Detecting and Tracking Coastal Precipitation in the Tropics: Methods and Insights into Multiscale Variability of Tropical Precipitation

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ABSTRACT

We propose an algorithm to detect and track coastal precipitation systems and we apply it to 18 years of the high-resolution (8 km and 30 min) Climate Prediction Center CMORPH precipitation estimates in the tropics. Coastal precipitation in the Maritime Continent and Central America contributes to up to 80% of the total rainfall. It also contributes strongly to the diurnal cycle over land with the largest contribution from systems lasting between 6 and 12 h and contributions from longer-lived systems peaking later in the day. While the diurnal cycle of coastal precipitation is more intense over land in the summer hemisphere, its timing is independent of seasons over both land and ocean because the relative contributions from systems of different lifespans are insensitive to the seasonal cycle. We investigate the hypothesis that coastal precipitation is enhanced prior to the arrival of the Madden–Julian oscillation (MJO) envelope over the Maritime Continent. Our results support this hypothesis and show that, when considering only coastal precipitation, the diurnal cycle appears reinforced even earlier over islands than previously reported. We discuss the respective roles of coastal and large-scale precipitation in the propagation of the MJO over the Maritime Continent. We also document a shift in diurnal cycle with the phases of the MJO, which results from changes in the relative contributions of short-lived versus long-lived coastal systems.

1. Introduction

As an increasing number of people live within 100 km from the coast (Nicholls and Cazenave 2010; Small and Nicholls 2003; Klein et al. 2003), especially in the tropics, understanding the processes that control coastal climates becomes crucial for public safety and well-being. Precipitation on coasts can result from the advection over the coast of a long-lasting, large-scale storm or from local processes due to the adjacency of land and ocean. It is the latter that we will call coastal precipitation. How coastal precipitation influences the long-term climate

and interacts with lower-frequency variability is still an open question. Documenting in detail the diurnal cycle and intraseasonal variability of coastal precipitation is a first step in solving this question.

Tropical coastal precipitation overwhelmingly results from the diurnal cycle of clouds and winds at the coast. This diurnal cycle is characterized by a concentration of clouds and precipitation over land during daytime, with a peak in the early evening, and over sea at night, with a maximum in early morning (e.g., Yang and Slingo 2001; Nesbitt and Zipser 2003; Mori et al. 2004). It results from the moist-air convergence over the coastal land associated with the sea breeze circulation, which in turn is due to the strong temperature contrast between land and ocean generated by the solar diurnal forcing and the contrast between the low heat capacity of land and the much larger heat capacity of the ocean surface

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(Mori et al. 2004; Qian 2008). The complex topography of the Maritime Continent (MC) reinforces these circulations over the larger islands (Yang and Slingo 2001), as shown by local studies over the main islands of the archipelago: Sumatra (Mori et al. 2004; Yokoi et al. 2017), Borneo (Ichikawa and Yasunari 2007; Qian et al. 2013), and New Guinea (Ichikawa and Yasunari 2008; Vincent and Lane 2016; Hassim et al. 2016). At night, the diurnal sea breeze reverses into a land breeze. This land breeze and other processes such as gravity waves triggered by convection over the island are responsible for the offshore propagation of land convection over the surrounding ocean at night (Saito et al. 2001; Mori et al. 2004; Qian 2008; Love et al. 2011; Hassim et al. 2016; Vincent and Lane 2016; Li et al. 2017; Yanase et al. 2017; Coppin and Bellon 2019a,b).

Another focus of interest has been the interaction between the islands of the MC and the Madden–Julian oscillation (MJO) (Madden and Julian 1971, 1972, 1994; Zhang 2005). As the large-scale envelope of convective clouds associated with the MJO moves eastward from the Indian to the Pacific Ocean, it encounters the MC, which acts as a natural barrier for the eastward propagation of the MJO (Rui and Wang 1990; Hendon and Salby 1994). A detailed understanding of the interaction between the processes controlling the MJO and the land–sea breezes as it passes over the MC is necessary to model and predict the propagation of the MJO accurately. Failure to do so may explain the poor MJO forecast capability (Seo et al. 2009; Vitart and Molteni 2010) and a too strong barrier effect of the Maritime Continent in some models (Kim et al. 2009; Seo et al. 2009; Jiang et al. 2015).

The diurnal cycle of convection over islands has been mentioned as one of the factors responsible for the barrier effect of the MC. Hagos et al. (2016) show that enhanced convection over the islands competes with and disrupts the convective signal of MJO events that propagates over the waters surrounding the islands. The diurnal cycle also varies depending on the phase of the MJO. It is enhanced over the islands and suppressed over the surrounding seas prior to the arrival of the MJO envelope (Peatman et al. 2014, 2015; Moron et al. 2015). This early rainfall peak over islands is caused by strong sea breezes resulting from high surface insolation and heating of islands in clear-sky conditions and favored by a strong large-scale moisture flux convergence (Birch et al. 2016). Conversely, precipitation decreases over islands before it does over surrounding seas in the later phases of the MJO because the increased cloud cover due to the MJO envelope decreases the surface insolation over land and the resulting land–sea breeze circulation.

Coastal and diurnal processes also affect the long-term climatology. In particular, observations show that precipitation peaks at the coastline and is enhanced within 300 km from each side of the coastline (Ogino et al. 2016). Observations also show an enhancement of the diurnal cycle and rainfall over islands relative to the surrounding oceans (Sobel et al. 2011). This enhancement is particularly strong over large islands (a few hundreds of km² or greater) where it mainly results from a larger rain frequency and from mechanically forced upslope flow in mountainous regions (Sobel et al. 2011). Idealized experiments with an island surrounded by ocean suggest that the precipitation rate over the island can be more than double the domain average, with an intense and regular thunderstorm over the island in late afternoon to early evening (Cronin et al. 2015). These idealized experiments have confirmed that the size of the island affects the magnitude of the rainfall enhancement, with a maximum for islands of around 20 km in radius for which the atmospheric moisture convergence is maximized (Cronin et al. 2015). This is particularly true when the prevailing large-scale wind is small, in which case the sea breeze dominates and controls the moisture convergence (Wang and Sobel 2017). Island precipitation enhancement is much reduced when the wind speed increases. In this case, mechanically forced convection is favored and horizontal advection of oceanic temperature reduces the land–sea thermal contrast that drives the sea breeze. It rains all day long on the windward side of the mountain, leading to a local precipitation maximum. On the other hand, the leeward side experiences a stronger diurnal cycle of precipitation (Wang and Sobel 2017).

One approach that has proven very helpful to improve our knowledge of both intraseasonal tropical variability and the life cycle of clouds is the tracking of convective systems. Using satellite datasets, it is indeed possible to follow convective systems very precisely in space and time and study their life cycle with great precision. Originally designed to track mesoscale convective systems over specific regions for short campaigns (Aspliden et al. 1976; Martin and Schreiner 1981; Williams and Houze 1987), tracking has been progressively applied to larger areas (Mapes and Houze 1993; Boer and Ramanathan 1997; Futyan and Del Genio 2007; Fiolleau and Roca 2013a; Hagos et al. 2013) up to the whole tropical belt (Fiolleau and Roca 2013b). It has considerably improved our understanding of the factors controlling the life cycle of mesoscale convective systems and of their interaction with their environment. All these studies use the background state of cloud field and most of the time only consider cloud systems bigger than 5000 km² for the mesoscale convective system over the

FIG. 1. (top) (left) Mean precipitation and (middle) DJF and (right) JJA precipitation between 30°S and 30°N for the 18 years of our coastal precipitation dataset. (bottom) Fraction of coastal precipitation relative to CMORPH precipitation over the same period. The gray areas correspond to areas not considered by the tracking algorithm (farther than 400 km from the nearest coast).

tropical belt (Fioleau and Roca 2013b) or bigger than 400 km² for smaller regions (Feng et al. 2012; Hagos et al. 2013). Furthermore, they usually only focus on a few consecutive months of data. As far as we know, only Bergemann et al. (2015) focus on the coastal precipitation in the tropics and do not filter out small convective systems. But their algorithm, which specifically targets the coastal systems parallel to the coasts every 3 h at 0.25° of spatial resolution for several years, does not consider the small circular convective systems because of its criteria on eccentricity and alignment relative to the coast. It also detects but does not track the systems during their lifetime and thus cannot be used to look at the propagation of the convective systems and their life cycle.

