Diurnal Cycle of Precipitation Observed in the Western Coastal Area of Sumatra Island: Offshore Preconditioning by Gravity Waves

SATORU YOKOI, SHUICHI MORI, MASAKI KATSUMATA, AND BIAO GENG

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

KAZUAKI YASUNAGA

Department of Earth Science, University of Toyama, Toyama, and Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan

FADLI SYAMSUDIN

Agency for the Assessment and Application of Technology, South Tangerang, Indonesia

NURHAYATI

Agency for Meteorology, Climatology and Geophysics, Jakarta, Indonesia

KUNIO YONEYAMA

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

(Manuscript received 9 December 2016, in final form 14 June 2017)

ABSTRACT

This study analyzes data obtained by intensive observation during a pilot field campaign of the Years of the Maritime Continent Project (Pre-YMC) to investigate the diurnal cycle of precipitation in the western coastal area of Sumatra Island. The diurnal cycle during the campaign period (November–December 2015) is found to have a number of similarities with statistical behavior of the diurnal cycle as revealed by previous studies, such as afternoon precipitation over land, nighttime offshore migration of the precipitation zone, and dependency on Madden–Julian oscillation (MJO) phase. Composite analyses of radiosonde soundings from the Research Vessel (R/V) Mirai, deployed about 50 km off the coast, demonstrate that the lower free troposphere starts cooling in late afternoon (a couple of hours earlier than the cooling in the boundary layer), making the lower troposphere more unstable just before precipitation starts to increase. As the nighttime offshore precipitation tends to be more vigorous on days when the cooling in the lower free troposphere is larger, it is possible that the destabilization due to the cooling contributes to the offshore migration of the precipitation zone via enhancement of convective activity. Comparison of potential temperature and water vapor mixing ratio tendencies suggests that this cooling is substantially due to vertical advection by an ascent motion, which is possibly a component of shallow gravity waves. These results support the idea that gravity waves emanating from convective systems over land play a significant role in the offshore migration of the precipitation zone.

1. Introduction

While precipitation in the tropics varies across a wide range of time scales, the diurnal cycle is arguably the most dominant source of variability. The diurnal cycle is considered to modulate behavior of longer-scale phenomena such as intraseasonal variability (e.g., Hagos et al. 2016). It also has considerable impacts on Earth’s radiation budget, as clouds only in daytime have an opportunity of reflecting insolation. Therefore, understanding the whole picture of the diurnal cycle and the underlying physical processes is an important topic in climate science.

Previous studies have revealed that the diurnal cycle has different characteristics in different regions.
Over land, precipitation tends to have maxima in the late afternoon and early evening, whereas precipitation over open ocean tends to be maximal in the early morning. On the other hand, in coastal areas, where tropical precipitation tends to concentrate on average (Xie et al. 2006; Ogino et al. 2016), the diurnal cycle is known to have much more complicated characteristics; diurnal precipitation peaks tend to migrate offshore during nighttime. Mori et al. (2004) analyzed satellite-borne precipitation radar data to describe the statistical characteristics of the diurnal cycle over and around Sumatra Island, which is located in the westernmost region of the Indonesian Maritime Continent (see Fig. 1). They showed that, in the western coastal area of the island, a precipitation peak tended to emerge near the coast in early evening and then migrate offshore during nighttime to reach at least 400 km off the coast the following morning. This feature was confirmed by a number of papers, such as Sakurai et al. (2005, 2009), Wu et al. (2009), Mori et al. (2011), Fujita et al. (2011), and Kamimera et al. (2012). Similar offshore migration was also observed in other tropical coastal areas, including those of Borneo Island (Houze et al. 1981; Ichikawa and Yasunari 2006), New Guinea Island (Liberti et al. 2001; Zhou and Wang 2006; Ichikawa and Yasunari 2008), Panama, and Colombia (Mapes et al. 2003a).

Compared with the diurnal cycle over land, which can be considered a direct response to the diurnal cycle of insolation, mechanisms responsible for the nighttime offshore migration seem to be more elusive; thus, they have received a lot of attention from researchers. Several mechanisms have been proposed, most of which focused on how the offshore atmospheric condition became favorable for convection during nighttime. Warner et al. (2003) and Mapes et al. (2003b) proposed that an ascent motion in the lower troposphere, which was due to gravity waves emanating from the nighttime radiative cooling of the elevated terrain of the Andes, destabilized the offshore atmosphere west of the Pacific coast of Panama and Columbia. Love et al. (2011) and Hassim et al. (2016) suggested the role of gravity waves emanating from convective systems over land. While diabatic heating within convective systems can generate gravity waves with various vertical wavelengths, evaporative cooling may also generate gravity waves that have an ascent signal in the lower troposphere, which propagates offshore and destabilizes the atmosphere. In contrast to these studies focusing on gravity waves, others have proposed that gravity currents diurnally excited over land play a role in the offshore near-surface horizontal convergence, which provides favorable conditions for convection (Houze et al. 1981; Mori et al. 2004; Wu et al. 2009; Fujita et al. 2010; Wapler and Lane 2012). In particular, the convergence is thought to be caused by land breezes and environmental monsoonal winds (Houze et al. 1981), cold outflow from convection over land and environmental winds (Mori et al. 2004; Wu et al. 2009), opposing systems of land-originated cold outflow (Fujita et al. 2010), and opposing land-breeze systems (Wapler and Lane 2012). The advection of convective systems over land by environmental winds is another candidate for the mechanism. Mori et al. (2004) speculated that thick anvil clouds expanding from
convection over land were driven offshore by environmental wind in the middle and upper troposphere, from which precipitation activated offshore convection through the seeder–feeder effect.

These mechanisms were proposed via analyses of observations by satellites, ground-based weather radars, and radiosonde soundings from land, together with numerical simulation outputs. On the other hand, in order to evaluate the validity of these mechanisms, it seems necessary to compare modeling results with observations in terms of thermodynamic conditions over coastal waters. However, direct observations of thermodynamic variables there have been lacking, as there is no routine radiosonde observation sites over the sea. With one of the aims being to obtain such observations, we conducted an intensive observation campaign in the western coastal area of Sumatra Island in November–December of 2015. The name of the campaign was Pre-YMC, as this was a pilot study of the Years of the Maritime Continent (YMC) Project, a 2-yr (July 2017–July 2019) international project consisting of multiple intensive observation projects in meteorology and oceanography focusing on the Maritime Continent\(^1\); the Pre-YMC is expected to provide scientific basis for the planning of the main phase of the YMC. As a part of the Pre-YMC, we performed 3-hourly radiosonde observations on the Research Vessel (R/V) *Mirai*, deployed about 50 km off the coast, as well as at a coastal city near the vessel. We will examine the thermodynamic conditions through analysis of the observational data obtained during the Pre-YMC to verify the proposed mechanisms for the diurnal offshore migration of the precipitation zone; this is the main purpose of the present study.

