How Does the Air-Sea Coupling Frequency Affect Convection During the MJO Passage?

Ning Zhao1 © and Tomoe Nasuno1 ©

1Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan

Abstract  The importance of air-sea coupling in the simulation and prediction of the Madden-Julian Oscillation (MJO) has been well established. However, it remains unclear how air-sea coupling modulates the convection and related oceanic features on the subdaily scale. Based on a regional cloud-permitting coupled model, we evaluated the impact of the air-sea coupling on the convection during the convectively active phase of the MJO by varying the coupling frequency. The model successfully reproduced the atmospheric and oceanic variations observed by satellite and in situ measurements but with some quantitative biases. According to the sensitivity experiments, we found that stronger convection was mainly caused by the higher sea surface temperatures (SSTs) generated in high-frequency coupled experiments, especially when the coupling frequency was 1 hr or shorter. A lower coupling frequency would generate the phase lags in the diurnal cycle of SST and related turbulent heat fluxes. Our analyses further demonstrated that the phase-lagged diurnal cycle of SST suppressed deep convection through a decrease in daytime moistening in the lower troposphere. Meanwhile, in the upper ocean, the high-frequency air-sea coupling helped maintain the shallower mixed and isothermal layers by diurnal heating and cooling at the sea surface, which led to a higher mean SST. In contrast, the low-frequency coupled experiments underestimated the SST and therefore convective activities. Overall, our results demonstrated that high-frequency air-sea coupling (1 hr or shorter) could improve the reproducibility of the intensity and temporal variation in both diurnal convection and upper ocean processes.

Plain Language Summary  The Madden-Julian Oscillation (MJO) is one of the most important sources of atmospheric variability in tropical regions; however, even the modern numerical models could not well reproduce the MJO. We believe that the underestimation of air-sea coupling may cause some parts of such biases in simulating/predicting the MJO. Therefore, our study is aimed to uncover the impact of the air-sea coupling on convection and the related oceanic feature during the MJO. By varying the air-sea coupling frequency, our results showed that the 1-hr or higher-frequency coupled experiments had better performance due to the well-reproduced sea surface temperature, while suppressed convection was found in low-frequency coupled experiments. In general, our study suggested that the high-frequency air-sea coupling could improve the reproducibility of both convection and upper ocean features during the MJO.

1. Introduction

The Madden-Julian Oscillation (MJO) is the chief source of variability in tropical regions on the intraseasonal timescale, which features an eastward propagating convectively active envelope (Madden & Julian, 1972, 2005). Over the decades, increasing evidence has shown that the MJO not only influences the global climate system but also many types of extreme weather in the tropics and midlatitudes (e.g., Kayano & Kousky, 1999; Kessler, 2001; Lorenz & Hartmann, 2006; Wang & Moon, 2018; Zhang, 2013). Therefore, it is crucial to obtain the successful simulation/prediction of the MJO for tropical weather systems, extreme weather events, monsoons, and the El Niño–Southern Oscillation (Mishra et al., 2017; Vitart, 2014; Wu et al., 2019). To successfully simulate/forecast the MJO, numerous studies have been carried out; however, some systematic biases remained even in the state-of-the-art climate and forecast models due to the complexities of interactions between the multiscale atmospheric phenomena and the upper ocean related to the MJO (e.g., Peatman et al., 2014; Waliser et al., 1999; Zhang & Anderson, 2003). In particular, previous studies found that the moistening processes, which are crucial to the initiation of the MJO and its maintenance/propagation (e.g., Raymond & Fuchs, 2009; Zermeño et al., 2015), are highly correlated with the variations of the sea
Recent studies have made more progress toward understanding the effect of air-sea coupling on the MJO, especially on the subdaily scale (e.g., DeMott et al., 2015; Kim et al., 2018, and the references therein). For example, after highlighting the important role of the diurnally forced SST in the upper ocean variability (Bernie et al., 2005), Bernie et al. (2008) suggested the potential improvement in simulating the MJO by the inclusion of the high-frequency air-sea coupling. Furthermore, Seo et al. (2014) reported the improved representation of diurnal SST and the buildup of preconvection warming and moistening in a high-frequency coupled model. Based on in situ observations, Ruppert and Johnson (2015) demonstrated that diurnally varying SST could invigorate net column moistening aloft. Some studies also confirmed that the high-frequency air-sea coupling could enhance the reproducibility of convection and, therefore, the MJO (e.g., Chen & Zhang, 2019; Fu et al., 2013; Hagos et al., 2011).

Although the importance of air-sea coupling on convection and the MJO has been well established, few studies have examined detailed modulations on the diurnal scale (e.g., Crueger et al., 2013; Green et al., 2017; Neale & Slingo, 2003). Furthermore, some studies have shown that the air-sea coupling frequency may modulate the reproducibility of the diurnal cycle of SST (e.g., Bernie et al., 2005; Shinoda, 2005; Seo et al., 2014). It is reasonable to expect that the modulated SST may further influence the subdaily moistening processes and the MJO (e.g., Hagos et al., 2016; Katsumata et al., 2018; Ruppert & Johnson, 2015). However, such modulations on the SST by air-sea coupling were neglected in most studies and, therefore, remained unclear.

Thus, the goal of this study is to evaluate the effect of air-sea coupling (and the coupling frequency) on convective activities and related upper oceanic variations, especially on the subdaily scale. As part of the ongoing Year of the Maritime Continent (YMC) project (Nasuno, 2019; Yokoi et al., 2017, 2019, Wu et al., 2019), we focus on one MJO event (26 November to 4 December 2017; Figure 1b) captured during the YMC-Sumatra 2017 field campaign. In this period, a large number of land- and ship-based in situ observations are available, providing a great opportunity to validate the capability of our numerical experiments.
This paper is organized as follows. Section 2 includes descriptions of the model setup, data source, and sensitivity experimental designs. Section 3 includes validation of the model performance with and without cumulus parameterizations. Section 4 documents the impact of coupling frequencies on the convection and surface conditions, along with analyses of the heat and moisture budgets. In section 5, the role of the daily mean SST and how the SST was modulated by the coupling frequency are discussed. Finally, a summary of major findings is presented in section 6. Results on the cumulus parameterizations, extra experiment for the role of local SST, and the land-sea connections are presented in sections S1–S3 in the supporting information, respectively.

2. Model and Experiment Settings

2.1. Model

In this study, numerical experiments were based on the Coupled-Ocean-Atmosphere-Wave-Sediment Transport Modeling System (Warner et al., 2010), and, for simplicity, we excluded the wave and sediment components and only activated the atmospheric (Weather Research and Forecasting Model, WRF V4.1.2) and oceanic (Regional Ocean Modeling System, ROMS svn 980) components.

The model was designed to cover the region from 78°E to 122°E and 14°S to 14°N, where the center is located at Sumatra Island (0°, 100°E; Figure 1a). All simulations started at 0:00 UTC on 21 November and ran until 0:00 UTC on 6 December, which was 5 days prior to the active phase of the MJO over the Maritime Continent (Phases 4 and 5, Figure 1b). The first day of each simulation was regarded as the spin-up period and was not used for analyses.

