Diurnal variations of the areas and temperatures in tropical cyclone clouds

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Diurnal variations of the areas and temperatures in tropical cyclone convective cloud systems in the western North Pacific were estimated using pixel-resolution infrared (IR) brightness temperature (BT) and best-track data for 2000–2013. The mean areal extent of very cold cloud cover (IR BTs < 208 K) reached a maximum in the early morning (0000–0300 local solar time (LST)), then decreased after sunrise. This was followed by increasing cloud cover between 208 and 240 K, reaching its maximum areal extent in the afternoon (1500–1800 LST). The time at which cloud cover reached a maximum was sensitive to the temperature thresholds used over the ocean. IR BTs < 240 K reached minima in the morning (0300–0600 LST), and IR BTs > 240 K reached minima in the afternoon (1500–1800 LST). The out-of-phase relationships between IR BTs < 240 K and IR BTs > 240 K, and between the maximum coverage times of IR BTs < 208 K and 208 K < IR BTs < 240 K, can both lead to the radius-averaged IR temperature having two minima per day. The different diurnal evolutions under different cloud conditions suggest tropical cyclone convective cloud systems are best described in terms of both areal extent and cloud-top temperature. Maximum occurrence of clouds with IR BTs < 208 K in the morning and maximum occurrence of clouds with 208 K < IR BTs < 240 K in the afternoon suggest that two different mechanisms might be involved in causing diurnal variations under these two types of tropical cyclone cloud conditions.

Key Words: tropical cyclone; diurnal cycle; cloud

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1. Introduction

Tropical cyclones (TCs) are major producers of both cloud cover and precipitation in the Tropics and Subtropics. Cloud cover and precipitation in TCs both show marked diurnal cycle signatures (Shu et al., 2013; Dunion et al., 2014; Bowman and Fowler, 2015; Wu et al., 2015). Recently acquired cloud-resolving numerical modelling results have suggested that radiative forcing accelerates the rate of tropical cyclogenesis and causes early intensification (Melhauser and Zhang, 2014). It has been suggested that the TC diurnal cycle has an important influence on the structure of a TC and possibly on its intensity as well (Dunion et al., 2014; Ge et al., 2014), but the mechanisms involved in causing diurnal cycles in TCs remain unclear.

The diurnal convection cycle is caused by incoming solar radiation, which peaks at local noon. Convective precipitation over land reaches a maximum in the late afternoon and is thought to be a direct response to daytime heating of the surface and the planetary boundary layer (e.g. Janowiak et al., 1994; Yang and Slingo, 2001). Maximum cloud cover over the open ocean tends to occur in the afternoon or early evening, whereas maximum deep cloud coverage occurs in the early morning (Yang and Slingo, 2001). Tropical ocean deep convective peaks were also found in the early morning in the idealized modelling studies of Liu and Moncrieff (1998). The processes controlling diurnal cloudiness and rain cycles over the ocean are the subject of ongoing debate and are less well understood than those over land. Differential radiative heating between the convective region and the surrounding cloud-free region is considered important according to some theories (Gray and Jacobson, 1977). It has also been suggested that the morning maximum deep cloud cover is caused by a direct radiation–convection effect in which afternoon convection is suppressed because more solar radiation is absorbed by the cloud tops, stabilizing the air and suppressing convection, and night-time convection is enhanced because radiative cooling of the cloud tops increases instability and promotes convection (Randall et al., 1991; Yang and Slingo, 2001). Chen and Houze (1997) linked the morning maximum deep cloud cover to the life cycle of cloud systems and diurnal solar heating of the ocean surface and atmospheric boundary layer. Nesbitt and Zipser...
(2003) argued that the morning maximum precipitation rate is caused by increased numbers of mesoscale convective systems, the growth of which is favoured and the lifetimes of which can be long during the night. In addition to those theories associated with solar radiation, Li and Wang (2012) provided an alternative explanation on the diurnal variation of the cloud canopy of observed TCs. A period of 22–26 h of outer spiral rain bands (outside a radius of about three times the radius of maximum wind) was simulated in a TC in the full compressible, non-hydrostatic cloud-resolving Tropical Model version 4 (TMC4) even without diurnal radiative forcing included in the model simulation (Wang, 2009).

The quasi-diurnal occurrence of outer spiral rain bands was considered to be associated with the boundary-layer recovery from the effect of convective downdrafts and the consumption of convective available potential energy by convection in the previous outer spiral rain bands (Li and Wang, 2012).