One of the scientific goals of the YMC is to understand multiscale interactions between the diurnal cycle and the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972) over the Maritime Continent. The MJO is the dominant mode of atmospheric variability in the tropics on intraseasonal time scales, and it is characterized by a convective envelope with horizontal scales of a few thousand kilometers and associated circulation field anomalies. In general, the convective envelope migrates eastward from the central Indian Ocean and across the Maritime Continent to the central Pacific at approximately 5 m s\(^{-1}\). The manner in which the characteristics of the diurnal cycle depend on the MJO is known to be different in different regions. In the western coastal area of Sumatra Island in particular, the diurnal cycle of precipitation tends to be more distinct when the MJO convective envelope is located over the eastern Indian Ocean, whereas it becomes obscured when the envelope arrives at the Maritime Continent (Fujita et al. 2011; Kamimera et al. 2012; Vincent and Lane 2017). On the other hand, the diurnal cycle off the north coast of New Guinea Island is more active when the MJO convective envelope is over the Maritime Continent and western Pacific (Ichikawa and Yasunari 2008; Peatman et al. 2014). As will be explained in section 2, an MJO convective envelope arrived at the Maritime Continent in the latter half of the Pre-YMC intensive observation period, providing an opportunity to examine this dependency. This is another purpose of the present study.

The remainder of the paper is organized as follows. Section 2 gives an overview of the Pre-YMC field campaign and MJO conditions during the intensive observation period. In section 3, we explain the observational data analyzed in this study. Section 4 presents the results and discussion. Analyses of the data related to precipitation and cloud are presented in section 4a, while those of thermodynamic profiles are in section 4b. In sections 4c and 4d, we examine the thermal stability of the offshore atmosphere and attempt to estimate the vertical velocity in late afternoon and early evening. We also discuss which of the proposed mechanisms is able to explain the observed results. Finally, a summary of this paper is presented in section 5.

2. Overview of the Pre-YMC campaign

The Pre-YMC intensive observation campaign was conducted in the western coastal area of Sumatra Island (Fig. 1) in November–December of 2015 by the Japan Agency for Marine–Earth Science and Technology (JAMSTEC), the Indonesian Agency for the Assessment and Application of Technology (BPPT), and the Indonesian Agency for Meteorology, Climatology and Geophysics (BMKG). To observe the offshore migration of the precipitation zone, we set up two observation sites located on coastal water and coastal land on a line nearly perpendicular to the coastline that runs roughly linearly from northwest to southeast. For the coastal water site, the R/V *Mirai* of JAMSTEC was deployed about 50 km off the coast at around 4.07°S, 101.90°E (as indicated by the closed circle in Fig. 1b) over the period from 23 November to 16 December 2015. We used the BMKG observatory in Bengkulu city (3.86°S, 102.34°E, 16 m above sea level, as indicated by the square in Fig. 1b), located just a few kilometers inland from the coast, as the coastal land site to perform observations over the period from 9 November to 25 December 2015.

\(^1\) 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.
To give an overview of the large-scale atmospheric conditions in the tropics during the observation periods, Fig. 2 presents the Hovmöller diagram of several variables averaged between 10°S and the equator. Figure 2a plots precipitation of the Global Satellite Mapping of Precipitation (GSMaP; Okamoto et al. 2005) near-real-time product. Until early December, precipitation occurred mostly over the eastern Indian Ocean and the central Pacific around the date line, with relatively less precipitation over the Maritime Continent (100°–150°E). The area with a large amount of precipitation over the Indian Ocean then started to migrate eastward, as indicated by the dashed line, and passed over the study area in mid-December, whereas precipitation around the date line appeared to be suppressed. In late December, the eastward-migrating area reached the date line, again producing high precipitation rates in this region. The eastward migration speed can be estimated at 5–7 m s\(^{-1}\), which is close to the speed of typical MJO.

The Hovmöller diagram of 850-hPa zonal wind of the Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) is shown in Fig. 2b. In November and early December, easterly wind dominated over the Maritime Continent. Over the central Indian Ocean, westerly wind displaced easterly wind in the second half of November and intensified gradually until early December. Then, the westerly wind area expanded eastward as far as the date line in mid-December. Comparison with Fig. 2a reveals that while the westerly wind followed the eastward-migrating precipitation area over the Indian Ocean and western part of the Maritime Continent, the former seemed to catch up with the latter over the western Pacific. This contrast between the Indian Ocean and the western Pacific in the positional relationship between precipitation and zonal wind is a well-known feature of the MJO (Hendon and Salby 1994). Figure 2c plots the anomalies of column-integrated water vapor from a two-month average of JRA-55. While westward-migrating signals with relatively small zonal scales dominated in November, a positive anomaly with a zonal scale comparable to that of the MJO was found in December, migrating eastward along with the precipitation area, which is also consistent with the characteristics of typical MJO.

Based on the above analyses, we can consider this eastward disturbance to be an MJO event. To confirm this, Fig. 3 plots a phase diagram of the Real-time Multivariate MJO (RMM) index designed by Wheeler and Hendon (2004). Whereas the RMM index rotated counterclockwise from phase 3 to 4 in early November, its amplitude decreased and its phase progression was less obvious after 10 November. The amplitude increased again in early December, and then the index resumed the counterclockwise rotation in mid-December. The obvious rotating behavior can be observed at least until the end of the month when the index reached phase 7, which corresponds to the eastward disturbance shown by the dashed line in Fig. 2. Therefore, it can be said that the Pre-YMC field campaign successfully observed the passage of the convective envelope of the MJO event over the study area.
3. Data and methods

Among various kinds of observation we performed at the two sites, this study analyzes data of weather radar, radiosonde sounding, rain gauge, and ceilometer.