The horizontal resolution of WRF and ROMS is 7 km with matching grids and land-sea masks (Nasuno 2019). The 30-s Global Multiresolution Terrain Elevation Data 2010 (Danielson & Gesch, 2011) and ETOPO1 (Amante & Eakins, 2009) were used for WRF and ROMS, respectively. WRF has 45 sigma layers from the surface to the top (50 hPa) with about 20 layers under 850 hPa, and ROMS has 50 layers based on the quadratic Legendre polynomial function (Souza et al., 2015) with over 18 vertical levels in the upper 50 m on average and the first layer was at about 0.3-m depth.

WRF uses the single-moment 7-class microphysics scheme (Bae et al., 2018), the Yonsei University PBL scheme (Hong et al., 2006), the Revised MM5 surface layer scheme (Jiménez et al., 2012), the Unified Noah Land Surface Model (Tewari et al., 2004), the RRTMG Shortwave and Longwave Schemes (Iacono et al., 2008), and the Grell-Freitas Ensemble (GFE) cumulus scheme (Grell & Freitas, 2014). ROMS uses the Mellor-Yamada Level 2.5 closure scheme associated with the third-order upstream horizontal advection, harmonic horizontal mixing, and fourth-order centered vertical advection, and no nudging term and tidal forcing were considered in this study.

It is worth noting that at a so-called gray zone resolution (Gerard, 2007), the cumulus parameterization does not always enhance the reproducibility in our 7-km model. Thus, to obtain the best model performance, several cumulus schemes were tested before the sensitivity experiments of coupling frequency, and these tests can be found in the supporting information. The schemes we tested included the GFE scheme (Grell & Freitas, 2014), the New Simplified Arakawa-Schubert scheme (Kwon & Hong, 2017), and the Multi-scale Kain-Fritsch scheme (Zheng et al., 2016). We also tested the New Tiedtke (NT) scheme (Zhang & Wang, 2017), which is not scale-aware but includes both deep and shallow cumulus components. Note that because the New Simplified Arakawa-Schubert scheme does not have the shallow convection component, which was proved to be important in simulating convection (Hagos et al., 2011; Pilon et al., 2016), we applied the Global/Regional Integrated Model system shallow convection scheme (Hong & Jang, 2018) in CP1HC3 following Kwon and Hong (2017). Finally, according to these extra experiments, the GFE scheme showed the best performance on both intensity and the diurnal cycle of the precipitation and was applied in all experiments of this study (see Figures S1 and S2 in the supporting information).

2.2. Data

In this study, the National Centers for Environmental Prediction (NCEP) Final (FNL) Operational Global Analysis data (NCEP, 2000) and Global Ocean Forecasting System (GOFS) 3.1 (Cummings, 2005) were used as the initial and lateral boundary conditions for WRF and ROMS, respectively. Additionally, in uncoupled experiments, the daily optimum interpolation SST (OISST) data set was used for the lower boundary condition (Banzon et al., 2016; Reynolds et al., 2007). The modeled atmospheric and oceanic properties were saved every 1 hr over the course of each computation in all experiments.
Table 1

Descriptions of Sensitivity Experiments

| Experiment | Settings                                      |
|------------|-----------------------------------------------|
| CP30MC     | 30-min WRF-ROMS                               |
| CP1HC      | 1-hr WRF-ROMS                                 |
| CP3HC      | 3-hr WRF-ROMS                                 |
| CP6HC      | 6-hr WRF-ROMS                                 |
| CP12HC     | 12-hr WRF-ROMS                                |
| CP1DC      | 1-day WRF-ROMS                                |
| NOCPC      | WRF only (OISST)                              |
| NOCPC+     | WRF only (OISST and CP1HC daily mean SST within Sumatra region) |
| NOCPC++    | WRF only (CP1HC daily mean SST everywhere)    |

For the model validation, we used the high-resolution satellite-based precipitation data (0.1° and hourly) from the Global Rainfall Map (GSMaP) data set together with ship-based (research vessel [R/V] Mirai) and land-based (Bengkulu, Indonesia) radiosonde data obtained during the YMC-Sumatra 2017 field campaign (Nasuno, 2019; Yokoi et al., 2019; Wu et al., 2019). The radiosondes were launched eight times per day, and the measurement was saved every 1 s during the ascending.

2.3. Sensitivity Experiments of Coupling Frequency

To investigate the influence of the coupling frequency on convection during the MJO active phase, a set of experiments was carried out by varying the coupling interval from 30 min to 1 day and three uncoupled (WRF only) experiments, while all the other settings remained identical. ROMS and WRF were coupled at the first time step of each experiment and then coupled after the specified intervals. For example, the 6-hourly coupled model (i.e., CP6HC) would exchange atmospheric and oceanic information at 0:00 UTC, 6:00 UTC, 12:00 UTC, and 18:00 UTC. In all coupled experiments, WRF uses the SST calculated in ROMS, while ROMS receives heat fluxes, wind stress, surface temperature, relative humidity (RH), and freshwater fluxes from WRF. In addition, three uncoupled (WRF only) experiments were further conducted to discuss the role of daily mean SST with different daily mean SSTs. Specifically, we used the OISST for the NOCPC experiment, and we further replaced the OISST by the daily mean SST obtained from CP1HC in the other two experiments (within the Sumatra region and the whole domain, respectively). Detailed descriptions of the sensitivity experiments can be found in Table 1.

3. Model Validation

Figure 2 shows the horizontal distributions of the mean precipitation rates and maximum precipitation time during the active phase of the MJO (26 November to 4 December) obtained by satellite and CP1HC. The heaviest rain occurred over the Gulf of Thailand near the eastern coast of the Malay Peninsula and the southern Andaman Sea (Burma Sea), along with a weak but widely distributed rainy zone covering the Indian Ocean and the Maritime Continent (Figure 2a). In the Southern Hemisphere, clear rainy zones were observed over the Indian Ocean, south of the equator and south of Java Island, but the precipitation rates were smaller. All the patterns mentioned above were successfully reproduced in CP1HC, although the rainy regions were not as concentrated as the observations in the Southern Hemisphere. It should be noticed that our model slightly underestimated the rainfall in the large area west of northern Sumatra, although it was quite weak.

In addition to the precipitation rate, the time of maximum precipitation can also be regarded as an important indicator representing the diurnal cycle of precipitation. Note that, in this study, we determined the time based on the diurnal composite of precipitation. As shown in Figure 2c, the maximum precipitation mainly occurred in the evening near the coastal region and in the midnight inland (see the patterns over Sumatra and Kalimantan). Meanwhile, over the shallow coastal seas, the heaviest rain occurred in the early morning, especially near the Gulf of Thailand and west of Sumatra Island where the heaviest rainfall was observed.

In general, the observed diurnal cycle was consistent with the diurnal cycle revealed in previous studies (Mori et al., 2004; Neale & Slingo, 2003). Moreover, in comparison with the observations, the precipitation
in CP1HC reached its maximum rate at the same time (Figure 2d), although the simulated precipitation showed more small-scale features.

In addition, we further compared our model results with the in situ radiosonde profiles obtained during the YMC-Sumatra 2017 field campaign (Figure 3; also see Figure 1a for the locations of radiosonde observations). The atmosphere near Sumatra Island was dominated by the westerly wind over the entire active phase of the MJO, which could also be seen in the model. The meridional wind component ($v$) suggested that a transition of wind field occurred after 30 November, as shown by the opposite meridional wind direction before and after the day. A similar transition of the meridional wind could also be found in our model, but with some underestimations, especially the northerlies near Bengkulu prior to the active phase of MJO (Figures 3c and 3d). Moreover, both observations and simulations showed that very high RH was dominant from the surface to the upper troposphere, indicating that the vigorous convective activities occurred in both the model and real atmosphere. Note that the model results were obtained at the same locations of radiosonde observation based on the bilinear interpolation.