The aim of this study is to describe and evaluate a new method that detects and tracks coastal precipitating systems in the tropics, and to evaluate the climatology and variability of these systems. Section 2 describes the data and tracking method. Section 3 describes the main characteristics of the convective systems tracked over the tropics and section 4 further documents the role of coastal precipitating clouds and their diurnal cycle in the propagation of the MJO.

2. Methods

a. Precipitation data

We use 18 years (1998–2015) of satellite rainfall estimates CMORPH-CRT from the Climate Prediction Center (CPC) morphing technique (CMORPH; Joyce

et al. 2004) for the tropical region (30°S–30°N; Fig. 1a) with a spatial resolution of 8 km and a temporal resolution of 30 min. CMORPH uses motion vectors derived from geostationary satellite infrared imagery (sampled at 30-min intervals, the same frequency as the microwave retrievals) to produce high-quality precipitation estimates from passive microwave data. The 30-min analyses at 8-km resolution depend on microwave retrievals exclusively, and only use infrared data to propagate microwave-derived precipitation features between microwave measurements. Time-weighted linear interpolation is used to modify the shape and intensity of precipitation features for time steps between passive microwave sensor scans to ensure precipitation estimates are temporally and spatially complete. CMORPH retrieves passive microwave precipitation estimates from the NOAA polar-orbiting operational meteorological satellites (including NOAA-15, -16, and -18); DMSP-13, -14, and -15; and NASA's Aqua and TRMM spacecraft. Infrared images over different time intervals are provided by the GOES-8, GOES-10, Meteosat-5, Meteosat-7, and Geostationary Meteorological Satellite 5 (GMS-5) satellites. The corrected dataset (CMORPH-CRT) differs from the satellite-only dataset (CMORPH-RAW) over land, where the probability density function of CPC unified daily rain gauge analysis (Xie and Xiong 2011) is used to calibrate the satellite estimates. Compared to CMORPH-RAW, CMORPH-CRT better estimates the amplitude of the diurnal cycle (Rauniyar et al. 2017).

The dataset shows substantial improvement over methods that blend infrared and microwave data, as

well as methods that average passive microwave precipitation datasets (Janowiak et al. 2005; Rauniyar et al. 2017). Even though it is one of the best satellite-based rainfall estimates available, it cannot capture precipitation forming or evaporating over regions not covered by passive microwave sensors. It also lacks accuracy for rainfall over ice or snow cover (Joyce et al. 2004), which does not affect our study. Like all the other microwave-dominated products, CMORPH-CRT generally overestimates the frequency of days with no rain because it fails to accurately detect light rainfall from shallow and warm clouds (Ebert et al. 2007; Guo et al. 2015). It also tends to overestimate extreme rainfall over islands (Rauniyar et al. 2017).

b. Tracking algorithm

1) GENERAL PRINCIPLE

For the rest of the paper, we use “cluster” to mean a contiguous precipitation area isolated by the watershed technique [see section 2b(2) below]. The pixels composing the clusters are called cells. A track refers to an ensemble of clusters that interact at some point during their lifetime. Individual clusters inside a track can appear, split, merge, and die (includes moving out of the zone of study) while the track continues.

We choose ellipses to fit the convective clusters because their geometric properties (orientation, position and length of the main axes) are easily retrieved. But even with this simple formulation, fitting ellipses to convective clusters is the most computationally costly section of our algorithm due to the large numbers of clusters at each time step. To reduce this computing cost, we exploit the sparsity of the CMORPH data (large regions without precipitation) by representing clusters as a set of cells rather than using image masking as in Bergemann et al. (2015).

Once the cells of a cluster are identified, their moment of inertia formula can be applied to compute the best-fitting ellipse without the need to recover the boundary as in the image-masking method. The latter operation scales with the number of cells in the cluster, but not with the number of grid points in the dataset, which gives this method a more favorable scaling, particularly for high-resolution precipitation data. Equations (1)–(3) show how the moment method can be used to compute the size (or total mass) of the cluster (zeroth-order moment), the center of gravity (first-order moment), and the characteristics of a cluster’s fitting ellipse (second-order moment):

$$\sum_i m_i = \text{total mass}, \quad (1)$$

$$\begin{aligned} \sum_i m_i \times (x_i, y_i) &= \text{total mass} \times \text{position} \\ &= \text{center of gravity}, \end{aligned} \quad (2)$$

$$\sum_i m_i \begin{pmatrix} x_i^2 & x_i y_i \\ x_i y_i & y_i^2 \end{pmatrix} = \text{total mass} \times \text{inertia tensor}, \quad (3)$$

with m_i being the “weight” of each cell, and x_i and y_i the longitude and latitude at the cell location. In our case we use $m_i = 1$; that is, all the cells have the same weight.

Note that the inertia tensor is symmetric; hence the eigenvalues are real numbers and the eigenvectors must be perpendicular to each other. The eigenvalues of the matrix in Eq. (3) give the length of the two perpendicular axes of the ellipse while the eigenvectors point in the direction of the axes. The orientation of the cluster is defined as the angle between the long axis and the north direction.

We prescribe a minimum length for both axes of the ellipse in order to properly track very small cloud systems. This is required because our tracking algorithm relies on the overlap of a cluster’s fitting ellipses at two consecutive time steps, and in the case of small clusters advected by strong winds, these two ellipses, as defined by Eqs. (2) and (3) above, might not overlap. The minimum axis-length value is chosen to ensure that clusters of a few grid points moving at 30 km h^{-1} are tracked properly (min_axis in Table 1).

2) DESCRIPTION OF THE ALGORITHM

In this description of the algorithm, old clusters/tracks refer to clusters/tracks at the previous time step (t_0), while new clusters/tracks are related to the present time step (t_1). The parameters are specified in Table 1. The main steps of the algorithm are as follows:

- Create the masks used to filter the coastal precipitation (Fig. 2). The larger mask covers the islands and all areas within 400 km from a coast. The smaller one, used to select the tracks that interact with islands and coasts, covers a band of 48 km on both sides of the coastline. The final output is fairly insensitive to the choice of the large mask. It is slightly more sensitive to the distance chosen for the coastal mask. We choose a width of 48 km for the latter as we want this coastal band to be as narrow as possible to ensure that the convective systems result from coastal processes, but we also want to avoid filtering out convective systems that develop from undetectable, small or shallow coastal clouds. For example, if shallow convection develops at the coast while being advected by the wind, it can have moved a few tens of kilometers before it transforms into deep convection that will be clearly visible in the precipitation

TABLE 1. Details of the parameters. Parameters in bold have a significant impact on the output.

Parameter	Default	Description
min_prec	0	Low threshold for watershed on precipitation data (in mm h^{-1})
max_prec	3.0	High threshold for watershed on precipitation data (in mm h^{-1})
szone	6	Distance to coast in pixels used to create mask of islands and close surrounding areas (land and ocean)
lzone	50	Distance to coast in pixels used to create mask of islands and large areas of surrounding ocean
frac_mask	1.0	Threshold to keep tracks once clusters overlap with masks (minimum overlap necessary)
frac_ellipse	1.0	Threshold to merge overlapping ellipses (minimum overlap necessary)
min_axis	6	Minimum ellipse axis length used for tracking (in pixels)
min_size	0	Area below which islands are deleted from the masks (in km^2)
max_size	800 000	Maximum area of islands (in km^2) that are filled by the land–sea mask. Above, for continents, only coast.
max_cells	4400	Minimum size of tracks (in pixels) that can be filtered out as large-scale systems (equivalent circle of 300 km of diameter)
t_life	5	Minimum lifetime for tracks to be filtered out as large-scale systems (in days)
t_life_min	2	Same as t_life, but for tracks that touch the northern or southern boundaries of the domain
frac_decrease	0.95	Decrease in size above which a track is cut (equivalent to 95%)

dataset. Also, 48 km is a reasonable size to detect coastal systems as it is well within the area influenced by land–sea breeze interactions (Coppin and Bellon 2019a,b) and other definitions of the coastal region can include up to 300 km on each side of the coastline (Ogino et al. 2016, 2017). Choosing this distance also allows us to detect small coastal systems developing over the ocean and propagating offshore such as the ones visible north of New Guinea in Hassim et al. (2016) and Vincent and Lane (2016). These systems are not detected if the width is decreased further.