Weather radars for monitoring precipitation particles were installed at both sites. Over the vessel, we utilized a dual-polarized C-band Doppler radar to obtain volume scan data at 6-min intervals with a ray and gate spacing of $0.78^-18$–150 m, respectively, at 10 elevation angles ranging from 0.5$^\circ$ to 40$^\circ$, with a 100-km range. At the Bengkulu observatory, BMKG routinely operates a C-band Doppler radar to obtain volume scan data at 10-min intervals with a ray and gate spacing of 1$^\circ$ and 250 m, respectively, at 10 elevation angles ranging from 0.5$^\circ$ to 40$^\circ$, with a 120-km range. The observation ranges are indicated by circles in Fig. 1b. Both radars covered the coastal waters as well as coastal land. Note, however, that the BMKG radar is unable to observe precipitation beyond the ridge of the mountain range, which runs parallel to the coast at about 50-km distance, due to interception of the radar beam by the topography. Furthermore, whereas the vessel’s radar was in operation during the entire observation period (23 November–16 December), the BMKG radar was out of service until 11 December; thus, its data during 12–25 December are used for analysis.

We estimate precipitation intensity from radar reflectivity at 2-km altitude above sea level, which is obtained by interpolation of volume scan, with the use of a simple reflectivity–rainfall relation:

$$Z = \alpha R^\beta,$$

where $Z$ indicates reflectivity (in mm$^6$ m$^{-1}$) and $R$ indicates precipitation (in mm h$^{-1}$). The parameters $\alpha$ and $\beta$ for the vessel’s radar are estimated through comparison between the reflectivity obtained by the radar and rain gauge data at Bengkulu for the 23 November–14 December period, following a method described by Yokoi et al. (2012). The estimated values of $\alpha$ and $\beta$ are 216 and 1.28, respectively. Through comparison of data of the two radars, the parameters $\alpha$ and $\beta$ for the BMKG radar are then estimated at 157 and 1.28, respectively. Although we acknowledge that this estimation method is simplistic, we believe that it is sufficient for the present purpose of examining spatiotemporal precipitation variability.

Because of the quasi-linear configuration of the coastline and mountain range, previous studies (e.g., Mori et al. 2004; Kamimera et al. 2012) assumed that the characteristics of the diurnal cycle did not vary considerably for the direction parallel to the coastline. The present study will adopt this approach and average the estimated precipitation data in the northwest–southeast direction to examine precipitation as a function of distance from the large-scale coastline indicated by a dashed line in Fig. 1b. Hourly precipitation data are then obtained by averaging the 6- or 10-min data. Note that we have compared the precipitation estimated from the BMKG radar with that from the vessel’s radar for the overlapping period (12–16 December) and confirmed that these showed remarkably similar time series (figure not shown).

The radiosonde observations were performed at 3-h intervals [0100, 0400, ..., and 2200 local time (LT); LT = UTC + 7 h] at both sites with the use of RS92-SGPD sensors manufactured by Vaisala Ltd. We examine potential temperature, water vapor mixing ratio, and horizontal wind data with 10-hPa vertical resolution. A potential concern is that, as the radiosondes sometimes shift horizontally for several tens of kilometers by the time they reach the tropopause, we might care about their position when examining phenomena with horizontal scales comparable to the distance of the shift. However, we will focus on the lower troposphere, where the distance of the shift is expected to be small. In fact, for the radiosonde observation from the vessel, mean horizontal distance of the radiosondes from the vessel is about 2 km at the 900-hPa level, 3 km at the 800-hPa level, and 6 km at the 500-hPa level, which is considerably smaller than the distance between the vessel and Bengkulu. Therefore, we can conclude that the shift has only a small impact on our arguments.

We also analyze rain gauge precipitation for 1-h resolution at both sites. Additionally, cloud-base height

---

**Fig. 3.** The RMM index phase diagram for November (blue) and December (red) of 2015. Numbers over the trajectory indicate day of the months for every 5 days.
data obtained by the ceilometer (CL51 manufactured by Vaisala Ltd.) on board the vessel at a sampling interval of 36 s are used to calculate the occurrence frequency of the cloud base for 1-h temporal and 100-m altitude resolutions, as well as to determine the cloud-free frequency for the same temporal resolution.

4. Results and discussion

a. Precipitation and cloud

To understand the behavior of the diurnal cycle during the intensive observation period, we first examine the time series of radar-estimated precipitation as a function of distance from the coastline (Fig. 4). Here distance is defined with the direction toward land being positive, and the vessel’s average position and the coastline are indicated by vertical dashed lines. Horizontal dashed lines indicate 0700 LT (0000 UTC) on each day. It is clearly observed that the diurnal cycle bore different characteristics between the periods before and after 13 December. Precipitation time series in the former period showed a more regular diurnal cycle than in the latter period. Early afternoon precipitation maxima over land and offshore migration of the precipitation zone during nighttime were evident during the former period (except for 26–27 November and 8 December). Such a diurnal cycle was obscured in the latter period; instead, the precipitation zone tended to move onshore with shorter time scales. Furthermore, there was almost no rainfall on 13, 17, and 24 December.

Since the characteristics of the diurnal cycle appear to be different between the two periods, it seems natural to investigate the mean diurnal cycle (MDC) for these two periods separately. The MDCs of precipitation for the periods 23 November–12 December (period I) and