Overall, our model showed good agreement with both satellite-based and in situ observations, suggesting that our model is reliable; therefore, we conducted sensitivity experiments of coupling frequencies using the same schemes and settings as CP1HC.

4. Results
4.1. The Impact on the Atmosphere
To examine the impact of the air-sea coupling frequency on convection, we first focused on the moisture and low-level wind (850 hPa) in our sensitivity experiments. Figure 4a presents the map of the daily mean precipitable water (PW) and wind averaged during the active phase of the MJO (from surface to 300 hPa, 26 November to 4 December) in CP1HC. A large amount of PW was concentrated north of Sumatra Island.
Figure 3. Vertical profiles of wind (a–d) and relative humidity (e–h) obtained by radiosonde observations (left panels) obtained during the YMC-Sumatra 2017 field campaign and CP1HC (right panels).

and the Malay Peninsula, which consists of the location of the heaviest rainfall, as observed by satellite and CP1HC (Figure 2). The cyclonic gyre located in both north and south of the equator, associated with the jet-like wind originating from the Indian Ocean to the west (also from the South China Sea), is consistent with the existence of vigorous convective activities over the Maritime Continent (i.e., the active phase of the MJO).

In comparison with CP1HC, the differences in the mean PW and low-level wind were not obviously changed in CP30MC, CP3HC, or CP6HC. However, unlike the high-frequency coupled experiments (i.e., 6 hr or more coupled), the total amount of PW decreased by 10% or more (>5 mm) in most regions in the low-frequency coupled models, and the clearest reduction was found in the uncoupled experiment (NOCPC, Figure 4g). Meanwhile, the wind anomalies showed anticyclonic gyre-like and westward jet-like patterns, indicating greatly suppressed convective activities (Figures 4e–4g). Note that although the anticyclonic gyre-like pattern could also be seen in Figure 4d, no westward anomalies were found, suggesting that the convection was weakened in the CP6HC but not as much as that in the low-frequency coupled/uncoupled experiments.

Figure 5a presents the frequency-altitude distributions of RH over the seas near Sumatra Island (Nasuno et al., 2015). The lower troposphere (surface to 800-hPa level) was characterized by frequent occurrences of 80–100% RH. The frequency in the middle troposphere (850–500 hPa) ranged from 50% to 100% RH,
Figure 4. Horizontal distributions of the precipitable water (integrated from the surface to 300-hPa height) and the wind field/anomalies at 850-hPa level during the active phase of the MJO (26 November to 4 December) in (a) CP1HC and (b–i) the differences between CP1HC and other runs. The increased/reduced percentages of $\Delta PW$ were based on CP1HC, and only values within 95% confidence level were plotted.

where 70–80% RH occurred the most. Above the 500-hPa level, the atmosphere became increasingly drier as the height increased, partially because we calculated the RH based on the water vapor mixing ratio only (i.e., without the ice). The other high-frequency coupled experiments showed only small differences (Figures 5b–5d) from CP1HC. Specifically, in CP30MC, the increased frequency of high RH (>70%) suggested that the convective activities were more enhanced (e.g., DeMott et al., 2019). However, this was not the case in the low-frequency coupled/uncoupled experiments. The occurrences of high RH were greatly reduced from the lower troposphere to the upper levels, and >90% RH was nearly extinct in the midlevels (Figures 5e–5g). Accordingly, low RH appeared more frequently in nearly the entire column above the atmospheric boundary layer in the low-frequency coupled/uncoupled experiments, suggesting that convection was greatly weakened.

One may consider that the reduction in high RH may be caused by the modulated preconditioning of the MJO (e.g., Seo et al., 2014). However, our models showed different results. As shown in Figure 6, the occurrences of high RH (>70%) showed no significant differences before the MJO, even at in midlevels (Figure 6a). Nevertheless, the situation started to change only after the MJO entered Phase 4 (26–30 November), although the SSTS had already been modulated (i.e., became cooler) since the model initiation (see Figure S3 in the supporting information). In particular, the atmosphere was greatly moistened in the high-frequency coupled experiments (>90% occurrence of high RH), which was consistent with the vigorous convection during the MJO. Nonetheless, the low-frequency coupled/uncoupled experiments showed relatively lower values, as suggested by the clear separation in the middle troposphere among the experiments.
Figure 5. (a) The frequency-altitude distributions of the simulated relative humidity over the seas around Sumatra Island and (b-i) the differences between other experiments and CP1HC (10° S–10° N, 90°–110° E; red box in Figure 1a).

The occurrence of high RH in the low-frequency coupled/uncoupled experiments was approximately 10% lower than that in the high-frequency coupled experiments. As the MJO propagated to the east (after 1 December, i.e., Phase 5 of the MJO), convection was suppressed in all experiments (see the descending trend in Figure 6), but the differences became larger. The low-level moisture divergence also suggested the similar results. Overall, our results suggested that the high-frequency air-sea coupling enhanced the convective activities during the active phase of the MJO, and it also helped with the maintenance of the moist atmosphere after the MJO passed.

4.2. Modulated Diurnal Cycle at the Sea Surface

Figure 7 represents the diurnal composite of surface variables averaged in the Sumatra region (red box in Figure 1a and ocean only). The SSTs reached over 29.5°C in the high-frequency coupled experiments, along with a clear diurnal cycle that was absent in the low-frequency coupled/uncoupled experiments. Although the daily mean SST and its diurnal amplitude were nearly identical among the high-frequency coupled experiments (Table 2), the temporal variations were not. In CP1HC and CP30MC, the largest SST appeared at 8:00 UTC, which was consistent with a recent study based on buoy data (Morak-Bozzo et al., 2016). However,
Figure 6. Time series of frequencies (occurrences) of the grid with high relative humidity (> 70%) at the (a) 500-hPa level, (b) 700-hPa level, (c) 900-hPa level, and (d) low-level (1,000–800 hPa) accumulated divergence of moisture flux. The gray shading indicates the active phase of MJO. Note that the high-frequency coupled experiments were colored by warm colors and low-frequency coupled experiments were colored by cool colors, except the NOCPC++ (purple), which used the daily mean SSTs from CP1HC.

Table 2
Daily Mean SST and TFLX in the Sumatra Region

| Experiment | SST (°C) | TFLX (W/m²) |
|------------|----------|-------------|
| CP30MC     | 29.02    | 161.04      |
| CP1HC      | 29.02    | 161.30      |
| CP3HC      | 29.01    | 163.85      |
| CP6HC      | 28.96    | 158.55      |
| CP12HC     | 28.17    | 120.02      |
| CP1DC      | 28.36    | 135.95      |
| NOCPC      | 28.43    | 122.22      |
| NOCPC+     | 29.00    | 155.03      |
| NOCPC++    | 29.00    | 161.87      |

this was not the case in CP3HC and CP6HC, where the diurnal cycle of SST was delayed by 1 and 4 hr, respectively. On the other hand, CP12HC and CP1DC did not have any diurnal cycle of SST, and the mean values were also lower than those of high-frequency coupled experiments.