- For each day of data and each time step, the algorithm:
 - 1) Detects all convective clusters by applying a watershed technique on the CMORPH data (Fig. 3a) using two different thresholds. The low threshold (0 mm h^{-1} by default to keep all precipitation) creates a contour around precipitating regions (Fig. 3b). The high threshold serves to detect convective cores around which we build the different

clusters. Several values were tested for the high threshold and 3 mm h^{-1} was chosen as it enables us to discriminate between the convective cores clearly without missing the early development of convective clusters. To understand how the watershed technique works, let us consider the topography map obtained by considering minus the precipitation as an altitude. Each convective core is a local minimum (the center of a basin) and regions without precipitation form a plateau at the top. All basins are filled up to a progressively increasing altitude, and all regions below the top plateau are progressively attributed to a basin by detecting the basin center connected to each grid point. Where two basins meet, dams are built all the way to the top plateau to keep these basins distinct. At the end of this procedure, each convective cluster (or basin) is composed of a convective core (basin center) surrounded by an area of weaker precipitation delimited by areas without precipitation

FIG. 2. Coastal masks used for the tracking. Only the precipitation inside the blue mask is considered at the beginning of the tracking. The smaller coastal region (yellow) is used to decide if a track is coastal at the end of the tracking. The land–sea mask used is created from the open water class of the Global 1-km Consensus Land Cover (Tuanmu and Jetz 2014) downgraded at the exact same resolution as the 8-km CMORPH data.

FIG. 3. (a) Snapshot of the CMORPH precipitation over Sumatra at 2000 local solar time 25 Mar 1999. (b) Precipitation masks for the same day and area. The convective cores are represented in yellow, the rest in blue. (c) Coastal precipitation as shown in the final output of the algorithm. (d) Tracks discriminated by the algorithm (also in the final output). One number represents one track.

and/or one or several interfaces (dams) with other clusters.

- 2) Discards the precipitation clusters located farther than 400 km away from any coast (i.e., if they are not completely included in the large mask) (blue mask in Fig. 2). Without this step, this algorithm can be used to track all precipitating systems, at a high computational cost. Here, precipitation away from coasts is not of interest.
- 3) Combines clusters whose ellipses completely overlap because they belong to the same system. This is generally the case when a strong convective core appears within a big convective envelope.
- 4) Tracks clusters between time steps (t_0 and t_1) using their ellipses (Fig. 4). Each cluster is followed in time as a group of cells to which a single identity label (ID) is associated. The complexity of the algorithm stems from the fact that we want to follow convective clusters during their whole lifetime with the same ID, even as they merge or break up.

- (i) Simple tracking: detect new clusters pertaining to existing tracks (all but gray and purple clusters in Fig. 4c). If the center of the ellipse of a new cluster (at t_1) is inside the ellipse of an old cluster (at t_0) or vice versa, the new cluster is added to the track of the old cluster. If the new cluster is inside several ellipses from different tracks, we choose the track of the nearest old cluster. As mentioned above, setting

a minimum ellipse size allows us to track small convective systems advected by the large-scale wind (e.g., red cluster in Fig. 4c).

- (ii) Tracking of splitting clusters: takes care of the splitting cases and ensures that we track as many clusters as possible, even very small. If several new clusters are inside the ellipse of an old cluster, both new clusters are included in the track of the old cluster (gray and purple clusters in Fig. 4c). A new ellipse is calculated (over purple clusters in Fig. 4d) for the next step (Fig. 4e) but the individual ellipses of each cluster will be used for the next “simple tracking” to avoid adding to the track new systems developing between split clusters. Additionally, the total area covered by all the clusters of a track is computed.
- (iii) Tracking of merging clusters: it is possible that two previously distinct clusters merge (green and blue clusters in Fig. 4e for example). In that case, we merge their tracks. To decide if two tracks merge, we build an ellipse for each old and new track (only one per track, and each track can be composed of several clusters). We call this ellipse the “track ellipse.” The merging occurs if the center of an old track ellipse is inside a new track ellipse (Fig. 4e). This ensures that two big convective tracks do not merge just because a few marginal clusters have overlapping ellipses. When two tracks merge, their

FIG. 4. Details of the different steps of the tracking algorithm. (a),(b) Precipitation clusters at t_0 and t_1 . (c) “Forward” tracking to check if new systems originate from an existing one, and take into account splitting cases (e.g., gray and purple systems). (d) New ellipses are calculated for the systems that have been merged. (e) “Backward” tracking to check if several previously distinct systems have merged (e.g., blue and green systems). (f) If so, both tracks are merged into one at t_0 and t_1 . (g)–(l) If several tracks are merged, we change the ID of the track for as long as it existed [green track becoming blue in (j) and (k)]. Note that for clarity purpose, the same clusters from the same track in (a) and (b) have the same ID, represented by the same color. But in reality, clusters in (a) and (b) have different colors until they are identified as belonging to the same track in (c).

total area becomes the sum of their individual areas (Fig. 4f). Track merging also implies merging these tracks for all previous time steps since their geneses (e.g., changing the color of the green track in Fig. 4g to blue at previous time steps, as in Fig. 4h).

(iv) Stops a track and starts a new one if the total area of a track decreases by more than a given fraction (frac_decrease in Table 1). This is to avoid continuing a track while most of its convective activity has died down but one marginal cluster remains active.

- At regular intervals, we archive information on all the tracks that are finished. To specifically target coastal convective systems, we only keep tracks that originate from the coastal region: more than 80% of the track area must overlap with the coastal mask (yellow mask in Fig. 2) at the time step this track is first detected. Any track which starts over the center of the big islands (orographic convection) or more than 48 km away from the coast, whether inland or offshore, is removed. A track is also declared large-scale and removed if its size ever becomes larger than the area of a circle of 300 km of diameter and if its lifetime exceeds 5 days (2 days if it touches the latitudinal border of the domain of study). The size and lifetime thresholds are designed to remove large mesoscale and synoptic systems that usually affect coastal regions, without removing coastal systems that stay in the same region for several days while being reinforced by daily land convection. The rest of the filtering also removes tropical storms, superclusters, mesoscale systems associated with equatorial gravity waves, and small oceanic systems advected over the coastal regions. The convective systems triggered by orography above central Borneo, Papua New Guinea, and Madagascar are also filtered out. Convective systems forced by orography near the coastline (i.e., southwest of Sumatra or the eastern part of Papua New Guinea) are still included in our analysis.
- These steps are repeated until the end of the period of study.

To reduce the computing time, the postprocessing (generating NetCDF files) is performed separately. The tracks are extracted from the files saved during the archive step of the main algorithm and each given a unique ID. Coastal precipitation is defined as precipitation associated with the tracks as illustrated in Figs. 3c and 3d. In this example, note how large-scale convection west and south of Sumatra is removed. Instances of tracks composed of independent clusters at that specific time because they previously split (dark blue track south of Sumatra) or adjacent clusters merging into one band (darker blue along the mountain range of Sumatra) are also visible.