![Fig. 4. Time series of radar-estimated precipitation (mm h⁻¹) as a function of distance from the coastline. The distance is defined as the direction toward land being positive. Data plotted were obtained by the vessel’s radar for 23 Nov–16 Dec and by the BMKG radar for 17–24 Dec. Vertical dashed lines represent the positions of the vessel (left) and the coast (right), and horizontal dashed lines indicate 0700 LT on each day. Gray shading indicates missing data (3, 7, and 12 Dec) or mountain shadow (17–24 Dec).](image-url)
14–24 December (period II) are shown in Figs. 5a and 5b, respectively. Here, the MDC is calculated by making composites of the data at each hour of the day during periods I and II. The MDC for period I indicates that precipitation starts to increase over land around noon and reaches a maximum at 1500 LT at about 30 km inland. The precipitation zone then migrates in the offshore direction and crosses the coastline at 1800 LT. It migrates farther offshore, passes over the vessel at around 2100–2200 LT, and eventually reaches the boundary of the study area (140 km from the coastline) at around 0100 LT. Precipitation activity seems to intensify near the boundary after gradual weakening at 50–100 km from the coast. The migration speed of the front of the zone (identified as an edge of the blue shadings) over the sea, and that of the precipitation maximum over the sea farther than 40 km from the coast, can be estimated at about 8 m s$^{-1}$ (indicated by the dot–dashed line). This speed is a little bit slower but broadly consistent with a result of Mori et al. (2004), which estimated the speed averaged over the sea within 400 km from the coast at $\sim 11$ m s$^{-1}$. On the other hand, the precipitation maximum over the sea farther than 40 km from the coast, can be estimated at about 8 m s$^{-1}$ (indicated by the dot–dashed line). This speed is a little bit slower but broadly consistent with a result of Mori et al. (2004), which estimated the speed averaged over the sea within 400 km from the coast at $\sim 11$ m s$^{-1}$ (note that their Fig. 5a suggests that the speed increases after the precipitation zone reaches about 150 km from the coast). On the other hand, the precipitation maximum over the sea farther than 40 km from the coast, can be estimated at about 8 m s$^{-1}$ (indicated by the dot–dashed line). This speed is also consistent with a result of Mori et al. (2011), which estimated the migration speed over the sea within 80 km from the coast at $\sim 4$ m s$^{-1}$. Note that the contrast in the migration speed between the front and precipitation maximum, and the contrast between near-coast and far-offshore areas, were also pointed out by Hassim et al. (2016) and Vincent and Lane (2016), respectively, both mainly through numerical simulation focusing on the northeastern coastal area of New Guinea Island.

Figure 6 plots the MDC of the rain gauge data at both sites, as well as that of radar-estimated precipitation over Bengkulu. At Bengkulu (Fig. 6a), the gauge precipitation starts to increase at 1400 LT and reaches its maximum at 1800 LT, followed by a gradual decrease until 0200 LT. Such MDC is consistent with the radar-estimated precipitation. The gauge precipitation at the vessel (Fig. 6b) also exhibits the MDC that is found to be consistent with the radar-estimated precipitation shown in Fig. 5a, with a preference of the large precipitation rate during nighttime and a maximum at 2200 LT.

Figure 7 shows the MDC of the occurrence frequency of the cloud base over the vessel for period I, which also exhibits characteristics consistent with the radar-estimated precipitation. In the morning and early afternoon, the cloud-free frequency exceeds 30%, with a maximum of nearly 60% at 1300 LT. The cloud base is mostly confined at around 0.5-km altitude. At around 1600 LT, the cloud-base frequency above 5-km altitude starts to increase. Based on our experience at the vessel during the campaign, these clouds were perhaps thick anvil clouds moving from the direction of land. Then the altitude of the midtropospheric maximum frequency slightly descends with time from about 5.5 km at 1900 LT to 5 km at 0200 LT. In the lower troposphere, the cloud-base
frequency increases at altitudes between 0.5 and 2 km in late afternoon and night, which was probably related to active convective systems. The cloud-free frequency at these times is less than 10%.

During period II, on the other hand, the MDC of radar-estimated precipitation (Fig. 5b) is much less clear, as seen in Fig. 4. The propagation signal seems to be reversed in direction. Furthermore, the time of day showing large offshore precipitation is completely different; precipitation tends to be greater during daytime than during nighttime, which is rather similar to the MDC over open ocean.

To argue what causes the change in the behavior of the diurnal cycle on 13 December, Fig. 8 shows time series of zonal wind obtained by radionsonde soundings at Bengkulu. Strong westerly wind in the lower troposphere below the 300-hPa level seemed to take place on 12 or 13 December and lasted at least until 22 December, which was arguably related to the MJO event. As the offshore migration is considered to be obscured when lower-tropospheric wind blows toward land (e.g., Ichikawa and Yasunari 2008), the

Fig. 6. MDC of rain gauge precipitation at (a) Bengkulu and (b) the vessel for period I. Dots in (a) indicates MDC of radar-estimated precipitation over Bengkulu.

Fig. 7. MDC of (a) occurrence frequency of the cloud base [% (100 m$^{-1}$)] and (b) occurrence frequency of a cloud-free condition (%) observed by the ceilometer on board the vessel for period I.

Fig. 8. Time–vertical cross section of zonal wind observed at Bengkulu. Running diurnal average has been applied.
change in direction of lower-tropospheric zonal wind seemed to play a role in the change in the behavior of the diurnal cycle of precipitation.

Overall, the observed characteristics of the diurnal cycle of precipitation, including mean features during period I and modulation due to the MJO, are considerably similar to the statistical characteristics of the diurnal cycle in the western coastal area of Sumatra Island as revealed by previous studies introduced in section 1. These similarities encourage us to analyze observational data further to discuss physical processes for the nighttime offshore migration of the precipitation zone during period I.

b. MDC of thermodynamic profile

One of the unique features of the Pre-YMC campaign is the radiosonde observation over coastal waters as well as over coastal land. Analyses of the radiosonde data are expected to provide information helpful for examining the proposed mechanisms for the offshore migration of the precipitation zone. To focus on the diurnal cycle, we analyze anomalies from the diurnal running average, which is defined as a 9-point average of 3-hourly time series, with the weights for the first and last points being half of the others. We also examine differences in the data between the two sites, which are expected to make clear land–sea contrasts. The MDC of the anomalies and the differences for period I are obtained by making composites at each of 3-hourly time steps of the day during the period. The statistical significance of the composite anomalies and differences are then assessed by the Student’s t test with an assumption that each of the 20 days of period I is independent of the others.

As the intensity of convective activity is deeply influenced by the atmospheric thermodynamic profile before convection takes place, here we examine the MDC of the potential temperature and the water vapor mixing ratio. Figures 9a and 9b show the MDC of the potential temperature anomaly over Bengkulu and the vessel, respectively. In general, the composite anomalies below the 750-hPa level are positive during daytime and negative during nighttime, which is consistent with a diurnal cycle of insolation. At Bengkulu, positive temporal maxima are observed at 1300 LT from the surface through the 750-hPa level, whereas negative maxima are observed at 0400 LT below the 950-hPa level and at 0700 LT above this level. The amplitudes near the surface are much larger than that at the other pressure levels. At the vessel, the daytime positive anomaly seems to lag 3–6 h behind that at Bengkulu, and a phase difference appears to exist between the pressure levels below and above the 950-hPa level. Positive maxima in the 950–750-hPa layer are observed at 1600 LT, whereas those below the 950-hPa level are observed at 1900 LT. Figure 10a shows the MDC of potential temperature anomaly averaged over the 1000–950-hPa (black) and 900–800-hPa (red) layers and confirms the vertical phase difference. As a result, thermal instability in the lower troposphere is increased by 1900 LT when precipitation starts to increase (Figs. 5a and 6b). In fact, a difference of equivalent potential temperature averaged over the 1000–950-hPa layer from saturated equivalent potential temperature over the 900–800-hPa layer (Fig. 10b), which is a measure of the lower-tropospheric thermal instability, is increased greatly at 1600–1900 LT, with its sign changing from negative to positive. This means that the atmosphere becomes more favorable for convection by 1900 LT. This feature is an interesting phenomenon and will be discussed in the next section.