The surface turbulent heat fluxes (TFLX; upward positive), which are mainly controlled by the latent heat flux (see Figure S5 in the supporting information), had more complex variations. The TFLX in CP30MC started increasing at 16:00 UTC and finally reached its maximum at 7:00 UTC (14:00 LT in the Sumatra
Figure 7. Diurnal composites of (a) sea surface temperature, (b) mixed-layer temperature, (c) turbulent heat fluxes (the sum of surface sensible and latent heat fluxes, upward positive), (d) surface air temperature at 2 m, (e) specific humidity at 2 m, and (f) surface wind speed at 10 m. Colors represent different experiments. Note that the correspondence of color and experiment is consistent in this paper, except in Figure 13.

Among the high-frequency coupled experiments, CP6HC showed totally different trends: the TFLX monotonically decreased during the daytime and then suddenly increased after 12:00 UTC. Such unrealistic variations were generated by the coupling procedure used in our model. Except for the time step of coupling, both WRF and ROMS used a “constant” boundary forcing at most of the time steps. For example, in the CP6HC, the increasing SST from 6:00 UTC to 12:00 UTC only occurred in the ocean (ROMS), while WRF used the temporally constant SST (which was obtained at 6:00 UTC) until the next coupling time. Therefore, the underestimation and overestimation occurred continuously due to the “constant” forcing.

Unlike the SST and TFLX, the surface air temperature (T2m) and specific humidity (q2m) experienced the same diurnal cycle in all experiments, although they were 0.5°C and 0.25 g/kg higher in the high-frequency coupled experiments (Figures 7d and 7e). Despite the unique variations in CP6HC, the heating of surface air started at 1:00 UTC, leading to the highest T2m at 9:00 UTC. The increase in q2m started slightly later at 3:00 UTC and then reached its maximum at 13:00 UTC (Figure 7e).

The surface wind speed had nearly no diurnal cycle regardless of coupling frequencies, but it was 1 m/s higher in the high-frequency coupled experiments (Figure 7f). The bimodal variations in T2m and its earlier increase than the SST in the CP6HC were mainly caused by phase-delayed heating from the ocean after 12:00 UTC, which alleviated the nighttime temperature decrease. Note that the diurnal amplitude of SST region), followed by a 9-hr decrease. Similar variations could be found in CP1HC, although its largest TFLX appeared 1 hr later and smaller. The lag became larger in CP3HC, along with a 2-hr period fluctuation caused by the different response times of surface air temperature (T2m) and specific humidity (q2m) (Figures 7d and 7e).

Among the high-frequency coupled experiments, CP6HC showed totally different trends: the TFLX monotonically decreased during the daytime and then suddenly increased after 12:00 UTC. Such unrealistic variations were generated by the coupling procedure used in our model. Except for the time step of coupling, both WRF and ROMS used a “constant” boundary forcing at most of the time steps. For example, in the CP6HC, the increasing SST from 6:00 UTC to 12:00 UTC only occurred in the ocean (ROMS), while WRF used the temporally constant SST (which was obtained at 6:00 UTC) until the next coupling time. Therefore, the underestimation and overestimation occurred continuously due to the “constant” forcing.

Unlike the SST and TFLX, the surface air temperature (T2m) and specific humidity (q2m) experienced the same diurnal cycle in all experiments, although they were 0.5°C and 0.25 g/kg higher in the high-frequency coupled experiments (Figures 7d and 7e). Despite the unique variations in CP6HC, the heating of surface air started at 1:00 UTC, leading to the highest T2m at 9:00 UTC. The increase in q2m started slightly later at 3:00 UTC and then reached its maximum at 13:00 UTC (Figure 7e).

The surface wind speed had nearly no diurnal cycle regardless of coupling frequencies, but it was 1 m/s higher in the high-frequency coupled experiments (Figure 7f). The bimodal variations in T2m and its earlier increase than the SST in the CP6HC were mainly caused by phase-delayed heating from the ocean after 12:00 UTC, which alleviated the nighttime temperature decrease. Note that the diurnal amplitude of SST
became smaller during the Phase 4 (∼0.4° C) due to the relative stronger surface wind (∼6.5 m/s) and related stronger vertical mixing, and as the wind became weaker, the amplitude increased to about 0.7° C in Phase 5 (see Figure S3 in the supporting information). In general, our results suggested that the strong surface wind did not influence the daily mean SSTs obviously, although it may induce strong vertical mixing within the upper ocean and reduce the diurnal amplitude of SSTs. Therefore, the impact of coupling frequency is unlikely sensitive to the wind condition.

Without the diurnally varying surface forcing, the SSTs were nearly constant in the low-frequency coupled experiments. Unlike the increasing trends during the daytime in the high-frequency coupled experiments, their TFLXs decreased during the daytime and increased during the nighttime, following T2m and q2m with an opposite sign (Figures 7d and 7e). Such variations were similar to the uncoupled experiment (NOCPC). Note that the relatively higher daily mean SST and TFLX in CP1DC (comparing to CP12HC) were caused by the nonzero solar radiation throughout the day, while the heating was updated to zero in CP12HC after 12:00 UTC (Figure 12c). Moreover, the influence of the mean SSTs and its diurnal cycle can be found in section 5.

4.3. Heat and Moisture Budget Analysis

To further elucidate the influence of the coupling frequency on subdaily moistening processes, we executed an area-averaged heat and moisture budget analysis. We rearranged the budget equations including the apparent heat source (Q1) and moisture sink (Q2) following Yanai et al. (1973):

$$\frac{\partial s}{\partial t} \equiv Q_1 - \tilde{U} \cdot \nabla s - \omega \frac{\partial s}{\partial p},$$

$$L_v \frac{\partial q}{\partial t} \equiv -Q_2 - L_v \tilde{U} \cdot \nabla q - L_v \omega \frac{\partial q}{\partial p},$$

where $s \equiv c_p T + g z$ is the dry static energy, $c_p$ is the specific heat at constant pressure, $T$ is the temperature, $q$ is the specific humidity, $L_v$ is the latent heat of condensation, and $\tilde{U}$ and $\omega$ are the horizontal wind vector and the vertical wind component in pressure coordinates, respectively. All terms were calculated based on the hourly output from the model and averaged over the Sumatra region (ocean only, red box in Figure 1a) during the MJO active phase.

Figure 8 represents the diurnal composite time-altitude distributions of the $s$ and $q$ budgets averaged in the Sumatra region during the MJO active phase based on CP1HC. The atmosphere became warmer during the local daytime and cooler during the local nighttime, shifting its phase at 12:00 UTC (Figure 8a). On the other hand, diurnal moistening mainly occurred from 4:00 UTC (11:00 LT) at surface levels, and such moistening generally takes 4–5 hr to extend to the entire lower levels (Figure 8b). Although the positive Q1 could be seen all day long during the MJO, the strong heating started at 16:00 UTC (23:00 LT) in the middle troposphere and reached its maximum at 4:00 UTC (Figure 8c), while the corresponding moisture sink occurred slightly earlier at lower levels before the heating started (Figure 8d), which well corresponded to the precipitation occurred over the ocean (Figure S8 in the supporting information). Both Q1 and Q2 were basically balanced by vertical advection ($s_{\text{adv}}$ and $q_{\text{adv}}$, Figures 8g and 8h, respectively), indicating the existence of vigorous convection, while horizontal moisture advection ($q_{\text{radv}}$, Figure 8f) tended to dry the atmosphere due to the background eastward wind during the MJO (Figure 4a). Note that, in this study, we only focused on the air-sea interaction and convection above the ocean; however, the dynamics revealed here are also highly connected with those over the land (see section S3 in the supporting information for details).