3. Climatology of tropical coastal rainfall

The final output of the tracking algorithm consists of 18 years of tropical coastal precipitation systems at 30-min and 8-km resolution, with advanced information on the systems' life cycles. Here we present briefly the climatology, seasonal cycle, and diurnal cycle of these precipitating systems. We also analyze in more detail the intraseasonal variability and its interaction with the diurnal cycle. Note that all diurnal cycle analyses are

composited based on the local time at the center of each track for each time step.

a. Geographic importance of coastal precipitation

The climatology of coastal precipitation follows the expected seasonal variability (Fig. 1 first row). The monsoonal northern shift of precipitation toward the Bay of Bengal and Southeast Asia is visible between June and August (JJA). JJA is also characterized by high coastal precipitation around Central America, northern South America, the Caribbean, and tropical West Africa. From December to February (DJF), coastal precipitation is strong around Madagascar, northern Australia, and the South Pacific islands. On average, the highest coastal precipitation amounts are localized around the Maritime Continent and the Bight of Panama.

These high coastal precipitation values are responsible for the vast majority of the total local precipitation, with up to 80% of the total precipitation being detected as coastal (Fig. 1, second row). The seasonal differences between DJF and JJA are clearly visible for Madagascar, around India, the Bight of Panama, and the Caribbean. In local winter, coastal precipitation stays more concentrated at the coast while in summer, it often propagates over the surrounding ocean. Because precipitation rates are higher in local summer, the annual composites are more similar to the local summer composites. Note that the same composites for March–May (MAM) and September–November (SON) resemble the composites for JJA and DJF respectively (not shown).

Over the Maritime Continent, the fraction of coastal precipitation is relatively independent of the season and shows that the majority of precipitation results from convective systems initiated in the coastal region. The relatively low percentages over the central part of Borneo and Papua New Guinea are caused by the fact that tracks initiated over these areas are not taken into account because they start more than 48 km away from the coast (the same is true for Madagascar). Including them increases the percentage back to 80%–85% (not shown) and shows the importance of orographic effects in generating convection in these regions (Mapes et al. 2003; Qian et al. 2012).

The fraction of coastal precipitation is also noticeably high around the Red Sea, the Persian Gulf, and the coast of East Africa all year long, presumably because of the strong land–sea interaction of these arid or semiarid regions.

Overall, these results are very consistent with those of Bergemann et al. (2015). The main difference is the percentage of the coastal fraction relative to the total rainfall. While the fraction of coastal regions can reach 90% in specific regions of the Maritime Continent (Strait of Malacca, Makassar Strait, and Java) and goes up to 75%–80% in the areas mentioned above, this

FIG. 5. (a) Diurnal cycle of coastal precipitation for winter (black lines) and summer (red lines) over land (solid lines) and surrounding oceans (dotted lines) between 30°S and 30°N. Summer refers to the summer semester for each hemisphere, that is, from May to October in the Northern Hemisphere and from November to April in the Southern Hemisphere for the 18 years of the dataset. The same applies for winter. (b) Diurnal cycle over land for tropical islands of different sizes [size categories from Ulrich and Bellon (2019)] and for the continental coasts. The diurnal cycle is an average over 18 years.

fraction rarely exceeds 65% in their study. This results from their use of more restrictive criteria on the shape of the ellipse and its alignment with the coastline, as well as of more restrictive thresholds for rainfall intensity and for the overlap between coastal area and convective systems. Since we intend to include all convective systems initiated within 48 km from each side of the coast, and therefore are not limited to systems parallel to the coasts, the higher fraction of coastal precipitation seems reasonable and emphasizes the importance of coastal convective systems in many tropical regions.

b. Population of convective systems and diurnal cycle

We composite the coastal precipitation over the whole dataset to get a general idea of the characteristics of the coastal precipitation in summer and winter. The winter season is defined from November to April for the Northern Hemisphere and from May to October in the Southern Hemisphere, and vice versa for the summer. The diurnal cycle is calculated over the land regions of the coastal mask and over the oceans within the larger mask (400 km from

FIG. 6. Probability density function of the mean size of the coastal precipitation tracks as a function of their lifetime (shading). The normalized probability density function of the lifetimes is indicated by the red line.

the coast). Coastal precipitation increases over both land and ocean in summer (Fig. 5a). The characteristic peaks in the diurnal cycle over land in late afternoon and over the surrounding seas in the morning are visible in both seasons without change in its timing. The intensity of the diurnal precipitation maximum and the amplitude of the diurnal cycle are almost doubled in summer compared to winter, over both land and ocean, probably due to a stronger land–sea thermal contrast. Looking at annual-mean diurnal cycle over the different islands depending on their size as well as for the coasts of continents highlights the strong dependence of the timing of the diurnal cycle to the size of the island (Fig. 5b). These results show that our algorithm is able to isolate characteristic behaviors of tropical precipitation and confirm that coastal systems are the ones contributing to the well-known diurnal cycle over tropical islands (usually not distinguished from other convective systems when looking at diurnal cycle composites). Indeed, we find a progressively delayed peak of the diurnal cycle as the island size grows. The difference in amplitude is mostly an artifact of the islands included in each category. Indeed, the diurnal cycle is dominated by islands in the warm and humid Maritime Continent in the biggest category and the 8000–32 000 km² category. The dependence of the diurnal cycle on the size of the islands is beyond the scope of this study; it is further investigated in Ulrich and Bellon (2019).

Our algorithm allows us to investigate the relationship between the mean area covered by a track and its lifetime (Fig. 6). We checked that the growth and decay of the tracks' area for different lifetime categories are realistic and are not artifacts of the tracking algorithm (not shown). Not surprisingly, longer-lived tracks are associated with larger mean area. However, there is a transition in behavior for lifetimes inferior or superior to 6 h. This emphasizes the fact that young coastal systems only have two choices: they either grow in size and merge

FIG. 7. Diurnal cycle of coastal precipitation for each lifetime category over land (solid lines) and ocean (dotted lines) in (a) winter and (b) summer. For lifetimes above one day, the lines represent the average diurnal cycle for the category.

with other coastal systems (and live longer) or they die. The short-lived systems are mainly convective (high precipitation rates averaged over the track lifetime) and their mean precipitation increases with lifetime, probably because stronger systems have a higher chance to live longer (not shown). On the other hand, the long-lived systems that grow in size with lifetime have a somewhat constant and smaller mean precipitation, indicating similar fractions of convective and stratiform precipitation, even though the larger the track, the more likely it is to have a larger stratiform part and a reduced mean precipitation.

Therefore, it seems there is a clear limit over which we can predict that a system will either grow in size and live longer, or dissipate. The probability density function of the lifetime (red line in Fig. 6) shows that such growth is unlikely. Most tracked life cycles do not last longer than 3 h. It is noteworthy to point out that such systems are generally discarded by most existing tracking algorithms. This may not be a problem when studying mesoscale convective systems, but it ignores a large fraction of the coastal cloud population and probably a significant amount of coastal rainfall.

c. Contribution of systems to the diurnal cycle by lifetime

To clarify how much rainfall can be attributed to these small convective systems, we calculate the contributions of convective systems to the diurnal cycle as a function

of their lifetime, for winter and summer, using 3-h bins (Fig. 7). Overall, the contribution from each category is independent of the season. The systems contributing most, on average and to the afternoon peak over land, last between 6 and 12 h. Longer-lived systems (up to one day) also contribute a lot to land precipitation, especially at night, and control the diurnal cycle of coastal precipitation over the ocean. The precipitation over land increases for each category (except the smallest and the biggest) around 10 a.m., which shows that it is likely triggered by the establishment of the sea breeze and that the lifetime of a system is insensitive to its time of genesis. This is not true for systems lasting less than 6 h. They exhibit two distinct peaks of contribution: around 1 p.m. and 4 p.m., which indicate different triggering times. This might result from different large-scale conditions. Long-lived tracks (>15 h) also have a stronger increase in precipitation later in the afternoon. This increase may indicate that these systems are likely to get triggered slightly later and/or in more favorable conditions, which might explain why they also last longer.

