temperature increment has been applied to the sea surface in the entire simulation

domain, whereas all we know is that it is only around *Mirai* that the prescribed SST is lower than the observed. On top of that, the increment is 1.1 K higher than the real SST bias, which we did intentionally to amplify the effects. Therefore, the results from this simulation must be carefully interpreted. For this reason, we have only selected a few results from this simulation to support arguments which are more straightforward to relate to the increased SST effect.

d. What causes weaker convection at Bengkulu in the model?

Environment humidity, concentrated in the lower troposphere is known to impact moist convection significantly. Derbyshire et al. (2004) demonstrated through single column and cloud-resolving model simulations that a moist free troposphere results in strong, deep convection and a dry free troposphere results in shallow convection. In a more relevant study over western Sumatra, Sasaki et al. (2004) found a consistent moistening of the lower troposphere near the mountains, between 2 and 3 km, in the radiosonde data. They further argued that it has a significant impact on the diurnal cycle of precipitation in the region. Keil et al. (2008), using numerical experiments, showed that increasing the humidity in the midtroposphere (830–600 hPa) by 10% increases the 6-hourly convective precipitation by 200%. They also showed that an increase in the boundary layer humidity produces deeper convection than a similar increase in the midtroposphere humidity.

Figure 11a shows the time–height contours of humidity bias at Bengkulu, calculated with respect to the observed humidity. Clear dry bias is seen in the lower troposphere below $z = 4 \text{ km}$. About 10% of dry bias is

FIG. 10. As in Fig. 3b, but for the $SST_{\text{control}} + 2\text{ K}$ case at Bengkulu.

seen throughout the day between 1 and 3 km, except around 1900 LT when the dry bias is reduced to roughly 7%. We hypothesize that it is entrainment of the drier environment air in the model which makes the convection weaker by reducing the buoyancy of the updrafts. To demonstrate this, we have compared the buoyancy of an undiluted rising parcel to the one in which dilution is allowed at a fixed entrainment rate of $3 \times 10^{-4} \text{ m}^{-1}$. The calculations are performed at 2200 LT, which is the time when convection peaks in the model. The parcel buoyancy in the model is smaller than the observation, above the lifting condensation level, irrespective whether dilution is allowed or not. This can be seen in Fig. 11b, which shows the difference of parcel buoyancy in the observation and model ($B_{\text{OBS-MODEL}}$) for the undiluted (solid) and diluted (dashed) ascent. The positive values indicate higher buoyancy in the observation. It is noted that $B_{\text{OBS-MODEL}}$ for diluted ascent is greater than $B_{\text{OBS-MODEL}}$ for undiluted ascent between $z = 1$ and 5 km. Difference between the two first increases from $z = 1$ km to $z = 3$ km, coinciding with the region of maximum dry humidity bias, and then starts to decrease. Maximum difference of 0.3 K is found at $z = 3$ km.

That is, model loses more buoyancy when dilution is allowed and this loss is concentrated in the region of dry humidity bias seen in Fig. 11a, which supports our hypothesis. The reasons for this dry humidity bias are not fully understood, but it appears that the model inherited it, at least partly, from the initial condition. It is demonstrated in Fig. 11c, which shows the difference of mean specific humidity between ECMWF analysis and radiosonde at Bengkulu at 0000 UTC. The mean is calculated by averaging all 0000 UTC analyses and

radiosonde data over the simulation period. Clearly, ECMWF analysis also suffers from a dry bias in the lower troposphere, which peaks between 1 and 2 km to a maximum of 0.7 g kg^{-1} and a layer mean of roughly 0.4 g kg^{-1} .

e. What causes weaker convection at Mirai in the model?