The MDC of potential temperature difference between the two sites (Fig. 9c) indicates that the boundary layer at Bengkulu is warmer than that at the vessel during daytime and cooler during nighttime. Such differences are arguably due to differences in heat capacity between land and sea. Whereas the daytime positive and significant difference extends almost to the 900-hPa level, the nighttime negative and significant difference is confined to the layer below the 950-hPa level. This contrast in depth is consistent with typical land–sea-breeze circulation (i.e., the sea breeze generally has a deeper vertical extent than the land breeze).

The MDC of the mixing ratio anomaly is presented in Figs. 9d and 9e. In general, the anomaly seems to synchronize with precipitation at both sites, particularly in the boundary layer. At Bengkulu, the anomalies below the 950-hPa level and around the 800-hPa level have their maxima at 1600 LT, when precipitation starts to increase, while those around the 900-hPa level record maxima at 1900 LT, when precipitation shows its maximum value (Fig. 6a). Similar characteristics are observed at the vessel. Below the 850-hPa level, positive maxima are observed at 2200 LT, when precipitation is greatest (Fig. 6b). The correspondence between the lower-tropospheric mixing ratio and precipitation suggests that the mixing ratio maxima may be at least partly due to evaporation of hydrometeors in the subcloud layer. Differences in mixing ratio between the two sites (Fig. 9f) indicate that the near-surface atmosphere below the 900-hPa level is more humid over the vessel than over Bengkulu, which is presumably due to the abundant moisture supply from the sea surface. On the other hand, above the 900-hPa level, the atmosphere over Bengkulu is more humid, especially in the afternoon, which may indicate that the convective mixed layer is deeper over the coastal land than it is over the coastal waters.
Fig. 9. (a), (b) Vertical profile of the MDC of potential temperature anomaly at (a) Bengkulu and (b) the vessel for period I, and (c) that of the difference in potential temperature between the two sites (Bengkulu minus the vessel). Contour intervals are 0.2 K and yellow (pale blue) shading indicates that the anomaly or difference is positive (negative) and statistically significant at the 95% confidence level. (d)–(f) As in (a)–(c), but for water vapor mixing ratio, with contour intervals of 0.2 g kg$^{-1}$. 
c. Late-afternoon cooling in the lower free troposphere

An interesting finding in the last section is the vertical phase difference in the MDC of potential temperature over the vessel. In particular, potential temperature below the 950-hPa level increases until 1900 LT, whereas potential temperature above this level starts to decrease at 1600 LT, which makes the lower troposphere become thermally more unstable by 1900 LT. To further make sure of this vertical difference, Fig. 11a shows the MDC of lower-tropospheric thermal instability defined as a difference of equivalent potential temperature averaged over the 1000–950-hPa layer from saturated equivalent potential temperature over the 900–800-hPa layer, at the vessel.

To discuss the relationship between the cooling and the diurnal cycle of precipitation, we examine the daily time series of the radar-estimated precipitation at a particular location and hour of day (Fig. 12b). The top of this figure plots the regression coefficients between the late-afternoon cooling and precipitation at 0700 LT of the same day of the cooling, while the bottom of the figure plots the coefficients between the cooling and precipitation at 0700 LT of the next day of the cooling. Negative coefficients (shown by reddish colors), which indicate more precipitation on days with stronger cooling, migrate offshore nearly along with the large precipitation zone found in the MDC plot (Fig. 5a). While close examination reveals that the negative coefficients lag about 1 h behind the composite precipitation in several locations, it seems inappropriate to discuss the meaning of such lags, since the temporal interval of the radiosonde observation is as long as 3 h. On the other hand, it is interesting to note that negative and statistically significant coefficients are observed over land a couple of hours prior to the cooling. The coefficients are also negatively large and statistically significant at 100–150 km offshore for several hours after the cooling, where offshore intensification takes place (Fig. 5a). These results suggest a close relationship between the late-afternoon cooling in the lower free troposphere and the diurnal cycle of precipitation. As the cooling precedes the precipitation, it is possible that the cooling contributes to the offshore migration of the precipitation zone via destabilization of the offshore atmosphere. Note that Birch et al. (2016) also demonstrated that the thermal instability played a significant role in active convection over the coastal waters in MJO phases 3 and 4. Negative and significant coefficients at 50–100 km offshore are also found at 0400–0700 LT during the next day, although at this moment we do not have any idea that physically explains this significant relationship, which may be a subject of our future research.

Next, we discuss the factors responsible for this cooling. In general, the local tendency of potential temperature below the 950-hPa level increases until 1900 LT, whereas potential temperature above this level starts to decrease at 1600 LT, which makes the lower troposphere become thermally more unstable by 1900 LT. To further make sure of this vertical difference, Fig. 11a shows the MDC of potential temperature tendency over the vessel. It is clearly seen that the tendency in the 900–800-hPa layer is negative and statistically significant at 1600–1900 LT, when the tendency below this layer is nearly zero and insignificant.