The only notable difference between seasons is the larger contribution of the 3–6-h systems to the diurnal cycle in winter. As a result of the interseasonal similarity, the normalized cumulative distributions of coastal precipitation are almost identical in winter and summer for both land and ocean (Fig. 8) even though their diurnal cycles exhibit a large difference in amplitude (Fig. 5a). Figure 8 shows a significant difference between

FIG. 8. Normalized cumulative coastal precipitation in winter and summer (black and red, respectively) for land (solid lines) and the surrounding seas (dotted lines).

coastal precipitation over land and over ocean in terms of systems' size and lifespan: 65% of the precipitation over land originates from systems with a lifetime shorter than one day, whereas such systems contribute to 40% of the precipitation over ocean. This means that once coastal precipitation systems reach the ocean, they are likely to be sustained by mechanisms of oceanic convection, see their lifetime prolonged, and possibly interact with coastal systems formed on the following day or oceanic systems advected from the open ocean.

4. Diurnal cycle and MJO propagation

We now focus on the interaction between coastal and large-scale precipitation in the Maritime Continent. More specifically, we aim to assess the hypothesis from Peatman et al. (2014, hereafter P14) that the diurnal cycle over the Maritime Continent is amplified prior to the MJO envelope and results in increased precipitation over land. To compare with P14, only the austral summer MJO events (November to April) of the 18 years of coastal data are used. The MJO phase index is calculated following Wheeler and Hendon (2004) on the basis of the first two principal components of outgoing long-wave radiation (OLR) and zonal wind at 200 and 850 hPa, averaged over 15°S–15°N. We used the National Oceanic and Atmospheric Administration (NOAA) interpolated OLR (Liebmann and Smith 1996) and the zonal wind from the National Centers for Environmental Prediction–Department of Energy Reanalysis II (NCEP2; Kanamitsu et al. 2002).

This index classifies days into nine categories: one of nonactive MJO and eight phases of the MJO: phase 1 with a dry anomaly over the Indo-Pacific region and a hint of convection off the coast of eastern Africa, phase 2 with MJO convection developing over the western equatorial Indian Ocean, phase 3 with MJO convection reaching its mature phase over the eastern Indian Ocean, phase 4 with

the MJO convection encroaching over the Maritime Continent, phase 5 with the MJO convection spread out over the Maritime Continent, phase 6 with MJO convection withdrawing from the western Maritime Continent (and a dry anomaly developing over the Indian Ocean), phase 7 with MJO convection decreasing and propagating toward the central Pacific while the dry anomaly expands over the eastern Maritime Continent, and phase 8 with the dry anomaly over the Indian Ocean and most of the Maritime Continent. Using the MJO indices available online at www.bom.gov.au/climate/mjo/graphics/rmm.74toRealtime.txt gives very similar results.

a. Coastal precipitation and MJO phases

For each active MJO phase, a composite mean coastal precipitation is computed by averaging over all days within that phase. The coastal precipitation anomaly is calculated by subtracting the mean coastal precipitation from the composite for each phase (Fig. 9). This figure is very similar to Fig. 5 in P14 and shows that convection develops over land prior to the arrival of the MJO envelope: over Sumatra during phases 7–8–1, over Borneo during phases 8–1–2, and over New Guinea during phases 2–3. The enhancement of land precipitation occurs even slightly earlier than in P14 over Sumatra in phases 7–8, over Java in phase 1, and Papua New Guinea in phase 2. We checked that this does not come from differences between TRMM 3B42HQ precipitation used in P14 and CMORPH-CRT or from the period of study. P14 does not differentiate between coastal precipitation and large-scale precipitation. Considering the diurnal cycle of the “vanguard” of precipitation, it is not surprising that our dataset dedicated to coastal precipitation emphasizes a slightly earlier increase in coastal precipitation over land.

To emphasize both large-scale and coastal precipitation signals over the ocean and land in the Maritime Continent, we estimate the large-scale precipitation by subtracting the coastal precipitation from the CMORPH MJO composites of total precipitation. For this plot only, all the tracks initiated over the center of Borneo and Papua New Guinea contribute to coastal precipitation as it would not make sense to consider them as large scale. The latitudinal average (between 15°S and 15°N) by type of surface allows us to quantify when, where, and how much of the precipitation anomaly propagates over land and ocean during the MJO (Fig. 10). The large-scale MJO envelope appears over the ocean west of the MC in phase 3, covers the MC in phase 4, and maximizes over the eastern MC in phase 5 (bottom left). This large-scale signal slightly reinforces precipitation over New Guinea in phase 5 but does not really impact convection over land elsewhere and for other phases (bottom right). The coastal precipitation transect confirms that the

FIG. 9. Daily mean coastal precipitation anomaly for each phase of the MJO as defined by Wheeler and Hendon (2004). Only the austral summer MJO events (from November to April) are considered. Phases advance in the counterclockwise direction.

precipitation is reinforced over both land and ocean prior to the MJO (top row). The positive signal over ocean is smoother than over land and lags it by one phase (3–5 days). It is maximum during phase 4 over most of the

MC and contributes as much as the large-scale precipitation systems to the precipitation anomaly over the ocean in the central MC. This means that, while coastal precipitation is the main contributor to the “vanguard” of

FIG. 10. (top) Coastal precipitation anomaly for each phase of the austral summer MJOs averaged between 15°S–15°N over (a) ocean and (b) land. (bottom) As in (top), but for the large-scale precipitation (filtered out by the tracking algorithm) over (c) ocean and (d) land. The running average of the altitude over land (over 3° of longitude) is shown as a gray line in (b) and (d). The Joint Institute for the Study of the Atmosphere and the Ocean Terrain Base (TBASE) 0.25° elevation data are used to compute the altitude (https://www.research.jisao.washington.edu/data_sets/elevation/elev.0.25-deg.nc). Ocean is set to missing.

FIG. 11. Diurnal cycle of coastal precipitation for each phase of the MJO over land (solid lines) and the surrounding seas (dotted lines) for the Maritime Continent region (15°S – 20°N and 90° – 154°E). Only the austral summer MJO events (from November to April) are considered.

the MJO propagation over the MC, it also significantly contributes to the MJO envelope precipitation signature in the most active phases (4–5). It may also explain why land and coastal precipitation in central MC is crucial to the barrier effect of the MC (Hagos et al. 2016). They showed that because the diurnal cycle over land competes with the seas through land–sea breezes and moisture supply, when convection over the islands dominates, the MJO is unable to propagate through the MC. The fact that the central MC is the region where land and coastal precipitation have the strongest influence in phase 4

supports this conclusion. Once the MJO envelope has propagated farther east, it also tends to damp precipitation on land, which is consistent with Birch et al. (2016).

b. Modulation of the diurnal cycle by the MJO

To understand if the increase in coastal precipitation prior to the MJO is just related to a daily mean increase in precipitation or an increase in the diurnal cycle, we look at the amplitude of the diurnal cycle for each MJO phase. The diurnal cycle over land and ocean for each phase of the MJO is shown in Fig. 11. It is the same as Fig. 5, but for the Maritime Continent region rather than for the whole tropics. The diurnal cycle over the MC is much larger than the average tropical diurnal cycle, especially over land. It experiences considerable variability, which was already visible in the precipitation anomalies (Fig. 9). Over the ocean, the amplitude is slightly increased in phases 3 and 4. Over land, phases 2 and 3 have the largest diurnal cycle while phases 5 and 6 have the smallest, highlighting the fact that the MJO envelope reduces the diurnal cycle amplitude by up to 40%. From phase 3 on, the maximum of precipitation gradually extends toward early evening before peaking in late afternoon from phases 7 to 2.

This shift in time results mainly from a change in lifetime of the coastal convective systems (Fig. 12). Prior

FIG. 12. Diurnal cycle of coastal precipitation averaged for each lifetime category for each phase of the austral summer MJO. For lifetimes above one day, the lines represent the average diurnal cycle for the category.