The diurnal cycle of convection over *Mirai* is not as distinctive as at Bengkulu, which makes it harder to analyze. Therefore, instead of analyzing the mean quantities calculated as a composite of time of the day, composites of time series centered at a locally generated convective peak are analyzed for *Mirai*. The 90th percentile is set for both modeled and observed rainfall to identify the convective peaks, for the entire simulation period. The rainfall events that cross this threshold, and are at least 3 h apart, are finally considered for the analysis. A 3-h window is selected to remove any residual effect of one event on to another. This procedure resulted in 12 localized convective rainfall events in the observation and 8 in the model. All these events took place on different days suggesting that out of 20 days nearly half of the time both model and observation had very intense localized rainfall over *Mirai*.

The composited time series of all these events for precipitation, air temperature, specific humidity, and along-transect wind at 25 m above the sea level are shown in Fig. 12. As these events are generally short lived, a 10-min running mean is used to filter the observation dataset to make sure that no event is lost. For UM in Figs. 12 and 13, we use the same dataset at before, which are obtained using a circular averaging of 10-km radius and a 10-min mean. Results from circular averaging of a radius of 5 km are also indicated to show that the model behavior is unaffected by a smaller averaging radius.

Focusing on the thermodynamic conditions first, we see that the observed air temperature in Fig. 12b shows a sharp decrease, which starts about 1 h before the precipitation peak, followed by a slow increase. UM also shows a drop followed by a rise in the air temperature, but it is much slower compared to the observation. The rate of temperature drop is about 0.25 K h^{-1} in the simulation which is ten times smaller than what is observed (2.5 K h^{-1}). This drop in the air temperature is accompanied by a corresponding drop in specific humidity before the precipitation peak, followed by an increase (only in observation) as seen in Fig. 12c. As with the air temperature, the rate of change in specific humidity in the observation is much larger than the simulation. Such drops in environmental (near surface) temperature and specific humidity have been associated

FIG. 11. (a) Time–height contour of specific humidity bias (UM – OBS) in g kg^{-1} at Bengkulu; (b) difference in buoyancy (OBS – UM) of diluted (dashed) and undiluted (solid) parcel lifted from the surface layer ($z = 18$ m) at 2200 LT; and (c) vertical profile of mean specific humidity bias (analysis – OBS) at 0000 UTC in g kg^{-1} at Bengkulu.

with the passage of cold outflow from mature convective systems in the literature (Feng et al. 2015; Zuidema et al. 2017). Stronger temperature drops are accompanied by stronger decreases in the specific humidity in the observation, which is indicative of a stronger cold outflow resulting from a deeper convection. Passage of the front of this cold outflow, in the observation, is noticeable in Fig. 12c (marked with a circle) with humidity excess of about 0.2 g kg^{-1} at lag = -1 h. It corroborates with the earlier findings in Zuidema et al. (2017) and the references therein, which reported an excess of 0.25 g kg^{-1} at the cold outflow fronts. Its presence in Zuidema et al. (2017) has been attributed to the convergence of near surface moist air. Similar humidity anomaly can also be seen in the simulation with magnitude less than 0.1 g kg^{-1} (see the marked circle); however, as this anomaly is seen after precipitation has already started, it is questionable if it indicates the passage of the cold outflow front.

As for the dynamic conditions, Fig. 12d shows a composite of the along-transect horizontal wind for both model and observation. A positive value indicates landward wind and negative indicate offshore wind. The figure shows that the observed wind is always landward with a peak at lag = -1 h, followed by a sharp drop until lag = -0.5 h, after which it rises again until lag = 0. As this peak in wind speed coincides with the peak in specific humidity in Fig. 12c, it is likely that the subsequent drop of landward wind (about 1.4 m s^{-1}) from lag = -1 to -0.5 h (see the black line on the blue curve in Fig. 12d) is a result of the deceleration of the prevailing landward wind due to the opposing cold outflow. It is noted that the wind in the observation remains landward after converging with the cold outflow (but with a smaller speed, which suggest that cold outflow has probably become weak by the time it reached *Mirai*. Modeled wind shows a continuous drop in speed from