To discuss the relationship between the cooling and the diurnal cycle of precipitation, we examine the daily time series of the potential temperature tendency at 1600–1900 LT in the 900–800-hPa layer over the vessel (Fig. 12a). The cooling dominated only during period I, whereas warming took place in the following days (13 December onward) when the diurnal cycle of precipitation was obscured. Using this time series in period I as a reference, we then perform regression analysis of daily time series of the radar-estimated precipitation at a particular location and hour of day (Fig. 12b). The top of this figure plots the regression coefficients between the late-afternoon cooling and precipitation at 0700 LT of the same day of the cooling, while the bottom of the figure plots the coefficients between the cooling and precipitation at 0700 LT of the next day of the cooling. Negative coefficients (shown by reddish colors), which indicate more precipitation on days with stronger cooling, migrate offshore nearly along with the large precipitation zone found in the MDC plot (Fig. 5a). While close examination reveals that the negative coefficients lag about 1 h behind the composite precipitation in several locations, it seems inappropriate to discuss the meaning of such lags, since the temporal interval of the radiosonde observation is as long as 3 h. On the other hand, it is interesting to note that negative and statistically significant coefficients are observed over land a couple of hours prior to the cooling. The coefficients are also negatively large and statistically significant at 100–150 km offshore for several hours after the cooling, where offshore intensification takes place (Fig. 5a). These results suggest a close relationship between the late-afternoon cooling in the lower free troposphere and the diurnal cycle of precipitation. As the cooling precedes the precipitation, it is possible that the cooling contributes to the offshore migration of the precipitation zone via destabilization of the offshore atmosphere. Note that Birch et al. (2016) also demonstrated that the thermal instability played a significant role in active convection over the coastal waters in MJO phases 3 and 4. Negative and significant coefficients at 50–100 km offshore are also found at 0400–0700 LT during the next day, although at this moment we do not have any idea that physically explains this significant relationship, which may be a subject of our future research.

Next, we discuss the factors responsible for this cooling. In general, the local tendency of potential temperature is caused by horizontal and vertical advection and the diabatic processes of atmospheric radiation, latent heating due to phase transition of water molecules, and turbulent heat flux. The MDC of mixing ratio tendency (Fig. 11b) turns out to provide a clue to this issue. At 1600–1900 LT, a positive (moistening) tendency with statistical significance accompanies the cooling tendency. There are two possibilities regarding what process causes this pair of tendencies. The first one is vertical advection due to ascent motion, as potential temperature generally increases with height while mixing ratio generally decreases with height. The other one
is evaporation of liquid water, which also cools and humidifies the atmosphere. Since a ratio of potential temperature tendency to mixing ratio tendency induced by the two processes is generally different from each other, we can examine which process is more presumable from the viewpoint of this ratio of tendencies. If this pair of tendencies is caused by the vertical advection process, the ratio of tendencies should be equal to a ratio of vertical gradient of potential temperature to that of mixing ratio. The ratio of vertical gradients in the 900–800-hPa layer at 1600–1900 LT can be estimated at \(-2.1 \times 10^3\) K. On the other hand, if the pair of tendencies is caused by the evaporation process, the ratio of tendencies should be equal to \(-L/C_p\), where \(L\) is the specific heat of vaporization and \(C_p\) is the specific heat of dry air at constant pressure. Since the ratio of tendencies in the 900–800-hPa layer at 1600–1900 LT is \(-1.4 \times 10^3\) K, it can be said that the vertical advection is more likely a major cause of the pair of tendencies than the evaporation, although there still

![Fig. 11. Vertical profile of the MDC of the tendencies of (a) potential temperature and (b) water vapor mixing ratio at the vessel for period I. Contour intervals are (a) 0.2 K (3 h) \(^{-1}\) and (b) 0.2 g kg \(^{-1}\) (3 h) \(^{-1}\). Yellow (pale blue) shading indicates that the tendency is positive (negative) and statistically significant at the 95% confidence level.](image)

![Fig. 12. (a) Daily time series of potential temperature tendency at 1600–1900 LT in the 900–800-hPa layer over the vessel. (b) Regression coefficients (unit: mm h \(^{-1}\) [K (3 h) \(^{-1}\)] \(^{-1}\)) of daily time series of radar-estimated precipitation at a particular location and hour of day, against the time series shown in (a) for period I. Negative (positive) coefficients indicate more (less) precipitation on days with stronger cooling. Hatched areas are those where correlation coefficients are statistically significant at the 95% confidence level, as assessed by the Student’s \(t\) test. The location and time of the reference potential temperature tendency is shown by a black box labeled as C. In the ordinate, “07(0),” for example, indicates the precipitation at 0700 LT of the same day of the cooling, while “07(+1)” indicates the precipitation at 0700 LT of the next day.](image)
exists a difference between the ratio of tendencies and that of vertical gradients.

The above argument suggests that ascent motion may exist in the offshore lower free troposphere at 1600–1900 LT. Therefore, we experimentally estimate the vertical velocity profile from composite profiles of potential temperature at 1600 and 1900 LT, with an assumption that its tendency is solely due to the vertical advection. As potential temperature generally increases monotonically with height, for an air parcel at each pressure level at 1600 LT, we can find only one pressure level where the composite potential temperature at 1900 LT is the same as the air parcel. From the difference between this pressure level and the pressure level of the parcel at 1600 LT, we can estimate the vertical velocity at 1600–1900 LT, which is plotted with a solid line in Fig. 13a. Likewise, we can estimate vertical velocity with an assumption that mixing ratio tendency is solely due to vertical advection, which is also plotted in Fig. 13a with a dashed line. Although these assumptions seem to be quite simplistic, the estimated vertical velocity still has several characteristics worth noting. First, both are negative with comparable magnitude (0.1–0.2 Pa s$^{-1}$) in the 950–750-hPa layer, and have a negative maximum at around the 850-hPa level. The quantitative correspondence of the two estimations supports our hypothesis that the ascent motion exists in this layer, as argued in the last paragraph. Second, the depth of the ascent is somewhat shallow; thickness of the 950–750-hPa layer is only 2–2.5 km. Third, the ascent layer does not reach the surface. Both estimations below the 950-hPa level exhibit a descent motion with comparable magnitude (except in the layer near the surface), although corresponding warming and drying are not statistically significant (Fig. 11).

Using horizontal wind and potential temperature profiles at both sites, we also examine whether horizontal advection contributes to the cooling tendency. It turns out that the horizontal advection is about one order of magnitude smaller than the tendency (figure not shown) and thus does not seem to have significant impact.