FIG. 13. Normalized cumulative coastal precipitation for each phase of the austral summer MJO, over land (solid lines) and the surrounding seas (dotted lines).

to the MJO (phases 1 and 2), the early peak in precipitation results from the large contribution of clouds lasting between 6 and 12 h. From phase 1 to phase 5, the contribution of these short-lived systems gradually weakens (as well as that of the shortest systems) while that of systems lasting between 15 h and one day increases and becomes dominant in phase 4. Everything reverses from phase 5 to phase 1. This emphasizes the progressive reinforcement of the diurnal cycle over land prior to the MJO, when the large-scale conditions become favorable for the development and maintenance of convection. It translates into a larger fraction of the rainfall from clouds lasting more than one day in phases 2 to 4: 40%–45% compared 35% during the drier phases 7 to 1 (Fig. 13). Cloud systems last longer and can be advected over the surrounding oceans. Additionally, the stronger diurnal cycle over land can also reinforce the diurnal cycle over the surrounding ocean. That is why the contribution of clouds lasting more than a day to the oceanic coastal precipitation is 10% larger just before or during the MJO, and lags the land maximum by one phase.

It is important to keep in mind that even though the MJO envelope is moving through in the active phases, our coastal dataset does not include the large-scale convective systems advected with the MJO or the oceanic systems within the MJO envelope. In that regard, the last figures give us valuable information on the coastal aspect of the MJO propagation over the Maritime Continent. It shows how the life cycle of coastal convective systems evolves depending on the large-scale conditions and how these systems can interact with the large-scale MJO envelope and contribute to its propagation.

5. Summary and conclusions

In summary, we propose a tracking algorithm and apply it to 18 years of high-resolution 30-min, 8-km

CMORPH-CRT precipitation data over the tropics. The main purpose of the algorithm is to isolate precipitation that is only influenced by coasts and islands, and to follow the coastal convective systems as they evolve with time. To do so, we define tracks that encompass all the individual cloud clusters belonging to the same convective system. These tracks grow in size as separate convective clusters merge and, in case of splitting, follow the smaller distinct systems keeping record of their common origin. Since we want to study coastal precipitation, we remove all the tracks that are not initiated within 48 km on each side of the coastline. Synoptic-scale tracks are also removed. This algorithm provides early detection of coastal convective systems and great detail on their life cycle. In particular, our algorithm is able to detect and track short-lived systems (with a lifespan shorter than 6 h), which are usually discarded in tracking algorithms of convective systems.

Another strength of the tracking algorithm presented here is that it is relatively straightforward to apply. It tracks coastal systems as soon as they appear with a time-efficient moment method that fits ellipses to the convective clusters. For this reason, it could be used on cloud-resolving models outputs to understand the physical mechanisms responsible for the diurnal propagation of convective systems or other phenomena. It could help distinguish the large-scale conditions in which coastal precipitation is likely to occur on average over the tropics or specific regions. Because the algorithm is built to work on any 2D image, it could also be used for variables other than precipitation with some adjustments of parameters and thresholds.

We exploit the new dataset of coastal precipitation produced by this algorithm to highlight key features. We find the following:

- On average in the tropics, the diurnal cycle over land and the surrounding oceans has a similar timing in all seasons but is much stronger in local summer.
- The vast majority of the tracked coastal systems last less than 6 h in winter and summer. They contribute to less than 10% of the average precipitation and the diurnal cycle. They are triggered at two different times in the afternoon, with a slightly larger contribution in local winter. For those short-lived systems, the maximum size of the track is small and insensitive to its lifetime.
- For longer-lived tracks, there is a linear relationship between lifetime and maximum size: the longer a track lives, the larger it becomes.
- The contribution of short-lived coastal systems (6–12 h) dominates the signal in early afternoon over land, while longer-lived systems (15–24 h) dominate

the late afternoon and evening precipitation over land, as well as the early morning rain over the oceans.

- Overall, the relative contribution of each lifetime category to the diurnal cycle is independent of the season. Over land, 65% of the rain originates from cloud systems lasting less than a day. This proportion goes down to 40% over the ocean.

We apply our coastal precipitation dataset to confirm results from Peatman et al. (2014), who posited that coastal diurnal precipitation over land is enhanced prior to the MJO over the Maritime Continent. Our findings confirm their results and complement them. When only considering coastal precipitation, the diurnal cycle over land appears amplified slightly earlier in the MJO propagation than what they found (at least for some islands).

The distinction between large-scale precipitation and coastal precipitation allows us to quantify the role coastal precipitation plays in the propagation of the MJO over the Maritime Continent. The increased diurnal cycle over land during phases 1–3 slowly propagates over the surrounding seas during phases 2–4. This is visible in the shift in lifetime of clouds toward longer-lasting clouds over land and then over the ocean prior to the arrival of the MJO envelope. On the other hand, large-scale precipitation only increases in phases 4 and 5 over the seas and phase 5 over New Guinea. When the MJO envelope arrives over an island, the precipitation diurnal cycle is systematically reduced, as shown by several studies which attribute it to a larger cloud cover over the islands and a resulting smaller diurnal heating (Birch et al. 2016; Hagos et al. 2016). In phase 4, when the MJO envelope is over the Maritime Continent, large-scale precipitation contributes to approximately 50% of precipitation over both ocean and land in the central Maritime Continent (110°–140°E). This emphasizes the importance of the diurnal cycle and coastal precipitation for the barrier effect of the Maritime Continent as presented by Hagos et al. (2016).

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programme (<https://www.nesi.org.nz>). The tracking algorithm in Python is available at <https://github.com/DavidCoppin/tracking>. The tracked coastal dataset is available at <https://www.dropbox.com/sh/fpwl5dmsvyaby9q/AADAE-O5y3R3TdLu4WYTkqgLa?dl=0>. The CMORPH satellite-based rainfall estimates were obtained from the Climate Prediction Center (CPC) of the NOAA. NOAA Interpolated OLR and NCEP2 data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, from their website at <https://www.esrl.noaa.gov/psd/>.

REFERENCES

- Aspliden, C. I., Y. Tourre, and J. B. Sabine, 1976: Some climatological aspects of West African disturbance lines during GATE. *Mon. Wea. Rev.*, **104**, 1029–1035, [https://doi.org/10.1175/1520-0493\(1976\)104<1029:SCAOWA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1976)104<1029:SCAOWA>2.0.CO;2).
- Bergemann, M., C. Jakob, and T. P. Lane, 2015: Global detection and analysis of coastline-associated rainfall using an objective pattern recognition technique. *J. Climate*, **28**, 7225–7236, <https://doi.org/10.1175/JCLI-D-15-0098.1>.
- Birch, C. E., S. Webster, S. C. Peatman, D. J. Parker, A. J. Matthews, Y. Li, and M. E. Hassim, 2016: Scale interactions between the MJO and the western Maritime Continent. *J. Climate*, **29**, 2471–2492, <https://doi.org/10.1175/JCLI-D-15-0557.1>.
- Boer, E. R., and V. Ramanathan, 1997: Lagrangian approach for deriving cloud characteristics from satellite observations and its implications to cloud parameterization. *J. Geophys. Res.*, **102**, 21 383–21 399, <https://doi.org/10.1029/97JD00930>.
- Coppin, D., and G. Bellon, 2019a: Physical mechanisms controlling the offshore propagation of convection in the tropics: 1. Flat island. *J. Adv. Model. Earth Syst.*, **11**, 3042–3056, <https://doi.org/10.1029/2019MS001793>.
- , and —, 2019b: Physical mechanisms controlling the offshore propagation of convection in the tropics: 2. Influence of topography. *J. Adv. Model. Earth Syst.*, **11**, 3251–3264, <https://doi.org/10.1029/2019MS001794>.
- Cronin, T. W., K. A. Emanuel, and P. Molnar, 2015: Island precipitation enhancement and the diurnal cycle in radiative-convective equilibrium. *Quart. J. Roy. Meteor. Soc.*, **141**, 1017–1034, <https://doi.org/10.1002/qj.2443>.
- Ebert, E. E., J. E. Janowiak, and C. Kidd, 2007: Comparison of near-real-time precipitation estimates from satellite observations and numerical models. *Bull. Amer. Meteor. Soc.*, **88**, 47–64, <https://doi.org/10.1175/BAMS-88-1-47>.
- Feng, Z., X. Dong, B. Xi, S. McFarlane, A. Kennedy, B. Lin, and P. Minnis, 2012: Life cycle of midlatitude deep convective systems in a Lagrangian framework. *J. Geophys. Res.*, **117**, D23201, <https://doi.org/10.1029/2012JD018362>.
- Fiolleau, T., and R. Roca, 2013a: An algorithm for the detection and tracking of tropical mesoscale convective systems using infrared images from geostationary satellite. *IEEE Trans. Geosci. Remote Sens.*, **51**, 4302–4315, <https://doi.org/10.1109/TGRS.2012.2227762>.
- , and —, 2013b: Composite life cycle of tropical mesoscale convective systems from geostationary and low Earth orbit satellite observations: Method and sampling considerations. *Quart. J. Roy. Meteor. Soc.*, **139**, 941–953, <https://doi.org/10.1002/qj.2174>.
- Futyan, J. M., and A. D. Del Genio, 2007: Deep convective system evolution over Africa and the tropical Atlantic. *J. Climate*, **20**, 5041–5060, <https://doi.org/10.1175/JCLI4297.1>.