Infrared (IR) satellite images have been used in a number of previous studies to identify diurnal maxima and minima associated with tropical convection and TC cloud patterns. However, there are some inconsistencies among the specific features of this well-documented diurnal cycle, particularly in the phases of the cycles. Diurnal variations in the areal extents of TC clouds have been studied using cloud-top temperatures below specific thresholds (e.g. Browner et al., 1977; Muramatsu, 1983; Lajoie and Butterworth, 1984; Steranka et al., 1984). Browner et al. (1977) analysed eight Atlantic tropical storms and found that the cloud area reached a maximum at 1700 local solar time (LST) and a minimum at 0300 LST. Similar results were found by Steranka et al. (1984) for the outer rain-band regions of 23 Atlantic TCs. However, the cloud area in the inner core region, with very low brightness temperatures (BTs), reached a maximum in the early morning (Steranka et al., 1984). Lajoie and Butterworth (1984) analysed data for 11 TCs near Australia and observed a marked diurnal oscillation with a maximum area within 3 h of 0300 LST and a minimum area within 3 h of 1800 LST, and also found a weaker daytime oscillation with maximum and minimum areas that occurred most frequently within 3 h of 1200 and 0900 LST, respectively.

Diurnal variations in IR BTs associated with TC cloud-top temperatures have been evaluated using average temperatures within a fixed radius or annulus (e.g. Steranka et al., 1984; Kossin, 2002; Dunion et al., 2014). Steranka et al. (1984) found a significant diurnal oscillation in the cloud-top temperature that explained a large percentage of the variance in each annulus ranging from the inner core to the storm periphery, hundreds of kilometres from the centre. Besides diurnal cycles, Steranka et al. (1984) found semidiurnal cloud-top temperature cycles in the outer peripheries of tropical storms. Kossin (2002) used IR cloud-top temperature measurements to analyse, separately, 21 Atlantic storms that occurred in 1999, and also found semi-diurnal oscillations. These semi-diurnal oscillations were found within all annuli, but were especially prevalent in the innermost and outermost regions. A few of the storms even had powerful spectral peaks at high frequencies and periods of 7–10 h. A general absence of significant diurnal oscillations in BT near the convective centres of hurricanes led Kossin (2002) to conclude that diurnal oscillations of cirrus canopies might not be physically linked to convection. Kossin (2002) suggested that the semi-diurnal solar atmospheric tide is linked to semi-diurnal cloud variations via a mechanism based on the variability of the convergence. Dunion et al. (2014) recently found diurnal pulses in cloud fields that propagate radially outward from the storm centres of mature hurricanes in low wind-shear environments in the North Atlantic. These mature hurricanes were constrained to their storm centres, 300 km from land. As well as this diurnal cycle, Dunion et al. (2014) found statistically significant cycles (of around 0.5–0.75 cycles per day) at 100–400 km radius, but the causes of these cycles were not clear.

The disagreements among the results of previous studies may be caused by the relatively small number of storms for which observational databases exist and the different analytical methods used. Diurnal cycles in the areal extent of clouds and in the cloud-top temperature in a TC may be caused by the presence of clouds with different properties. Satellite IR sensors only provide indirect estimates of the properties of deep convective clouds, and the properties of the interiors of such clouds cannot be determined. Cloud-top temperatures measured using satellite IR sensors are generally similar for deep convective clouds and cirrus clouds (e.g. Liu et al., 1995; Sui et al., 1997). The average temperature within a fixed radius or annulus includes diurnal signals from different types of cloud. Different cycle parameters are found when the signals for different cloud conditions are combined, once the diurnal cycles of the areal cloud extent and temperature are not in phase for the different cloud conditions. Rather than studying diurnal variations in TC clouds with a fixed radius or annulus, we herein consider daily variability for whole convective clouds in TCs in terms of both the areal cloud extent and temperature, in order to allow the discrepancies between previous studies to be resolved.

2. Data and methods

Best-track data for the western North Pacific were obtained from the US Navy Joint Typhoon Warning Center (JTWC: Chu et al., 2002). Storm parameters were typically recorded at 0000, 0600, 1200 and 1800 UTC. Six-hourly measurements of the location of the TC centre, the intensity of the TC, and other important parameters were included in the best-track data. We used 6-hourly observations for the period 2000–2013. A total of 391 storms that reached tropical storm intensity level or higher were recorded in the western North Pacific during the study period. The TCs were separated into weak (tropical storm to TC category 1) and strong storms (TC categories 2–5) to allow differences in diurnal variations in storms of different intensities to be examined. Storms of TC category 2 are classed as strong storms here because not many of the storms were in TC categories 3–5.