Although the differences between CP6HC and CP1HC were small in the PW and high RH occurrences (Figures 4 and 5), the budget analyses showed more detailed modulations. It is easy to find that the heating and moistening were stronger during the nighttime but became weaker during the daytime in the CP6HC (Figures 9a and 9b), which was likely due to the phase-lagged SST and related TFLX. In addition to the phase lag in the diurnal processes, both Q1 and Q2 were weakened, associated with the reduced vertical advection of moisture. It is suggested that the phase-delayed diurnal cycle of SST greatly weakened the daytime convection and slightly enhanced the nighttime convection, resulting in a net reduction in daily mean state.
In the uncoupled run, Q1 and Q2 were significantly reduced over 0.1 K/hr, associated with the weakening in vertical advection (Figure 10). Furthermore, the positive anomalies in horizontal moisture advection \( q_{\text{adv}} \) also indicated that the eastward moisture transport was reduced, which was consistent with the easterly low-level wind anomalies, as shown in Figure 4g. As a result, the atmosphere became warmer and drier during the daytime and cooler and moister during the nighttime in NOCPC (Figures 10c and 10d). In general, the results shown above suggested that the convection and diurnal heat/moisture processes were greatly modulated when using the daily mean SST; however, to investigate the roles of the mean SST and its diurnal cycle, further experiments were needed. Note that similar results were found in low-frequency coupled experiments.

5. Discussion

5.1. The Role of the Daily Mean SST

In our model, WRF received the SST from ROMS only. The SSTs in CP12HC, CP1DC, and NOCPC were >0.6°C lower than those in the high-frequency coupled experiments, which may therefore be the key factor in the
weakened convective activities (e.g., Dipankar et al., 2019). To confirm this, two extra WRF-only experiments were carried out by using the same settings in NOCPC but with the daily mean SST from CP1HC. One may notice that the SST in NOCPC+/NOCPC++ was slightly lower than that in CP1HC, which was induced by the linear interpolation of the daily mean SST into the 6-hourly SST, following the same procedure for the OISST in NOCPC. Fortunately, our results show that the biases were small and negligible (∼0.02°C, Table 2). Note that we only discussed the role of daily mean SST based on NOCPC++ here, and readers may refer to section S2 for more results of NOCPC+ (role of the local SST).

In general, the higher daily mean SST induced the higher TFLX (Table 2 and Figure 7), more active convection (Figure 6), and therefore a moister atmosphere (Figure 4), which was almost comparable with those in high-frequency coupled experiments. Furthermore, with the same daily varying SST, the time evolution of convective activities also showed good agreements with the high-frequency coupled models during the MJO. The anticyclonic gyre-like pattern of low-level wind anomalies remained, along with the negative but small biases in PW, although its center moved to the south. It is suggested that the underestimation of convection remained in NOCPC++, even with the same daily mean SST.

**Figure 9.** Same as in Figure 8 but for the differences between CP6HC and CP1HC. Only values within 90% confidence level were plotted based on Student’s $t$ test.
As showed in Figure 11, both Q1 and Q2 had small but nonnegligible negative anomalies from the surface to the midlevels. Moreover, unlike the monotonic reductions in NOCPC, the differences in NOCPC++ showed more subdaily variations. We found that the lower SST induced a drier boundary layer in NOCPC++ during the daytime, while the higher SST induced a moister boundary layer during the nighttime. Previous studies suggested that the pre-moistening of the lower troposphere is an important feature that can promote deep convection (Katsumata et al., 2018; Ruppert & Johnson, 2015; Shinoda & Uyeda, 2002). Therefore, it is likely that this weakened moistening in the lower levels of NOCPC++ (approximately 5:00–12:00 UTC, Figure 11b) suppressed the onset of the subsequent diurnal deep convection, resulting in the negative Q1/Q2 and related vertical advection.

Based on the results of CP6HC and NOCPC++, the daily mean SST did play a dominant role in controlling the convection intensity, although the higher daytime SST (hence, the diurnal cycle of SST) played a smaller but nonnegligible role in daytime moistening and therefore the onset of diurnal deep convection. Figure 6d also demonstrates the above conclusion that while the higher daily mean SST induced larger
moisture convergence, small but clear negative biases could be seen after 8:00 UTC in NOCPC++ and CP6HC, exhibiting weakened diurnal convection. Note that all high-frequency coupled experiments showed little differences with CP1HC, while the results of CP12HC and CP1DC were similar as NOCPC but with different magnitudes.

5.2. Modulations in the Upper Ocean

Since we found that the diurnal cycle of SST played a smaller role than the daily mean SST, one may ask why the high-frequency coupled experiments had higher daily mean SSTs. To answer this question, we focused on the modulations in the upper ocean by air-sea coupling, especially the dynamics of the oceanic mixed layer and upper isothermal layer.

In this study, we defined the oceanic mixed-layer depth (MLD) in terms of a depth with a density equal to that at the 1-m depth plus an increment in density equivalent to −0.2°C (Moteki et al., 2018), and therefore, the isotherm depth (ILD) is defined as the depth where the temperature is 0.2°C lower than that at 1-m depth. Note that the results were not significantly changed when the reference depth was set to 10 m.
Figure 12. Diurnal composites of (a) the mixed-layer (solid line) and isothermal-layer depth (dashed line), (b) thicknesses of the barrier layer, and (c) the net heat flux at the sea surface (downward positive). (d) The lead-lag correlations of mixed-layer depth (solid line) and mixed-layer temperature (dashed line) corresponding to the SST.

Table 3

| Experiment | MLT (°C) | MLD (m) | ILD (m) | BL (m) | Net heat flux at sea surface (W/m²) |
|------------|----------|---------|---------|--------|----------------------------------|
| CP30MC     | 28.98    | 28.17   | 37.89   | 9.72   | 27.27                            |
| CP1HC      | 28.98    | 27.78   | 37.89   | 10.11  | 28.15                            |
| CP3HC      | 28.97    | 27.54   | 38.05   | 10.51  | 24.65                            |
| CP6HC      | 28.91    | 27.56   | 39.18   | 11.62  | 17.54                            |
| CP12HC     | 28.23    | 36.86   | 49.11   | 12.25  | −99.64                           |
| CP1DC      | 28.41    | 33.90   | 47.35   | 13.45  | −72.49                           |

As shown in Figure 12b, the mixed-layer temperature (MLT, vertically averaged within the mixed layer) in the high-frequency coupled experiments was approximately 29.0° C, which was 0.8° C (0.6° C) higher than that in CP12HC (CP1DC). Both MLT and MLD had a weak but clear diurnal cycle, indicating the existence of stratification and destratification induced by the surface heating/cooling and mixing processes. The mixed layer became warmer and shallower after the sea surface was heated during the daytime (Figure 12a), and the largest MLT appeared 2 to 3 hr after the SST reached its maximum (Figures 7a and 7b), which was the time required by the adjustment processes (Figure 12d). Similar diurnal variations were found in ILDs, although it was generally over 10 m deeper than the MLD in all coupled experiments (Figure 12a).