The statistically significant relationship that a larger cooling rate in late afternoon is observed on days when early-afternoon precipitation over land is stronger (Fig. 12b) suggests that the cooling may be caused by the cumulus convection over land rather than, for example, sea-breeze circulation. Among the proposed processes for the offshore migration reviewed in section 1, the one that is able to explain the above observed results most consistently is convectively generated gravity waves. Hassim et al. (2016) simulated the offshore migration of the precipitation zone to the northeast of New Guinea Island and found that the evaporative cooling of the convective systems over land generated gravity waves with ascending wave fronts in the lower troposphere, which propagated offshore nearly horizontally, as predicted theoretically by Nicholls et al. (1991). If the estimated ascent motion (Fig. 13a) is a part of the gravity wave, the vertical wavelength $\lambda_z$ can be estimated from the depth of the ascent at as shallow as 4–5 km. Then the horizontal phase speed of the gravity wave $v_p = N\lambda_z/2\pi$, in which $N \approx 1.25 \times 10^{-2}$ s$^{-1}$ is the Brunt–Väisälä frequency, is calculated to be 8–10 m s$^{-1}$. Here we neglect

![Fig. 13. (a) Vertical pressure velocity over the vessel at 1600–1900 LT, as estimated from composite profiles at 1600 and 1900 LT of potential temperature (solid) and mixing ratio (dashed) of period I. See the text for details of the estimation method. (b) As in (a), but for 1900–2200 LT.](image-url)
background wind speed, as the horizontal wind component normal to the coastline averaged over period I and over the 950–750-hPa layer turns out to be less than 1 m s$^{-1}$. It should be noted that, if the gravity wave with these parameters emanates from the precipitation maximum at 1500 LT at 30 km inland and propagates offshore at such speed, it reaches the vessel’s position at around 1700–1730 LT, consistent with the timing of the late-afternoon cooling observed. This supports our hypothesis that the cooling and ascent over the vessel are caused by the shallow gravity waves emanating from the convection over land.

The gravity wave supposed here has characteristics similar to one of the gravity wave types simulated by Hassim et al. (2016). They argued that there were two gravity wave types in their simulations: one has an ascending wave front below 5-km altitude and propagates offshore at $\approx 15$ m s$^{-1}$, and the other has the wave front below 2–2.5 km altitude and propagates at $\approx 8$ m s$^{-1}$. These were classified into “slow” and “gust front” modes, respectively, identified by Tulich and Mapes (2008). Considering the depth of the ascent layer and the phase speed, the gravity wave argued here is similar to the latter mode rather than the former one.

Such gravity waves with ascending wave front in the lower troposphere are considered to contribute to the generation of new convective cells near the existing convective systems via destabilization of the lower atmosphere and near-surface convergence (e.g., Mapes 1993; Shige and Satomura 2000, 2001; Tulich and Mapes 2008). Love et al. (2011) and Hassim et al. (2016) reported that this process is at play in their realistic numerical simulations of the offshore migration. The observational evidence of the cooling and its tight relationship with offshore precipitation intensity found in the present study seems to support the idea that the gravity waves generated by evaporative cooling of the convective systems over land contribute to the nighttime offshore precipitation via destabilization of the lower troposphere. Note that the estimated horizontal phase speed of the gravity wave (8–10 m s$^{-1}$) is consistent with the migration speed of the precipitation zone offshore.

Mapes et al. (2003b) also demonstrated the importance of gravity waves with similar vertical structures in the destabilization of the lower troposphere. However, the wave they simulated was generated by nighttime radiative cooling of elevated terrain; thus, it propagated offshore during midnight and predawn hours. As the cooling found in the present study occurs in the late afternoon, it should not be caused by gravity waves generated in this way. There is another possibility that the cold outflow that emanates from convective systems causes convergence at their fronts and thus the ascent motion. However, the arrival of the cold outflow also induces surface temperature drop and wind speed increase, which are not observed in the radiosonde data (Fig. 9b) and surface meteorology data (not shown) at 1600–1900 LT. In addition, the convergence due to the cold outflow should occur near the surface, which is inconsistent with the vertical structure of the estimated vertical velocity, which has a node at the 950-Pa level.

d. Evening cooling in the boundary layer

In addition to the late-afternoon cooling in the lower free troposphere, Fig. 11a also indicates that, at 1900–2200 and 2200–0100 LT, there exist negative and significant potential temperature tendencies between the surface and the 880-hPa level, with a negative maximum at the surface. Since the cooling at 1900–2200 LT is accompanied by positive mixing ratio tendencies with statistical significance (Fig. 11b), there is a possibility that this cooling is also substantially due to ascent motion. Vertical velocity profiles at 1900–2200 LT estimated from potential temperature profiles and from mixing ratio profiles, shown in Fig. 13b, are consistent with each other below the 880-hPa level, where an ascent motion with magnitudes of 0.1–0.15 Pa s$^{-1}$ is suggested. Therefore, it can be said that the boundary layer cooling at 1900–2200 LT is also likely to be due mainly to vertical advection, except for at the surface layer where the ascent estimated from the potential temperature profiles is much larger than that from the mixing ratio profiles and diabatic processes may be at work.

Evidence for the boundary layer ascent can also be observed in the difference in southwest–northeast component of horizontal wind $u_{n}$ at Bengkulu from that at the vessel (Fig. 14). Here, the positive value of $u_{n}$ is defined as a southwesterly (onshore) wind. At 1900 LT, negative and significant differences are found near the surface, with positive and significant differences at around the 850-hPa level. When we make the assumption that the characteristics of the MDC of the circulation field do not change in the direction parallel to the coastline (i.e., that there is no horizontal gradient in this direction), we can estimate the horizontal divergence between the two sites from the difference in $u_{n}$, with positive (negative) values corresponding to horizontal divergence (convergence). Given the distance between the two sites, the surface convergence at 1900 LT can be estimated as $5–6 \times 10^{-5}$ s$^{-1}$ whereas the divergence at the 850-hPa level can be estimated as $2–3 \times 10^{-5}$ s$^{-1}$. A pair of convergence and divergence implies an ascent motion in between, whose maximum is estimated as about 0.2 Pa s$^{-1}$ around the 900-hPa level.

The difference in $u_{n}$ suggests the ascent at 1900 LT between Bengkulu and the vessel, while potential
temperature and mixing ratio tendencies suggest the ascent at 1900–2200 LT over the vessel. Combining these results suggests a possibility that the ascent signal may propagate from the direction of land. As the temporal and spatial interval of the radiosonde observations are as long as 3 h and \( \sim 50 \text{ km} \), respectively, it is difficult to estimate the propagation speed. Nevertheless, as a quite rough estimate, we assume that the ascent at 1900 LT occurs just in the middle of the two sites and the ascent over the vessel occurs at 2030 LT, to obtain the propagation speed of about 4 m s\(^{-1}\). This speed is close to the speed of the precipitation maximum over the sea nearer than 40 km from the coast. As explained earlier, the offshore-migrating precipitation maximum has characteristics similar to a squall line. If this is the case, the above argument suggests a possibility that this ascent might correspond to the leading edge of the squall line, where surface convergence of convective outflow and the environmental air mass generates new convection. However, it is difficult to discuss this issue further, because of long temporal and spatial intervals of the radiosonde observation.