We used IR BT (equivalent to the black-body temperature) data with a pixel size of 4 × 4 km² (Janowiak et al., 2001) from the US National Centers for Environmental Prediction Climate Prediction Center. Globally merged (60°S to 60°N) IR BT data were produced by merging data from all the available geostationary satellites (GOES-8/10, Meteosat-7/5 and GMS). The peak frequencies of the IR channels used were 10.7, 11.5 and 11.0 μm for the GOES-8/10, Meteosat-7/5 and GMS data, respectively. The IR data obtained from these instruments will not be averaged from scene to scene for simplicity. However, these effects are considerably smaller than the viewing geometry effects. For the same target in regions, the mean difference of each sensor is determined and ‘calibrated’ by the sensors aboard the neighbouring satellite. The IR satellite images used typically indicate high-level cirrus in the TC canopy and embedded deep convection. The data were corrected for ‘zenith angle dependence’. The IR temperatures at locations far from the satellite nadir would have been lower than the actual temperatures because of geometric effects and radiometric path extinction effects (Joyce et al., 2001). The zenith angle dependence correction removes, to a large extent, the discontinuities at the boundaries between the areas covered by the different geostationary satellites when IR data from the satellites are merged. GOES full-disc views are guaranteed only eight times daily at 0000, 0300 . . . 2100 UTC. For images not at these times, the GOES data may be assembled from various regional subsets of a full-disclosure view. Global IR composites are available for every half-hour via a weekly rotating file. The half-hour data were averaged to give hourly images to reduce the number of data gaps caused by satellite eclipse periods. A total of 34 186 satellite images were collected for weak storms (tropical storm to TC category 1) and 8274 satellite images were collected for strong storms (TC categories 2–5). The temperature data were adjusted to LST for each longitude grid line.
In many previous studies, IR BTs of 230–240 K have been used to indicate the presence of convective clouds over both land and ocean (e.g. Yang and Slingo, 2001; Wilcox, 2003; Tian et al., 2004). Machado et al. (2002) and Hong et al. (2006) used an IR BT < 210 K and an IR BT < 235 K to detect deep convective clouds and high clouds, respectively. It has been suggested that an IR BT < 208 K is a conservative indicator of precipitating deep convective clouds in the western Pacific (Chen and Houze, 1997). We refer to these previous studies in assigning IR BT ranges to three categories of clouds, namely very cold deep convective clouds (IR BT < 208 K), cold high clouds (208 K < IR BT < 240 K), and low-level clouds and clear sky (IR BT > 240 K).

Diurnal cycles in TC convective systems were identified by analysing all the IR BTs within 500 km of each TC centre. The same radius was used in previous studies of TC precipitation (e.g. Lau et al., 2008; Jiang and Zipser, 2010; Prat and Nelson, 2013; Wu et al., 2015) and reflects the typical radius of the curved TC cloud shield (550–600 km) (Prat and Nelson, 2013). Prat and Nelson (2013) found that TC rainfall was little different between 500 and 1000 km of a TC centre. Our analysis focused on the open ocean, and satellite images including land masses less than 300 km from a storm centre were not considered. We considered only large land masses to be ‘land’. Satellite images including islands less than 300 km from a storm centre were not excluded. The 6-hourly TC centre position data were linearly interpolated to give 3-hourly TC centre positions. The hourly IR satellite images were matched to the appropriate 3 h intervals for which the TC centre positions were interpolated.

3. Results

An example of the TC diurnal cycle of the areas and cloud-top temperatures for Typhoon Saola on 29 and 30 July 2012 is shown in Figure 1. Typhoon Saola was the ninth named storm and the fourth typhoon of the 2012 Pacific typhoon season. Typhoon Saola strengthened from an intensity of 35 kn (18 m s$^{-1}$) to 57.5 kn (29.6 m s$^{-1}$) between 0500 LST on 29 July and 1700 LST on 30 July. The IR images show that the areal extent (as a radius) of very cold clouds decreased from 500 to 300 km during the day (between 0500 and 1700 LST) on 29 July and on 30 July, and the areal extent of relatively warm clouds increased. During the night, from 1700 LST on 29 July to 0500 LST on 30 July, the areal extent of very cold clouds increased rapidly from 300 to 500 km and the areal extent of the warmer clouds decreased correspondingly. Diurnal variations in the areal extent of very cold clouds in Typhoon Saola were particularly evident in the southern half of the typhoon. The changes in the IR BTs associated with changes in the areal extent of the clouds between 0500 and 1700 LST and between 1700 and 0500 LST were as high as 50–70 $^\circ$C. Maximum cooling did not occur in a circle within the TC as observed by Dunion et al. (2014). Typhoon Saola is a clear example of different diurnal variations occurring under two different types of cloud.