- Guo, H., S. Chen, A. Bao, J. Hu, A. S. Gebregiorgis, X. Xue, and X. Zhang, 2015: Inter-comparison of high-resolution satellite precipitation products over Central Asia. *Remote Sens.*, **7**, 7181–7211, <https://doi.org/10.3390/rs70607181>.
- Hagos, S., Z. Feng, S. McFarlane, and L. R. Leung, 2013: Environment and the lifetime of tropical deep convection in a cloud-permitting regional model simulation. *J. Atmos. Sci.*, **70**, 2409–2425, <https://doi.org/10.1175/JAS-D-12-0260.1>.
- , C. Zhang, Z. Feng, C. D. Burleyson, C. De Mott, B. Kerns, J. J. Benedict, and M. N. Martini, 2016: The impact of the diurnal cycle on the propagation of Madden-Julian Oscillation convection across the Maritime Continent. *J. Adv. Model. Earth Syst.*, **8**, 1552–1564, <https://doi.org/10.1002/2016MS000725>.
- Hassim, M. E., T. P. Lane, and W. W. Grabowski, 2016: The diurnal cycle of rainfall over New Guinea in convection-permitting WRF simulations. *Atmos. Chem. Phys.*, **16**, 161–175, <https://doi.org/10.5194/acp-16-161-2016>.
- Hendon, H. H., and M. L. Salby, 1994: The life cycle of the Madden-Julian oscillation. *J. Atmos. Sci.*, **51**, 2225–2237, [https://doi.org/10.1175/1520-0469\(1994\)051<2225:TLCOTM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1994)051<2225:TLCOTM>2.0.CO;2).
- Ichikawa, H., and T. Yasunari, 2007: Propagating diurnal disturbances embedded in the Madden-Julian Oscillation. *Geophys. Res. Lett.*, **34**, L18811, <https://doi.org/10.1029/2007GL030480>.
- , and —, 2008: Intraseasonal variability in diurnal rainfall over New Guinea and the surrounding oceans during austral summer. *J. Climate*, **21**, 2852–2868, <https://doi.org/10.1175/2007JCLI1784.1>.
- Janowiak, J. E., V. E. Kousky, and R. J. Joyce, 2005: Diurnal cycle of precipitation determined from the CMORPH high spatial and temporal resolution global precipitation analyses. *J. Geophys. Res.*, **110**, D23105, <https://doi.org/10.1029/2005JD006156>.
- Jiang, X., and Coauthors, 2015: Vertical structure and physical processes of the Madden-Julian oscillation: Exploring key model physics in climate simulations. *J. Geophys. Res.*, **120**, 4718–4748, <https://doi.org/10.1002/2014JD022375>.
- Joyce, R. J., J. E. Janowiak, P. A. Arkin, and P. Xie, 2004: CMORPH: A method that produces global precipitation estimates from passive microwave and infrared data at high spatial and temporal resolution. *J. Hydrometeorol.*, **5**, 487–503, [https://doi.org/10.1175/1525-7541\(2004\)005<0487:CAMTPG>2.0.CO;2](https://doi.org/10.1175/1525-7541(2004)005<0487:CAMTPG>2.0.CO;2).
- Kanamitsu, M., W. Ebisuzaki, J. Woollen, S.-K. Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter, 2002: NCEP-DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc.*, **83**, 1631–1643, <https://doi.org/10.1175/BAMS-83-11-1631>.
- Kim, D., and Coauthors, 2009: Application of MJO simulation diagnostics to climate models. *J. Climate*, **22**, 6413–6436, <https://doi.org/10.1175/2009JCLI3063.1>.
- Klein, R., R. Nicholls, and F. Thomalla, 2003: The resilience of coastal megacities to weather related hazards: A review. *Building Safer Cities: The Future of Disaster Risk*, The World Bank, 111–137, <https://doi.org/10.1596/0-8213-5497-3>.
- Li, Y., N. C. Jourdain, A. S. Taschetto, A. Sen Gupta, D. Argüeso, S. Masson, and W. Cai, 2017: Resolution dependence of the simulated precipitation and diurnal cycle over the Maritime Continent. *Climate Dyn.*, **48**, 4009–4028, <https://doi.org/10.1007/s00382-016-3317-y>.
- Liebmann, B., and C. A. Smith, 1996: Description of a complete (interpolated) outgoing longwave radiation dataset. *Bull. Amer. Meteor. Soc.*, **77**, 1275–1277.
- Love, B. S., A. J. Matthews, and G. M. S. Lister, 2011: The diurnal cycle of precipitation over the Maritime Continent in a high-resolution atmospheric model. *Quart. J. Roy. Meteor. Soc.*, **137**, 934–947, <https://doi.org/10.1002/qj.809>.
- Madden, R. A., and P. R. Julian, 1971: Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific. *J. Atmos. Sci.*, **28**, 702–708, [https://doi.org/10.1175/1520-0469\(1971\)028<0702:DOADOI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1971)028<0702:DOADOI>2.0.CO;2).
- , and —, 1972: Description of global-scale circulation cells in the tropics with a 40–50 day period. *J. Atmos. Sci.*, **29**, 1109–1123, [https://doi.org/10.1175/1520-0469\(1972\)029<1109:DOGCC>2.0.CO;2](https://doi.org/10.1175/1520-0469(1972)029<1109:DOGCC>2.0.CO;2).
- , and —, 1994: Observations of the 40–50-day tropical oscillation—A review. *Mon. Wea. Rev.*, **122**, 814–837, [https://doi.org/10.1175/1520-0493\(1994\)122<0814:OOTDIO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1994)122<0814:OOTDIO>2.0.CO;2).
- Mapes, B. E., and R. A. Houze, 1993: Cloud clusters and superclusters over the oceanic warm pool. *Mon. Wea. Rev.*, **121**, 1398–1415, [https://doi.org/10.1175/1520-0493\(1993\)121<1398:CCASOT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1993)121<1398:CCASOT>2.0.CO;2).
- , T. T. Warner, and M. Xu, 2003: Diurnal patterns of rainfall in northwestern South America. Part III: Diurnal gravity waves and nocturnal convection offshore. *Mon. Wea. Rev.*, **131**, 830–844, [https://doi.org/10.1175/1520-0493\(2003\)131<0830:DPORIN>2.0.CO;2](https://doi.org/10.1175/1520-0493(2003)131<0830:DPORIN>2.0.CO;2).
- Martin, D. W., and A. J. Schreiner, 1981: Characteristics of West African and east Atlantic cloud clusters: A survey from GATE. *Mon. Wea. Rev.*, **109**, 1671–1688, [https://doi.org/10.1175/1520-0493\(1981\)109<1671:COWAAE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1981)109<1671:COWAAE>2.0.CO;2).
- Mori, S., and Coauthors, 2004: Diurnal land-sea rainfall peak migration over Sumatra Island, Indonesian Maritime Continent, observed by TRMM satellite and intensive rawinsonde soundings. *Mon. Wea. Rev.*, **132**, 2021–2039, [https://doi.org/10.1175/1520-0493\(2004\)132<2021:DLRPMO>2.0.CO;2](https://doi.org/10.1175/1520-0493(2004)132<2021:DLRPMO>2.0.CO;2).
- Moron, V., A. W. Robertson, J.-H. Qian, and M. Ghil, 2015: Weather types across the Maritime Continent: From the diurnal cycle to interannual variations. *Front. Environ. Sci.*, **2** (65), <https://doi.org/10.3389/fenvs.2014.00065>.
- Nesbitt, S. W., and E. J. Zipser, 2003: The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements. *J. Climate*, **16**, 1456–1475, <https://doi.org/10.1175/1520-0442-16.10.1456>.
- Nicholls, R. J., and A. Cazenave, 2010: Sea-level rise and its impact on coastal zones. *Science*, **328**, 1517–1521, <https://doi.org/10.1126/science.1185782>.
- Ogino, S.-Y., M. D. Yamanaka, S. Mori, and J. Matsumoto, 2016: How much is the precipitation amount over the tropical coastal region? *J. Climate*, **29**, 1231–1236, <https://doi.org/10.1175/JCLI-D-15-0484.1>.
- , —, —, and —, 2017: Tropical coastal dehydrator in global atmospheric water circulation. *Geophys. Res. Lett.*, **44**, 11 636–11 643, <https://doi.org/10.1002/2017GL075760>.
- Peatman, S. C., A. J. Matthews, and D. P. Stevens, 2014: Propagation of the Madden-Julian Oscillation through the Maritime Continent and scale interaction with the diurnal cycle of precipitation. *Quart. J. Roy. Meteor. Soc.*, **140**, 814–825, <https://doi.org/10.1002/qj.2161>.
- , —, and —, 2015: Propagation of the Madden-Julian oscillation and scale interaction with the diurnal cycle in a high-resolution GCM. *Climate Dyn.*, **45**, 2901–2918, <https://doi.org/10.1007/s00382-015-2513-5>.
- Qian, J.-H., 2008: Why precipitation is mostly concentrated over islands in the Maritime Continent. *J. Atmos. Sci.*, **65**, 1428–1441, <https://doi.org/10.1175/2007JAS2422.1>.
- , A. W. Robertson, and V. Moron, 2013: Diurnal cycle in different weather regimes and rainfall variability over Borneo