Our results suggested that the mixed-layer dynamics could be greatly modulated with or without high-frequency air-sea coupling (Figure 12d). In low-frequency coupled experiments, the MLTs were relatively higher than the SSTs (Figures 7a and 7b) because the ocean experienced net heat loss at the sea surface throughout the day (Figure 12c and Table 3). It is reasonable to consider that continuous surface cooling reduced the SST and broke down the upper layer instability, inducing strong vertical mixing and therefore the deepening of MLD/ILDs. Furthermore, the deeper MLD further reduced the MLT. On the other hand, in high-frequency coupled experiments, the net heat gain during the daytime would raise the SST and enhance the stratification that suppresses the mixing, leading to a shallower MLD/ILD (Figure 12a). It is easy to find that the high-frequency air-sea coupling helped to maintain the higher daily mean SSTs/MLTs and shallower MLDs/ILDs.
5.3. The Drift of SST

In previous sections, we found that the high-frequency coupled model appeared to have warmer SSTs, which even warmer than the satellite observations (OISST). It is reasonable to consider that the SSTs may be overestimated in our models. On the other hand, some studies also suggested that the satellite-based SSTs may be underestimated due to the heavily spatial smoothing and the removal of diurnal variations (Clayson & Bogdanoff, 2013; Reynolds et al., 2007). Therefore, it is necessary to validate them with the in situ observations. In this study, we used the temperature measured by the R/V Mirai Surface Meteorological observation system at 5-m depth (SBE38, SeaBird Electronics), and details can be found in the MIRAI MR17-08 Cruise report (JAMSTEC & BPPT, 2018). Note that the modeled SSTs were nearly identical to the 5-m temperature due to the deep ILDs.

Figure 13 shows the time series of SSTs obtained in our models, satellite, and R/V Mirai during the active phase of the MJO, together with the precipitation rate observed by the optical rain gauge (ORG-815DR, Osi). The OISST (NOCPC) showed negative biases most times, while the models overestimated in some periods. The overestimation of SST on 28 November was mainly related to heavy rainfall, while the period from 3 December was related to the underestimation of convection and its related precipitation (hence, surface cooling) as showed in Figures 3e and 3f. Nevertheless, both OISST and our modeled SSTs generally followed the trend of in situ observations in the same order. In addition, as mentioned in section 3, our model showed good agreement with both satellite-based and in situ observations. Therefore, our conclusions on the importance of high-frequency air-sea coupling (and the higher and diurnally varying SST generated by that) are robust and reliable.

6. Concluding Remarks

A regional cloud-permitting coupled model was conducted to evaluate the impact of coupling frequency on convective activities during an MJO event captured in the YMC-Sumatra 2017 field campaign. By activating the scale-aware GFE cumulus scheme, the 1-hourly coupled model showed good agreements with both satellite-based precipitation and in situ radiosonde observations. Thus, a set of sensitivity experiments was carried out to investigate the impact of the air-sea coupling frequency on convective activities during the MJO.

By varying the coupling frequency from 30 min to 1 day, we found that the PW in the atmosphere was largely reduced in the low-frequency coupled experiments (12-hourly or daily coupled), associated with the westerly wind anomalies (CP12HC and CP1DC; Figure 4). Our analysis indicated that the occurrences of high RH (>70%) were significantly reduced in the low-frequency coupled experiments, especially at the 500-hPa level (middle troposphere), suggesting that deep convection was suppressed. Such a reduction occurred only after the MJO entered its active phase (Figure 6). Similar results were found in the uncoupled (atmosphere-only) model (NOCPC). The analysis of the apparent heat source (Q1) and moisture sink (Q2) budget confirmed that the vertical advection of heat and moisture played the dominant role during the active phase of the MJO, but both were weakened in the low-frequency coupled and uncoupled experiments.

According to our results, high-frequency air-sea coupling is necessary for representing the diurnal cycle and the daily mean of SST. Specifically, in 30-min and 1-hr coupled experiments (CP30MC and CP1HC), the SST successfully reproduced the diurnal cycle of SST, and the maximum SST appeared at 8:00 UTC, which was consistent with observations (Ruppert & Johnson, 2015). However, in CP3HC and CP6HC, the time of the diurnal maximum SST was delayed by 1 and 4 hr, respectively, and such phase lags were also found in surface turbulent heat fluxes (TFLX, Figure 7). The surface air temperature (T2m) and specific humidity (q2m) basically followed the same trend of SST in the high-frequency coupled experiments (with few hours
Acknowledgments
The authors greatly appreciate the fruitful comments and suggestions by three anonymous reviewers. The authors are grateful to all those who engaged in the YMC-Sumatra 2017 field campaign, and the observational data set is available (at www.jamstec.go.jp/jmc/jpn/jmcj_data.html). This study is supported by authors’ institute (JAMSTEC) as regular annual budget provided by the Ministry of Education, Culture, Sports, Science and Technology (MEXT) of Japan. Tomoe Nasuno is also supported by JSPS KAKENH Grant Number JP19H04248. We also thank the following institutions/agencies for data provision: Japan Aerospace Exploration Agency for the GSMaP data set (https://sharaku.eorc.jaxa.jp/GSMaP/), the National Centers for Environmental Prediction for the NCEP-FNL data set (https://rda.ucar.edu/datasets/ds083.2/), the National Centers for Environmental Information for the OISST data set (https://www.ncdc.noaa.gov/oisst), the Hybrid Coordinate Ocean Model (HYCOM) Consortium for Data-Assimilative Ocean Modeling for the GOF03.1 data set (https://www.bycom.org/datasetserver/gofs-3pt1/analysis), the U.S. Geological Survey for the GMED2010 data set (https://www.usgs.gov/land-resources/eros/coastal-changes-and-impacts/gmed2010), the U.S. National Centers for Environmental Information for ETOPO1 data set (https://www.ngdc.noaa.gov/mgg/global/), and the Bureau of Meteorology, Australia, for the RMM MJO index (https://www.bom.gov.au/climate/mjo/). The numerical experiments in this study were carried out on the JAMSTEC Super Computer System (DA System).