The temporal evolutions of the areal extents of clouds and the IR BTs during Typhoon Saola between 1700 LST on 28 July and 1700 LST on 30 July are shown in Figure 2. The areal extent was calculated from the total number of 4 × 4 km$^2$ pixels within the temperature range of interest. Most areas within a 200 km radius...
of the storm centre were covered with clouds colder than 240 K. The area covered by clouds with very cold (less than 208 K) tops reached a maximum in the early morning (0300–0500 LST) and then decreased after sunrise. The decrease in very cold cloud cover after sunrise was followed by an increase in the area covered by clouds with tops between 208 and 240 K. The average temperature of the cloud tops <208 K changed generally in phase with (but 3 h in advance of) the average temperature of the cloud tops between 208 and 240 K. The area mean temperature 200 km from the TC centre had a diurnal oscillation, with a minimum temperature in the morning and a maximum temperature at 1100 LST on 29 July. At 300–500 km from the centre, Typhoon Saola was covered with clouds colder than 240 K and clear sky or clouds with IR BT > 240 K. The area covered by cloud tops colder than 240 K reached a maximum areal extent in the afternoon (1400–1700 LST) and a minimum areal extent between midnight and early morning (2300–0500 LST). Each particular area fluctuated between being warmer than 240 K and covered with cloud tops colder than 240 K. A decrease in the area covered by clouds <240 K was therefore followed by an increase in the area covered by clouds >240 K, and vice versa. The out-of-phase relationship between area and cloud-top temperature for the two sets of conditions led to the average temperature 300–500 km from the centre having two peaks per day.

The radius–time plots of the 14-year mean IR BTs for weak and strong storms in the western North Pacific are shown in Figure 3, and were used to determine whether the semi-diurnal cycle in Typhoon Saola was either unique to that TC or a common feature of area-averaged TC IR BTs. Azimuthal IR BT calculations have been used in previous studies (e.g. Steranka et al., 1984; Kossin, 2002; Dunion et al., 2014) to analyse diurnal variations in TCs. At any particular LST, the mean IR BT for both weak and strong storms, except at 50–100 km (radius) from the centres of strong storms, increased as the radial distance from the TC centre increased. The IR BT for strong storms was lower at 50–100 km than at 50 km from the TC centre. The IR BT 50–200 km from the TC centre reached a minimum in the early morning (0300–0600 LST) in both weak and strong storms. The minimum IR BT at 300–500 km from the TC centre (an area mostly covered with mid-level clouds, low-level clouds, and clear sky) was in the late afternoon (1500 LST). Two minima, one in the early morning and one in the late afternoon, were found in the IR BT 200–300 km from the TC centre.

The semi-diurnal cycle in the radius-averaged IR BT could have had two causes, one being the out-of-phase relationship between the diurnal variations in the IR BT under two different

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**Figure 2.** Three-hourly GOES data for (a) areal extent, (b) brightness temperature, and (c) radius-averaged brightness temperature at 200 km from the storm centre. The black lines show IR brightness temperatures <208 K and the blue dashed lines show IR brightness temperatures between 208 and 240 K. Three-hourly GOES IR data for (d) areal extent, (e) brightness temperature, and (f) radius-averaged brightness temperature at 300–500 km from the storm centre. The blue dashed lines show IR brightness temperatures >240 K and the black lines show IR brightness temperatures <240 K.

**Figure 3.** Radius–time plots of the 14-year mean brightness temperatures (K) for (a) weak and (b) strong storms.
IR conditions (BT < 240 K or BT > 240 K) and the other being the out-of-phase relationship between the time at which maximum cloud cover occurred under two different cloud conditions (BT < 208 K or 208 K < BT < 240 K). The mean IR BT was about 240 K 200–300 km from the centres of weak storms. It is likely that the semi-diurnal cycle was mainly caused by the out-of-phase relationship between the diurnal variations in the IR BTs < 240 K and IR BTs > 240 K. TC conditions with IR BTs < 240 K had minimum mean IR BTs in the morning, whereas TC conditions with IR BTs > 240 K had minimum mean IR BTs in the afternoon, as for Typhoon Saola. The mean IR BT was about 220–230 K 200–300 km from the centres of strong storms. The semi-diurnal cycle in this radius range was mainly caused by the out-of-phase relationship between the time at which maximum cloud cover occurred with IR BTs < 208 K and IR BTs of 208–240 K. Clouds with IR BTs < 208 K and IR BTs of 208–240 K both had minimum mean temperatures in the morning, but clouds with IR BTs < 208 K reached maximum mean coverage in the morning, whereas clouds with IR BTs of 208–240 K reached maximum mean coverage in the afternoon.