FIG. 12. Observed and simulated composite time series centered at local precipitation maximum of (a) precipitation rate, (b) air temperature 25 m above the sea level (MSL), (c) specific humidity 25 m MSL, and (d) along-transect horizontal wind speed at 25 m MSL. These maxima are identified based on a threshold of 90th percentile of both model and observation. See text for details. Circles in (c) and (d) indicate the passage of the cold front and the slope of the black lines in (d) indicate strength of the convergence between prevalent landward wind and the seaward cold front.

lag = -3 h to lag = -1 h at which point it reverses its direction to become seaward. If we assume for now (to be clarified in Fig. 13) that this change in direction marks the passage of the cold outflow in the simulation, then the subsequent change in speed of around -0.5 m s^{-1} from lag = -1 to lag = 0 (see the black line on the red curve in Fig. 12d) is indicative of the strength of convergence between the cold outflow and the prevalent landward wind in the model, which is clearly weaker than what is observed.

Another way to arrive at the same conclusion is to use the temperature drop in Fig. 12a as an indicator of the temperature, and hence buoyancy, difference between the cold outflow and the environment. As the propagation

speed of a cold outflow is directly proportional to the buoyancy difference between its body and the environment (Tompkins 2001), the higher buoyancy difference in the observation in Fig. 12a clearly suggest a faster-moving cold outflow, relative to the respective environment, in the observation than the simulation.

To illustrate model's behavior further, time–distance plot of these four variables on 4 December 2015, along the transects in Fig. 1b, are shown in Fig. 13. On this day, the model captured an intense rainfall event at 0300 LT at *Mirai*, which is one of the eight model events used in Fig. 12. Note that temperature and specific humidity in Fig. 13 are anomaly with respect to the daily mean. As noted in Fig. 12a, the time–distance plot also shows that

FIG. 13. Time–distance plot of variables in Fig. 12 is shown along the transects in Fig. 1b for model simulation on 4 Dec 2015. The solid and dashed vertical lines indicate the locations of *Mirai* and *Bengkulu*, respectively, and the horizontal dashed lines indicate 0300 LT.

Mirai encounters rain about an hour before the peak is reached at 0300 LT. This appears to have come from the direction of *Bengkulu*. Temperature in Fig. 13 is found to be decreasing before and after the precipitation peak. Specific humidity perturbation shows a peak just before 0230 LT, similar to the specific humidity anomaly in Fig. 12c. Furthermore, as noted in Fig. 12c, specific humidity continues to decrease until early morning. The along-transect wind in Fig. 13 also corroborates with the finding in the previous paragraph that the model wind speed in the landward direction continues to slow down until the cold outflow hits it at around 0130 LT, at which time it changes its direction to become seaward and then stays marginally seaward afterward.

In summary, we infer that precipitation over *Mirai* is triggered when the cold outflow from a neighboring convective system, most of which come from the direction of *Bengkulu* as seen in Fig. 2b, converges with the prevailing landward wind. This convergence takes place about an hour before the precipitation peak,

pushing the warmer environment air up. While the starting buoyancy of this updraft depends on the temperature and the specific humidity of the air at the time of convergence, the strength of convergence is related to the magnitude of the drop in the near-surface wind speed as discussed earlier. Assuming that the marked circle in Fig. 12c indicates the time of convergence, it is clear from Figs. 12b,c that the thermodynamic conditions of the initial updrafts are more favorable (warmer and moister) for deeper convection in the observation than in UM. Favorable thermodynamic conditions, together with a stronger convergence, explain the very intense and localized precipitation in the observation as compared to UM.