References
Amante, C., & Eakins, B. W. (2009). ETOPO1 1 arc-minute global relief model: Procedures, data sources, and analysis. Maryland, United States: National Geophysical Data Center, NOAA. Accessed on 13 May, 2019.
Bae, S. Y., Hong, S.-Y., & Tiao, W. K. (2018). Development of a single-moment cloud microphysics scheme with prognostic hail for the Weather Research and Forecasting (WRF) model. Asia-Pacific Journal of Atmospheric Sciences, 55, 233–245.
Bananzon, V., Smith, T. M., Chin, T. M., Liu, C., & Hankins, W. (2016). A long-term record of blended satellite and in situ sea-surface temperature for climate monitoring, modeling and environmental studies. Earth System Science Data, 8(1), 165–176.
Bernie, D. J., Gualdier, E., Madec, G., Slingo, J. M., Woolnough, S. J., & Cole, J. (2008). Impact of resolving the diurnal cycle in an ocean-atmosphere GCM. Part 2: A diurnally coupled CGCM. Climate Dynamics, 31, 909–925.
Bernie, D. J., Woolnough, S. J., Slingo, J. M., & Gualdier, E. (2005). Modeling diurnal and intraseasonal variability of the ocean mixed layer. Journal of Climate, 18, 1190–1202.
Chen, X., & Zhang, F. (2019). Relative roles of pre-conditioning moistening and global circumnavigating mode on the MJO convective initiation during DYNAMO. Geophysical Research Letters, 46, 1079–1087. https://doi.org/10.1029/2018GL080987
Clayson, C. A., & Bogdanoff, A. S. (2013). The effect of diurnal sea surface temperature warming on climatological air-sea fluxes. Journal of Climate, 26(8), 2546–2556.
Crueger, T., Stevens, B., & Brokopf, R. (2013). The Madden-Julian oscillation in ECHAM6 and the introduction of an objective MJO metric. Journal of Climate, 26(10), 3241–3257.
Cummings, J. A. (2005). Operational multivariate ocean data assimilation. Quarterly Journal of the Royal Meteorological Society, 131(613), 3583–3604.
Danielson, J. J., & Gesch, D. B. (2011). Global multi-resolution terrain elevation data 2010 (GTOPO2010), Virginia, United States: U.S. Department of the Interior and U.S. Geological Survey. Accessed on 13 May, 2019.
DeMott, C. A., Klingman, N. P., Tseng, W.-L., Burt, M. A., Gao, Y., & Randall, D. A. (2019). The convection connection: How ocean feedbacks affect tropical mean moisture and MJO propagation. Journal of Geophysical Research: Atmospheres, 124, 11,910–11,931. https://doi.org/10.1029/2019JD031015
DeMott, C. A., Klingman, N. P., & Woolnough, S. J. (2015). Atmosphere-ocean coupled processes in the Madden-Julian oscillation. Reviews of Geophysics, 53, 1099–1154. https://doi.org/10.1002/2014RG000478
Dipankar, A., Webster, S., Huang, X.-Y., & Doan, V. Q. (2019). Understanding biases in simulating the diurnal cycle of convection over the western coast of Sumatra: Comparison with pre-YMC observation campaign. Monthly Weather Review, 147, 1615–1631.
Fu, X., Lee, J.-Y., Hsu, P.-C., Taniguchi, H., Wang, B., Wang, W., & Weaver, S. (2013). Multi-model MJO forecasting during DYNAMO/CINDY period. Climate Dynamics, 41, 1067–1081.
Gerard, L. (2007). An integrated package for subgrid convection, clouds and precipitation compatible with meso-gamma scales. Quarterly Journal of the Royal Meteorological Society, 133(624), 711–730.
Green, B. W., Sun, S., Bleck, R., Benjamin, S. G., & Grell, G. A. (2017). Evaluation of MJO predictive skill in multiphysics and multimodel global ensembles. Monthly Weather Review, 145(7), 2555–2574.
Grell, G. A., & Freitas, S. R. (2014). A scale and aerosol aware stochastic convective parameterization for weather and air quality modeling. Atmospheric Chemistry and Physics, 14(10), 5233–5250.
Hagos, S., Leung, L. R., & Dudhia, J. (2011). Thermodynamics of the Madden–Julian oscillation in a regional model with constrained moisture. Journal of the Atmospheric Sciences, 68(9), 1974–1989.
Hagos, S. M., Zhang, C., Feng, Z., Burleson, C. D., DeMott, C., Kerns, B., et al. (2016). The impact of the diurnal cycle on the propagation of Madden-Julian Oscillation convection across the Maritime Continent. Journal of Advances in Modeling Earth Systems, 8, 1552–1564. https://doi.org/10.1002/2016MS000725
Holloway, C. E., Woolnough, S. J., & Lister, GrenvilleMc. S. (2015). The effects of explicit versus parameterized convection on the MJO in a large-domain high-resolution tropical case study. Part II: Processes leading to differences in MJO development. Journal of the Atmospheric Sciences, 72(7), 2719–2743.
Hong, S.-Y., & Jiang, J. (2018). Impacts of shallow convection processes on a simulated boreal summer climatology in a global atmospheric model. Asia-Pacific Journal of Atmospheric Sciences, 54(5), 361–370.
Hong, S.-Y., Noh, Y., & Dudhia, J. (2006). A new vertical diffusion package with an explicit treatment of entrainment processes. Monthly Weather Review, 134(9), 2318–2341.
Iacono, M. J., Delamere, J. S., Mlawer, E. J., Shephard, M. W., Clough, S. A., & Collins, W. D. (2008). Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. Journal of Geophysical Research, 113, D13103. https://doi.org/10.1029/2008JD009944
JAMSTEC, & BPPT (2018). Study on air-sea interaction over upwelling region in the eastern Indian Ocean: Japan Agency for Marine-Earth Science and Technology, Japan and Agency for the Assessmentand Application of Technology, Indonesia.
Jiménez, P. A., Dudhia, J., González-Rouco, J. F., Navarro, J., Montávez, J. P., & García-Bustamante, E. (2012). A revised scheme for the WRF surface layer formulation. Monthly Weather Review, 140(3), 898–918.
Katsumata, M., Morii, S., Hamada, J.-I., Hattori, M., Syamsuddin, F., & Yamanaka, M. D. (2018). Diurnal cycle over a coastal area of the Maritime Continent as derived by special networked soundings over Jakarta during HARIMAUI2010. Progress in Earth and Planetary Science, 5, 64.
Kayano, M., T., & Kousky, V. E. (1999). Intraseasonal (30–60 day) variability in the global tropics: Principal modes and their evolution. Tellus A: Dynamic Meteorology and Oceanography, 51(3), 373–386.
Kessler, W. S. (2001). EOF representations of the Madden–Julian oscillation and its connection with ENSO. Journal of Climate, 14(13), 3055–3061.
Kim, H. M., Hoyos, C. D., Webster, P. J., & Kang, I.-S. (2010). Ocean-atmosphere coupling and the boreal winter MJO. Climate Dynamics, 35, 771–784.
Kim, H., Vitart, F., & Waliser, D. E. (2018). Prediction of the Madden–Julian oscillation: A review. Journal of Climate, 31(23), 9425–9443.
Kwon, Y. C., & Hong, S.-Y. (2017). A mass-flux cumulus parameterization scheme across gray-zone resolutions. Monthly Weather Review, 145(2), 583–598.
Lorenz, D. J., & Hartmann, D. L. (2006). The effect of the MJO on the North American monsoon. Journal of Climate, 19(3), 333–343.
Madden, R. A., & Julian, P. R. (1972). Description of global-scale circulation cells in the tropics with a 40–50 day period. Journal of the Atmospheric Sciences, 29(6), 1109–1123.
Madden, R. A., & Julian, P. R. (2005). Historical perspective. In W. K. M. Lau, & D. E. Waliser (Eds.), Atmospheric Sciences: An Assessment of the State of the Science (Vol. 29, pp. 1–16). Berlin: Springer-Praxis.
Mishra, S. K., Sahany, S., & Salunke, P. (2017). Linkages between MJO and summer monsoon rainfall over India and surrounding region. Meteorology and Atmospheric Physics, 129(6), 283–296.
Morak-Bozzo, S., Merchant, C. J., Kent, E. C., Berry, D. J., & Carella, G. (2016). Climatological diurnal variability in sea surface temperature characterized from drifting buoy data. Geoscience Data Journal, 3(1), 20–28.
Mori, S., Jun-Ichi, H., Taulid, Y. I., Yamanaka, M. D., Okamoto, N., Murata, F., et al. (2004). Diurnal land sea surface peak migration over Sumatera Island, Indonesian Maritime Continent, observed by TRMM satellite and intensive rawinsonde soundings. Monthly Weather Review, 132(8), 2021–2039.
Moteki, Q., Katsumata, M., Yoneyama, K., Ando, K., & Hasegawa, T. (2018). Drastic thickening of the barrier layer off the western coast of Sumatra due to the Madden-Julian oscillation passage during the Pre-Years of the Maritime Continent campaign. Progress in Earth and Planetary Science, 5, 35.
NCEP (2000). NCEP FNL Operational Model Global Tropospheric Analyses, continuing from July 1999. Research Data Archive at the National Center for Atmospheric Research, Computational and Information Systems Laboratory, Boulder CO, https://doi.org/10.5065/D6M043C6, Accessed on 24 May, 2019.
Nasuno, T. (2019). Moisture transport over the Maritime Continent during the 2015 and 2017 YMC Sumatra campaigns in global cloud-system-resolving simulations. SOLA, 15, 99–106.
Nasuno, T., Li, T., & Kikuchi, K. (2015). Moistening processes before the convective initiation of Madden-Julian oscillation events during the CINDY2011/DYNAMO period. Monthly Weather Review, 143(2), 622–643.
Neale, R., & Slingo, J. (2003). The Maritime Continent and its role in the global climate: A GCM study. Journal of Climate, 16(5), 834–848.
Petts, A. J., Matthews, M. J., & Stevens, B. (2014). Propagation of the Madden–Julian Oscillation through the Maritime Continent and scale interaction with the diurnal cycle of precipitation. Quarterly Journal of the Royal Meteorological Society, 140(680), 814–825.
Rao, R., Zhang, C., & Dudhia, J. (2016). Roles of deep and shallow convection and microphysics in the MJO simulated by the model for prediction across scales. Journal of Geophysical Research: Atmospheres, 121, 10,575–10,600. https://doi.org/10.1002/2015JD024697
Raymond, D. J., & Fuchs, Z. (2009). Moisture modes and the Madden–Julian oscillation. Journal of Climate, 22(11), 3031–3046.
Reynolds, R. W., Smith, T. M., Liu, C., Chelton, D. B., Casey, K. S., & Schlax, M. G. (2007). Daily high-resolution-blended analyses for sea surface temperature. Journal of Climate, 20(22), 5473–5496.
Ruppert, J. H., & Johnson, R. H. (2015). Diurnally modulated cumulus moistening in the preonset stage of the Madden–Julian oscillation during DYNAMO. Journal of the Atmospheric Sciences, 72(4), 1622–1647.
Seo, H., Subramanian, A. C., Miller, A. I., & Cavanagh, N. R. (2014). Coupled impacts of the diurnal cycle of sea surface temperature on the Madden–Julian oscillation. Journal of Climate, 27(22), 8422–8443.
Shinoda, T. (2005). Impact of the diurnal cycle of solar radiation on intraseasonal SST variability in the western equatorial Pacific. Journal of Climate, 18(14), 2628–2636.
Shinoda, T., & Ueda, H. (2002). Effective factors in the development of deep convective clouds over the wet region of eastern China during the summer monsoon season. Journal of the Meteorological Society of Japan. Ser. II, 80(6), 1395–1414.
Souza, J. M., Powell, B. Castillo-Trujillo, A. C. & Flament, P. (2015). The vorticity balance of the ocean surface in Hawaii from a regional reanalysis. Journal of Physical Oceanography, 45, 424–440. https://doi.org/10.1175/JPO-D-14-0074.1
Tewari, M., Chen, F., Wang, W., Dudhia, J., LeMone, M. A., Mitchell, K., et al. (2004). Implementation and verification of the unified Noah land surface model in the WRF model. Seattle, WA, US. Presentation.
Tseng, K.-C., Sui, C.-H., & Li, T. (2015). Moistening processes for Madden–Julian oscillations during DYNAMO/CINDY. Journal of Climate, 28(8), 3041–3057.
Vitart, F. (2014). Evolution of ECMWF sub-seasonal forecast skill scores. *Quarterly Journal of the Royal Meteorological Society, 140*(683), 1889–1899.