The minimum mean IR BT 300–500 km from the TC centre occurred in the afternoon, and could have been caused by the dominance of clouds with IR BTs > 240 K (which reached a minimum temperature in the afternoon) or by more cold clouds occurring in the afternoon than at other times. Cold clouds are more likely to reach 300–500 km from the centre in strong than in weak storms, so the minimum mean values of IR BT in the afternoon during strong storms were more likely to have been caused by cold clouds reaching 300–500 km from the TC centre in the afternoon, whereas the minimum mean values of IR BT in the afternoon during weak storms were more likely to have been caused by cloud-tops with IR BTs > 240 K themselves having minimum temperatures in the afternoon. The minimum mean IR BT found 50–200 km from the TC centre in the early morning and the minimum mean IR BT found 300–500 km from the TC centre in the late afternoon during strong storms (Figure 3) were consistent with the propagating diurnal pulse observed by Dunion et al. (2014).

The data shown in Figure 3 suggest that TC convective systems may be better described in terms of their areas and temperatures rather than their radius-averaged temperatures. The 14-year mean area of the IR BT in each 5 K bin is shown as a function of the

| BT (K) | 0000 | 0300 | 0600 | 0900 | 1200 | 1500 | 1800 | 2100 |
|--------|------|------|------|------|------|------|------|------|
| 180    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 190    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 200    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 210    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 220    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 230    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 240    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 250    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 260    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 270    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |
| 280    | 0.8  | 1.6  | 2.4  | 3.2  | 4.0  | 4.8  | 5.6  | 10.0 |

Figure 4. Temperature–time plots of the 14-year mean cloud coverage at particular brightness temperatures (within each 5 K interval) within 500 km of the storm centre for (a) weak and (b) strong storms. The dots indicate the times at which peak coverage occurred for clouds in the different 5 K temperature intervals.
which maximum cloud coverage occurred over the oceans was sensitive to the temperature thresholds used, for both weak and strong storms. It also indicates that the discrepancies between the results of previous studies using different satellite observations to track diurnal cycles in deep convection and cloud patterns in TCs could have been caused by different temperature thresholds.

The time at which maximum cold cloud coverage occurs has also been found to vary substantially depending on the IR BT thresholds used over tropical oceans (e.g. Janowiak et al., 1994; Chen and Houze, 1997; Yang and Slingo, 2001; Tian et al., 2004). Over tropical land, the time at which maximum cold cloud coverage occurs has been found to be independent of the temperature thresholds used (Janowiak et al., 1994; Hong et al., 2006). Insufficient data are available to determine whether the time at which maximum TC cloud coverage over land reached is sensitive to the temperature thresholds used (as is the case over the oceans). Therefore, the diurnal cycles in the total area covered by cloud tops colder than 208 K and cloud tops between 208 and 240 K over tropical oceans and land are examined using one year of data in Figure 6. Similar to the case for TC clouds, the maximum occurrence of very cold deep convective clouds was found to occur at 0400–0700 LST, and the maximum occurrence of cold high clouds was found to occur at 1600 LST over tropical oceans. Very deep cold convective clouds and cold high clouds were found to reach maxima at 1800–1900 LST over tropical land. We therefore inferred that the time at which maximum TC clouds occur over land will not be sensitive to the temperature thresholds used as deep convective clouds over tropical land.

The 14-year mean diurnal cycles in the total areas covered by cloud tops colder than 208 K and by cloud tops between 208 and 240 K, together with their respective mean temperatures, are shown in Figure 7. We used these results to determine whether the coverage of cloud tops colder than 208 K and cloud tops between 208 and 240 K were related or developed independently. In both weak and strong storms, the mean coverage of very cold cloud tops reached a maximum at 0600 LST, decreased after sunrise, and reached a minimum in the late afternoon (1500–1800 LST). The maximum area of very cold cloud tops in the morning suggests that very cold clouds, with IR BTs < 208 K, followed the cloud–radiation interaction hypothesis. The coverage of cloud tops between 208 and 240 K reached a maximum at 1500 LST, when the atmospheric surface layer overlying the ocean surface was at its warmest. The maximum

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cloud-top coverage decreased after 1500 LST, then decreased rapidly after sunset (1800 LST). The mean coverage of very cold cloud tops reached a maximum in the early morning (0000–0300 LST), then decreased after sunrise. The decrease in the coverage of very cold clouds was followed by an increase in coverage between 208 and 240 K. It is possible that clouds with IR BTs between 208 and 240 K evolved from very cold clouds after the very cold clouds had reached their maximum coverage. Further modelling experiments are needed to determine whether different mechanisms are involved in the diurnal cycles of these two types of cloud. In both weak and strong storms, the mean coverage of very cold clouds was about one third that of clouds with IR BTs between 208 and 240 K.