On the other hand, the boundary layer cooling at 2200–0100 LT is accompanied by a negative tendency of mixing ratio; therefore, neither the vertical advection nor evaporation can explain these tendencies. Furthermore, daytime warming and nighttime cooling tendencies near the surface are not accompanied by a significant mixing ratio tendency. These potential temperature tendencies are probably caused by turbulent heat flux and radiation.

5. Summary

The diurnal cycle of precipitation in tropical coastal areas bears much more complicated characteristics than that over open ocean and inlands; thus, it has attracted the interest of researchers. In this study, we have revealed the characteristics of the diurnal cycle in the western coastal area of Sumatra Island in November–December of 2015, during which we conducted the Pre-YMC field campaign. As part of this campaign, 3-hourly radiosonde soundings, continuous weather radar observations, and so forth, were performed at the R/V \textit{Mirai} of JAMSTEC, which was deployed about 50 km off the coast, and at the BMKG Bengkulu observatory which is located just a few kilometers inland from the coast. These observations were analyzed in this paper.

While an MJO convective envelope was over the Indian Ocean in the first half of the intensive observation period, it arrived at the study area around 13 December and passed over the Maritime Continent. The mean diurnal cycle (MDC) of precipitation also exhibited completely different characteristics for the 23 November–12 December period (period I) and the 14–24 December period (period II). During period I, the MDC is characterized by afternoon precipitation maxima over coastal land and the heavy precipitation zone migrates offshore during nighttime. The migration speed of the front of the zone over the sea can be estimated at about 8 m s\(^{-1}\), while that of the precipitation maximum around the coast is at 3–3.5 m s\(^{-1}\). The zone passes over the vessel at 2100–2200 LT and migrates farther offshore to at least 140 km from the coast. On the other hand, the MDC is obscured during period II, whereas onshore migration of the precipitation zone with shorter time scales is observed. These characteristics are consistent with the findings of previous statistical studies of observations such as Mori et al. (2004, 2011), Fujita et al. (2011), and Kamimera et al. (2012). Based on these results, it can be concluded that the diurnal cycle observed by the Pre-YMC field campaign is not peculiar; rather, it bears characteristics similar to the typical diurnal cycle in this area.

To better understand the mechanisms responsible for the nighttime offshore migration of the precipitation zone, we examined the MDC of thermodynamic profiles over the vessel as well as over Bengkulu for period I. While the MDC of potential temperature over Bengkulu is vertically in-phase in the lower troposphere during daytime, the potential temperature over the vessel is not vertically in-phase. Over the vessel, potential temperature in the lower free troposphere tends
to start cooling in late afternoon, about 3 h earlier than that in the boundary layer. This results in destabilization of the lower troposphere by the time that precipitation starts to increase. The intensity of nighttime offshore precipitation tends to be greater on days when the cooling is stronger, suggesting that the cooling and resultant destabilization is likely to enhance the convective activity of the migrating precipitation zone.

Comparison between the tendencies of potential temperature and water vapor mixing ratio suggests that this late-afternoon cooling in the lower free troposphere is due substantially to the vertical advection by ascent motion, with a magnitude of 0.1–0.2 Pas$^{-1}$. It is reasonable to speculate that this ascent motion is due to a gravity wave with a vertical wavelength of 4–5 km and horizontal phase speed of 8–10 m s$^{-1}$ that is generated by the evaporative cooling of convective systems over land. Furthermore, this gravity wave has similarities with a type of gravity wave known as the “gust front” mode determined by Tulich and Mapes (2008).

In addition, at 1900–2200 LT, just before the arrival of the precipitation maximum at the vessel’s position, cooling is also observed in the boundary layer and is due substantially to the vertical advection by ascent motion. Horizontal winds also imply an ascent motion in the boundary layer between the two sites at 1900 LT. If the precipitation zone has characteristics similar to a squall line, this ascent might correspond to the leading edge of the squall line.

The findings of the present study provide clues for understanding physical processes for the nighttime offshore migration of the precipitation zone. In particular, they support the idea that gravity waves emanating from the convective systems over land in early afternoon play a significant role in the offshore migration of the precipitation zone by making the offshore atmosphere more conductive to convection, as proposed by the numerical studies of Love et al. (2011) and Hassim et al. (2016). Furthermore, these results may be helpful for deepening the understanding of ensemble of cumulus convective systems and obtaining successful simulation of their behavior. As gravity waves are one of the phenomena through which the convective systems interact with each other, cloud resolving models should represent them accurately. The present results are expected to be used for evaluating the performance of the models in simulating such gravity waves.

Of course, there are a number of issues and subjects that should be addressed in our future research. For example, while this study investigates the diurnal cycle for the period when the MJO is inactive or the MJO convective envelope is over the Indian Ocean, how the convective envelope interacts with the diurnal cycle when the former arrives at the Maritime Continent (viz., period II) is a very interesting subject. Another subject that has not been addressed as much in this paper is the offshore intensification; to what extent such intensification can be observed and what physical processes cause the intensification are among some interesting questions. Some of them can be addressed by further analyses of the observational data obtained during the Pre-YMC, while others need another intensive observation campaign with slightly different configuration, such as deploying the vessel several tens of kilometers farther offshore to focus on the intensification process. It seems that the YMC is expected to be a nice opportunity to perform such a new campaign.

Acknowledgments. The authors are grateful to all those who were engaged in the Pre-YMC field campaign. The authors also thank Japan Aerospace Exploration Agency (JAXA) for the provision of the GSMaP dataset, the Japan Meteorological Agency (JMA) for the provision of JRA-55 dataset, and the Bureau of Meteorology, Australia, for the provision of the RMM index. This study is partly supported by Grant-in-Aid for Scientific Research (JP16K05560 and JP16H04048) of the Japan Society for the Promotion of Science, and the Environment Research and Technology Development Fund (2-1503) of Environmental Restoration and Conservation Agency, Japan.

REFERENCES


Ichikawa, H., and T. Yasunari, 2006: Time–space characteristics of diurnal rainfall over Borneo and surrounding oceans as

Unauthenticated | Downloaded 07/28/22 01:53 AM UTC