Waliser, D. E., Lau, K. M., & Kim, J.-H. (1999). The influence of coupled sea surface temperatures on the Madden-Julian oscillation: A model perturbation experiment. *Journal of the Atmospheric Sciences, 56*(3), 333–358.

Wang, B., & Moon, J.-Y. (2018). Subseasonal prediction of extreme weather events. In H.-S. Jung, & B. Wang (Eds.), *Bridging science and policy implication for managing climate extremes* (pp. 33–48). Singapore: World Scientific.

Warner, J. C., Armstrong, B., He, R., & Zambon, J. B. (2010). Development of a coupled ocean–atmosphere–wave–sediment transport (COAWST) modeling system. *Ocean Modelling, 35*(3), 230–244.

Webber, B. G. M., Matthews, A. J., & Heywood, K. J. (2010). A dynamical ocean feedback mechanism for the Madden–Julian oscillation. *Quarterly Journal of the Royal Meteorological Society, 136*(684), 740–754.

Wu, P., Ardiansyah, D., Mori, S., & Yoneyama, K. (2019). The effect of an active phase of the Madden-Julian oscillation on surface winds over the western coast of Sumatra Island. In *IOP Conference Series: Earth and Environmental Science, 301*, pp. 012,009.

Yanai, M., Esbensen, S., & Chu, J.-h. (1973). Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *Journal of the Atmospheric Sciences, 30*(4), 611–627.

Yokoi, S., Mori, S., Katsumata, M., Geng, B., Yasunaga, K., Syamsudin, F., et al. (2017). Diurnal cycle of precipitation observed in the western coastal area of Sumatra Island: Offshore preconditioning by gravity waves. *Monthly Weather Review, 145*(9), 3745–3761.

Yokoi, S., Mori, S., Syamsudin, F., Haryoko, U., & Geng, B. (2019). Environmental conditions for nighttime offshore migration of precipitation area as revealed by in situ observation off Sumatra Island. *Monthly Weather Review, 147*(9), 3391–3407.

Zermeño, D. M., Zhang, C., Kollias, P., & Kalesse, H. (2015). The role of shallow cloud moistening in MJO and non-MJO convective events over the ARM Manus site. *Journal of the Atmospheric Sciences, 72*(12), 4797–4820.

Zhang, C. (2013). Madden–Julian oscillation: Bridging weather and climate. *Bulletin of the American Meteorological Society, 94*(12), 1849–1870.

Zhang, C., & Anderson, S. P. (2003). Sensitivity of intraseasonal perturbations in SST to the structure of the MJO. *Journal of the Atmospheric Sciences, 60*(17), 2196–2207.

Zhang, C., & Wang, Y. (2017). Projected future changes of tropical cyclone activity over the western North and South Pacific in a 20-km-mesh regional climate model. *Journal of Climate, 30*(15), 5923–5941.

Zheng, Y., Alapaty, K., Herwehe, J. A., Del Genio, A. D., & Niyogi, D. (2016). Improving high-resolution weather forecasts using the Weather Research and Forecasting (WRF) model with an updated Kain–Fritsch scheme. *Monthly Weather Review, 144*(3), 833–860.

Zhu, J., Wang, W., & Kumar, A. (2017). Simulations of MJO propagation across the Maritime Continent: Impacts of SST feedback. *Journal of Climate, 30*(5), 1689–1704.