Very cold clouds reached minimum mean temperatures at midnight or before dawn for both weak and strong storms. The diurnal temperature variations were out of phase with the diurnal variations in temperature and coverage (percentage to 500 km from the storm centre) for weak storms. In both weak and strong storms the temperature of the clouds increased. The diurnal temperature variations were generally in phase with the areal extent of clouds occurred. The maximum and minimum temperatures of the cold clouds between 208 and 240 K were about 3 h later than the maximum and minimum temperatures of the very cold clouds.

4. Discussion

4.1. The discrepancies between diurnal pulses in IR BTs and precipitation in TCs

Dunion et al. (2014) examined all major North Atlantic hurricanes between 2001 and 2010 and found that TC diurnal pulses are a distinguishing characteristic of the TC diurnal cycle. The diurnal pulses (peak cooling in the IR field) reached 200 km from the TC centre at 0400–0800 LST, 300 km from the TC centre at 0800–1200 LST, 400 km from the TC centre at 1200–1500 LST, and 500 km from the TC centre at 1500–1800 LST. Wu et al. (2015) analysed satellite precipitation data from 1998 to 2012 and found a similar outward propagation of diurnal signals. The diurnal amplitude of precipitation decreased as the radial distance from the TC centre increased, and the timing of the peak was progressively later. In weak storms, precipitation peaked 2.5–4 h earlier in the inner core (within 100 km of the centre) than in the outer rain bands (100–500 km from the centre), whereas in strong storms the lead times were 2.5–5.5 h. The lead times for the TC precipitation peaks in the inner cores relative to the outer rain bands were different in different basins and between storms of different intensities. In any case, the lead times were several hours earlier for precipitation than for peak cooling in the IR field.

Apart from the differences in the diurnal phases, the locations of largest diurnal amplitudes were also different for the IR BTs and precipitation. Diurnal variations in major North Atlantic hurricanes in the IR field are strongest 300–500 km from the TC centres (Dunion et al., 2014), but stronger diurnal variations in TC precipitation are found in the inner core regions than in the outer rain bands (Wu et al., 2015). Although diurnal cycles in strong North Atlantic storms are exceptions (no significant diurnal cycles have been detected in inner core regions), the diurnal amplitude of precipitation 100–400 km from a TC centre has been found to decrease as the radial distance from the TC centre increases.

The discrepancies between diurnal variations in IR BTs and precipitation in TCs can be explained by the IR cooling detected by Dunion et al. (2014) being largely induced by the differences between the two IR cloud conditions at a particular distance from the TC centre, which reflects changes in clouds in terms of both temperature and areal extent. For instance, the cloud
field for Hurricane Felix cooled by as much as 40–85°C 200–300 km from the TC centre between 1215 and 1815 UTC on 3 September 2007 (Fig. 1 in Dunion et al. (2014)). This was achieved through very cold clouds extending from 200 to 300 km from the TC centre. Peak IR cooling occurring in the afternoon 400–500 km from the TC centre is likely because non-precipitating 208 K < IR BT < 240 K clouds (which reached maximum coverage in the afternoon) extended to 400–500 km from the TC centre. However, including large non-precipitating areas does not change the diurnal variation characteristics in precipitation significantly (Wu et al., 2015). The discrepancies between diurnal variations in the TC IR BTs and precipitation also suggests that diurnal cycles in a TC cannot be adequately described only in terms of IR BT changes at certain distances from the TC centre. Instead, TC diurnal cycles are better described in terms of both the temperature (representing different cloud conditions) and the time at which maximum cloud cover occurs.