- associated with ENSO. *J. Climate*, **26**, 1772–1791, <https://doi.org/10.1175/JCLI-D-12-00178.1>.
- Qian, T., C. C. Epifanio, and F. Zhang, 2012: Topographic effects on the tropical land and sea breeze. *J. Atmos. Sci.*, **69**, 130–149, <https://doi.org/10.1175/JAS-D-11-011.1>.
- Rauniyar, S. P., A. Protat, and H. Kanamori, 2017: Uncertainties in TRMM-era multisatellite-based tropical rainfall estimates over the Maritime Continent. *Earth Space Sci.*, **4**, 275–302, <https://doi.org/10.1002/2017EA000279>.
- Rui, H., and B. Wang, 1990: Development characteristics and dynamic structure of tropical intraseasonal convection anomalies. *J. Atmos. Sci.*, **47**, 357–379, [https://doi.org/10.1175/1520-0469\(1990\)047<0357:DCADSO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1990)047<0357:DCADSO>2.0.CO;2).
- Saito, K., T. D. Keenan, G. Holland, and K. Puri, 2001: Numerical simulation of the diurnal evolution of tropical island convection over the Maritime Continent. *Mon. Wea. Rev.*, **129**, 378–400, [https://doi.org/10.1175/1520-0493\(2001\)129<0378:NSOTDE>2.0.CO;2](https://doi.org/10.1175/1520-0493(2001)129<0378:NSOTDE>2.0.CO;2).
- Seo, K.-H., W. Wang, J. Gottschalck, Q. Zhang, J.-K. E. Schemm, W. R. Higgins, and A. Kumar, 2009: Evaluation of MJO forecast skill from several statistical and dynamical forecast models. *J. Climate*, **22**, 2372–2388, <https://doi.org/10.1175/2008JCLI2421.1>.
- Small, C., and R. Nicholls, 2003: A global analysis of human settlement in coastal zones. *J. Coast. Res.*, **19**, 584–599.
- Sobel, A. H., C. D. Burleyson, and S. E. Yuter, 2011: Rain on small tropical islands. *J. Geophys. Res.*, **116**, D08102, <https://doi.org/10.1029/2010JD014695>.
- Tuanmu, M. N., and W. Jetz, 2014: A global 1-km consensus land-cover product for biodiversity and ecosystem modelling. *Global Ecol. Biogeogr.*, **23**, 1031–1045, <https://doi.org/10.1111/geb.12182>.
- Ulrich, M., and G. Bellon, 2019: Super-enhancement of precipitation at the center of tropical islands. *Geophys. Res. Lett.*, <https://doi.org/10.1029/2019GL084947>, in press.
- Vincent, C. L., and T. P. Lane, 2016: Evolution of the diurnal precipitation cycle with the passage of a Madden–Julian oscillation event through the Maritime Continent. *Mon. Wea. Rev.*, **144**, 1983–2005, <https://doi.org/10.1175/MWR-D-15-0326.1>.
- Vitart, F., and F. Molteni, 2010: Simulation of the Madden–Julian oscillation and its teleconnections in the ECMWF forecast system. *Quart. J. Roy. Meteor. Soc.*, **136**, 842–855, <https://doi.org/10.1002/qj.623>.
- Wang, S., and A. H. Sobel, 2017: Factors controlling rain on small tropical islands: Diurnal cycle, large-scale wind speed, and topography. *J. Atmos. Sci.*, **74**, 3515–3532, <https://doi.org/10.1175/JAS-D-16-0344.1>.
- Wheeler, M. C., and H. H. Hendon, 2004: An all-season real-time multivariate MJO index: Development of an index for monitoring and prediction. *Mon. Wea. Rev.*, **132**, 1917–1932, [https://doi.org/10.1175/1520-0493\(2004\)132<1917:AARMMI>2.0.CO;2](https://doi.org/10.1175/1520-0493(2004)132<1917:AARMMI>2.0.CO;2).
- Williams, M., and R. A. J. Houze, 1987: Satellite-observed characteristics of winter monsoon cloud clusters. *Mon. Wea. Rev.*, **115**, 505–519, [https://doi.org/10.1175/1520-0493\(1987\)115<0505:SOCOWM>2.0.CO;2](https://doi.org/10.1175/1520-0493(1987)115<0505:SOCOWM>2.0.CO;2).
- Xie, P., and A.-Y. Xiong, 2011: A conceptual model for constructing high-resolution gauge-satellite merged precipitation analyses. *J. Geophys. Res.*, **116**, D21106, <https://doi.org/10.1029/2011JD016118>.
- Yanase, A., K. Yasunaga, and H. Masunaga, 2017: Relationship between the direction of diurnal rainfall migration and the ambient wind over the southern Sumatra Island. *Earth Space Sci.*, **4**, 117–127, <https://doi.org/10.1002/2016EA000181>.
- Yang, G., and J. Slingo, 2001: The diurnal cycle in the tropics. *Mon. Wea. Rev.*, **129**, 784–801, [https://doi.org/10.1175/1520-0493\(2001\)129<0784:TDCITT>2.0.CO;2](https://doi.org/10.1175/1520-0493(2001)129<0784:TDCITT>2.0.CO;2).
- Yokoi, S., S. Mori, M. Katsumata, B. Geng, K. Yasunaga, F. Syamsudin, Nurhayati, and K. Yoneyama, 2017: Diurnal cycle of precipitation observed in the western coastal area of Sumatra Island: offshore preconditioning by gravity waves. *Mon. Wea. Rev.*, **145**, 3745–3761, <https://doi.org/10.1175/MWR-D-16-0468.1>.
- Zhang, C., 2005: Madden–Julian Oscillation. *Rev. Geophys.*, **43**, RG2003, <https://doi.org/10.1029/2004RG000158>.