The reason for a colder and drier environmental air at the time of convergence is clearly the cold SST bias in the simulation. To understand the reason for a weaker (warm and slow) cold outflow in the simulation, we look at the vertical profiles of potential temperature around the precipitation peak in Fig. 14. A profile is marked as $T + 0$ if it falls within an hour of the precipitation peak; otherwise,

FIG. 14. Vertical profiles of potential temperature θ for (a) observation and (b) model composited around local precipitation maximum shown in Fig. 12a.

the event itself is removed from the list. This condition resulted in a decrease in the number of events in the observation to 9. The other two curves marked $T + 3$ and $T - 3$, are the profiles 3 h before and after the $T + 0$ profile. The observed profiles in Fig. 14a show that the boundary layer ($z < 300$ m), which was well mixed 3 h before the precipitation peak, has cooled down non-uniformly in the vertical to become a stable layer ($\Delta\theta/\Delta z \approx 0.01 \text{ K m}^{-1}$) by $T + 0$. As by this time the cold outflow has most likely already displaced the prevailing environmental air, we conclude that this stable layer depicts the temperature profile within the cold outflow body. Three hours later the boundary layer is seen transitioning back to a mixed layer as expected. Modeled profiles (see Fig. 14b) also show similar behavior that the boundary layer was well mixed 3 h before the precipitation peak, then it cooled down, and 3 h later it became well mixed again. The difference, however, is that the boundary layer in the model is nearly well mixed at $T + 0$, as opposed to being stable as in the observation. This vertical mixing causes the simulated cold outflow to be warmer than the observed cold outflow. As the speed of the cold outflow depends on the buoyancy difference between the environment and its own body, it is obvious that a warmer cold outflow would have a slower speed, as is found in the simulations.

To identify the origin of this mixing in the cold outflow body, we show the vertical profiles of simulated potential temperature, along the transect, at increasing distance from Bengkulu toward *Mirai* in Fig. 15.

The diurnal composite at 2230 LT, as shown in Fig. 8, is used to extract these profiles because by this time the cold outflow is seen to have reached the halfway point between *Mirai* and Bengkulu. Locations of these profiles are marked with the arrows in Fig. 8 at 2230 LT. Clearly, the temperature within the cold outflow body (see $z < 300$ m), which is stably stratified at Bengkulu, starts to mix in the vertical as soon as it reaches the sea surface 10 km away from Bengkulu. It is found to be well mixed (see the short dashed line) 10 km farther downstream over the sea surface.

It is reasonable to expect mixing over the warmer sea surface. However, it appears that the mixing is too rapid in the simulation as opposed to the observation. This situation of cold dense air overlying a warmer surface and then becoming mixed is similar to the daytime evolution of convective boundary layer over land. Drawing corollary from numerous studies on the evolution of the convective boundary layer, the two likely candidates to enhance vertical mixing in a convection-permitting model simulation are the diffusion in the turbulence scheme and the coarser resolution. The latter increases the mixing by enhancing entrainment across the inversion (Sullivan and Patton 2011). To check its sensitivity to the turbulence scheme, we have performed another experiment by reducing the turbulent coefficient by half. The resulting potential temperature profiles in Fig. 16 show a similar behavior as in Fig. 15. This suggests that it probably is the coarse model resolution which causes rapid mixing in the cold outflow

FIG. 15. Vertical profiles of potential temperature along the transect at indicated distance from Bengkulu at 2230 LT. Profiles correspond to the points marked by the arrows in the bottom panel of Fig. 8.

FIG. 16. As in Fig. 15, but for the simulation with reduced turbulent mixing coefficient.

as soon as it reaches the warmer sea surface. At this point of time we are unable to perform increased resolution simulations to demonstrate the same due to the high computational cost involved; however, we intend to pursue it in a future study.

4. Conclusions

The present article aims to diagnose reasons for the biases in the diurnal cycle of convection, over the western coast of Sumatra, at a convection-permitting (1.5 km) resolution using UM in the NWP mode. Focus is given on identifying issues in the input data from ECMWF high-resolution forecast, which can potentially affect the diurnal cycle, keeping in mind that the grid resolution and model physics also have an important control over these biases.