4.2. Mechanisms involved in two types of cloud

Our results suggest that the maximum occurrence of very cold clouds with IR BTs < 208 K in the morning can be explained using hypotheses based on cloud–radiation interactions. The maximum occurrence of clouds in the afternoon suggests that diurnal variations in the cloud tops between 208 and 240 K follow the diurnal solar heating of the ocean surface and the atmospheric boundary layer, as suggested by Chen and Houze (1997). Chen and Houze (1997) suggested that diurnal variations in the sea-surface temperature are instrumental in oceanic diurnal cycles, and that diurnal heating of the ocean surface during the day controls the time at which convective systems start in the afternoon. Tian et al. (2004) suggested that the lack of a diurnal cycle in the sea-surface temperature may limit the ability of boundary forcing in atmospheric models to simulate both the diurnal phase and amplitude of convection and cloud cover over the oceans.

The maximum cloud cover in the afternoon could also be related to the presence of cirrus clouds. Cirrus clouds, which are strongly connected to tropical deep convective clouds, can extend and persist for some hours after deep convective clouds dissipate (Gray and Jacobson, 1977). Cirrus clouds can explain the phase difference between IR BTs < 208 K and 208 K < IR BTs < 240 K over the oceans, but cannot explain the in-phase relationship between IR BTs < 208 K and 208 K < IR BTs < 240 K over land. Li and Wang (2012) considered the quasi-diurnal behaviour of outer spiral rain bands associated with the boundary layer recovery from the effect of convective downdraughts and the consumption of convective available potential energy by convection in the previous outer spiral rain bands. The boundary-layer air near the original location of convection initiation takes about 10 h to recover after extracting energy from the underlying ocean. However, this mechanism is unable to explain the timing of the maximum precipitation. In addition, Li and Wang (2012) specifically explained the periodic behaviour in the outer spiral rain bands (three times the radius of maximum wind). However, the diurnal cycle of TC convection is not unique to the outer spiral bands.

The maximum occurrence of cold clouds between 208 and 240 K in the afternoon over both tropical oceans and land (Figure 6) appears to support the idea that diurnal variations in cold clouds between 208 and 240 K are influenced by diurnal solar heating of the surface, but such a conclusion is not possible from our IR analyses. One conclusion that can be drawn is that two different mechanisms are involved in diurnal variations in very cold deep convective clouds and cold high clouds over the oceans, taking into account the out-of-phase relationships between the times at which maximum very cold convective cloud and cold high cloud coverage occur.

5. Concluding remarks

Diurnal variations of the areas and the cloud-top temperatures in deep convective cloud systems in TCs over the western North Pacific were analysed using pixel-resolution IR BT data and best-track data for 2000–2013, which included a total of 391 storms. Diurnal variations in the areas and cloud-top temperatures of very cold deep convection cloud tops (BT < 208 K) and cold high cloud tops (208 K < BT < 240 K) were considered so that diurnal variations in TC convective systems could be described as precisely as possible. The mean area covered by very cold cloud tops reached a maximum in the early morning (0300–0600 LST), and the mean area covered by cloud tops between 208 and 240 K reached a maximum in the afternoon (1500–1800 LST). The out-of-phase relationship between the areal extents under these different cloud types led to substantial variations in the time at which the maximum area of cold clouds occurred, depending on the IR BT thresholds used. TC conditions with IR BTs < 240 K had minimum mean IR BTs in the morning (0300–0600 LST), and TC conditions with IR BTs > 240 K had minimum mean IR BTs in the afternoon (1500–1800 LST). The out-of-phase relationship between cloud-top temperatures <240 K and IR cloud-top temperatures >240 K, and between the maximum times of coverage of clouds with cloud-top temperatures <208 K and of cold clouds between 208 and 240 K could both lead to two daily minima in the radius-averaged IR temperature. The diurnal cycles in TC convective cloud systems are complicated by diurnal variations in the horizontal sizes of clouds and by cloud temperatures having different phases under different cloud conditions. The differences between diurnal cycles in deep convection and cloud patterns in TCs found in previous studies are largely caused by the use of different temperature thresholds to represent deep convection or by the use of averages for different cloud conditions. The diurnal variations of the areas and the cloud-top temperatures analysed in this article not only provided an explanation for the semi-diurnal cycle in TC clouds, but also explained the discrepancies between diurnal pulses in the TC IR BTs (Dunion et al., 2014) and precipitation (Wu et al., 2015). It is worth note that the diurnal variations of the areas and temperatures in TC clouds found in this article are not unique to the western North Pacific Ocean, but are common features in all TC basins.