Data from the pilot field campaign of Years of the Maritime Continent (pre-YMC) are used to understand the model biases. The campaign was conducted nearly for a month from 22 November to 24 December of 2015 but only the first 20 days of data are used in the present article. Large-scale wind during these 20 days was found to be easterly in the lower troposphere, which assisted the convective systems generated over land to migrate over to the sea. This resulted in a distinct diurnal cycle of rainfall both on and offshore. During the campaign, 3-hourly radiosondes were released on a Research Vessel (R/V) *Mirai* and a land site, Bengkulu, only 55 km away from the R/V as shown in Fig. 1. Precipitation

characteristics at Bengkulu and *Mirai* are discussed in sections 3c, 3d, and 3e, respectively. Both observation and simulation results support the earlier findings that the convection over coastal land and sea is strongly coupled. Observed diurnal peak of precipitation at Bengkulu is noted in the evening around 1900 LT, which then moves offshore toward *Mirai*. The diurnal peak at *Mirai*, although not as distinct as at Bengkulu, is observed around 2230 LT. Simulated diurnal cycle agrees reasonably well with the observation but suffers from two biases at both the sites. These are (i) a delay of around 3 h in the diurnal peak and (ii) a weaker intensity. Findings of the present study are summarized below:

- Precipitation at Bengkulu is found to be triggered when the landward wind from the coastal sea merges with the local land breeze at Bengkulu. The reason for its delay is attributed to a delay in the onset of the local land breeze, which is found to be due to the weaker land-sea temperature contrast in the model and a delay in the arrival of the cold outflow from the adjacent mountain. The results show that the cold outflow at Bengkulu is a consequence of the evening precipitation over the mountain, which helps to develop a land breeze by cooling the air locally, as shown in Fig. 8. Using Fig. 3, it is shown that the delay in the arrival of the cold outflow at Bengkulu contributes to about 30% in the total delay; suggesting that the major reason for the delay in the onset of local land breeze at Bengkulu is the cold SST bias in the model (Fig. 4c). This is justified by performing another experiment by

increasing the model SST by 2 K. The results from this experiment, in Fig. 9, clearly show an improvement in the phase of the diurnal peak at Bengkulu, which is accompanied by an early onset of the local land breeze.

- Supporting the earlier findings, Figs. 2–3 show that the diurnal precipitation at *Mirai* is due to the propagating convective systems from the neighboring land. As the propagation speed of the offshore migrating system in the model (3.5 m s^{-1}) is nearly the same as in the observation (3.8 m s^{-1}), it is argued that the delay in the diurnal precipitation peak at *Mirai* is mainly due to the corresponding delay at Bengkulu.
- The reason for less intense rainfall at Bengkulu is attributed to the lower-tropospheric dry specific humidity bias (see Fig. 11a). Impact of the dry humidity bias is demonstrated by comparing the buoyancy of a parcel undergoing diluted ascent in the model and the observation in Fig. 11b. The figure shows that the entrainment of relatively drier air in the simulation decreases the buoyancy of the rising parcel, which is argued to cause weaker rainfall at *Mirai*. It is further argued that this dry humidity bias, if not fully, is partly from the ECMWF analysis.
- Using Figs. 8 and 12, it is shown that the deep convection at *Mirai* is a result of the convergence of the cold outflow (from the migrating convective system) with the prevailing landward wind. Both thermodynamic and dynamic conditions of the initial updrafts are found to be less favorable (cold, dry, and slow) for an intense convection in the simulation (see Fig. 12). It is argued that the near surface air in the simulation is colder and drier, as compared to the observation, because of the cold SST bias. A weaker cold outflow is found to be the reason for a weaker convergence in the simulation. Through Figs. 14–16, it is demonstrated that the weak cold outflow in the simulation is a result of an enhanced vertical mixing over the warm sea surface. It is further argued that the vertical mixing is stronger in the model because of its coarser resolution.

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