Hypotheses for cloud–radiation interactions have been developed to explain daytime minima and night-time maxima in cloud cover. Our results suggest that cloud–radiation interactions can only partly explain diurnal variations in deep convection in TCs over oceans. It appears that the maximum occurrence of very cold clouds with IR BTs < 208 K in the morning can be explained using hypotheses based on cloud–radiation interactions, whereas the maximum occurrence of clouds between 208 and 240 K in the afternoon need to be explained using hypotheses that include different physical mechanisms. Modelling experiments are needed to determine whether afternoon maximum occurrences of cold high clouds between 208 and 240 K are influenced by diurnal solar heating of the ocean surface and atmospheric boundary layer, as suggested by Chen and Houze (1997), by cold cirrus clouds generated by deep convective clouds, or by the boundary-layer recovery process proposed by Li and Wang (2012). Very cold clouds are closely associated with precipitating deep convective clouds, and precipitation in TCs is closely related to the release of latent heat and the development of the TC (e.g. Steranka et al., 1986; Rao and MacArthur, 1994; Kieper and Jiang, 2012), so diurnal variations under different cloud conditions could have important influences on the structure and intensity of TCs.

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References

Bowman KP, Fowler MD. 2015. The diurnal cycle of precipitation in tropical cyclones. J. Clim. 28: 5325–5334.

Browner SP, Woodley WL, Griffith CG. 1977. Diurnal oscillation of area of cloudiness associated with tropical storms. Mon. Weather Rev. 105: 856–864.

Chen S, Houze RA Jr. 1997. Diurnal variation and life-cycle of deep convective systems over the tropical Pacific warm pool. Q. J. R. Meteorol. Soc. 123: 357–388.

Dunion JP, Thorncroft CD, Velden CS. 2014. The tropical cyclone diurnal cycle of mature hurricanes. Mon. Weather Rev. 142: 3905–3919.

Hong G, Heygster G, Rodriguez CAM. 2006. Effect of cirrus clouds on the diurnal cycle of tropical deep convective clouds. J. Geophys. Res. 111: D06209, doi: 10.1029/2005JD006208.

Janowiak JE, Arkin PA, Morrissey M. 1994. An examination of the diurnal cycle in oceanic tropical rainfall using satellite and in situ data. Mon. Weather Rev. 122: 2296–2311.

Janowiak JE, Joyce RJ, Yanosh Y. 2001. A real-time global half-hourly pixel-resolution infrared dataset and its applications. Bull. Am. Meteorol. Soc. 82: 205–217.

Jiang H, Zipser EJ. 2010. Contribution of tropical cyclones to the global precipitation from eight seasons of TRMM data: Regional, seasonal, and interannual variations. J. Clim. 23: 1526–1543.

Joyce JR, Janowiak JE, Huffman G. 2001. Latiitudinally and seasonally dependent zenith-angle corrections for geostationary satellite IR brightness temperatures. J. Appl. Meteorol. 40: 689–703.

Kieper ME, Jiang H. 2012. Predicting tropical cyclone rapid intensification using TRMM-WETMARC/LBA. J. Geophys. Res. 107: 8064, doi: 10.1029/2000JD000338.

Kossin JP. 2002. Daily hurricane variability inferred from GOES infrared imagery. Mon. Weather Rev. 130: 2260–2270.

Lau K-M, Zhou YP, Wu H-T. 2008. Have tropical cyclones been feeding more extreme rainfall? J. Geophys. Res. 113: D23113, doi: 10.1029/2008JD009963.

Liu GS, Curry JA, Sheu R-S. 1995. Classification of clouds over the western equatorial Pacific Ocean using combined infrared and microwave satellite data. J. Geophys. Res. 100: 13811–13826.

Machado LAT, Laurent H, Lima AA. 2002. Diurnal march of the convection observed during TRMM-WETAMC/LBA. J. Geophys. Res. 107: 8064, doi: 10.1029/2000JD000338.

Melhauser C, Zhang F. 2014. Diurnal radiation cycle impact on the pregenesis environment of Hurricane Karl (2010). J. Atmos. Sci. 71: 1241–1259.

Muramatsu T. 1983. Diurnal variations of satellite-measured TBB areal distribution and eye diameter of mature typhoons. J. Meteorol. Soc. Jpn. 61: 77–89.

Nesbitt SW, Zipser EJ. 2003. The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements. J. Clim. 16: 1456–1475, doi: 10.1175/1520-0442-16.10.1456.

Prat OP, Nelson BR. 2013. Precipitation contribution of tropical cyclones in the southeastern United States from 1998 to 2009 using TRMM satellite data. J. Clim. 26: 1047–1062.

Randall DA, Harshvardhan, Duzlich DA. 1991. Diurnal variability of the hydrologic cycle in a general circulation model. J. Atmos. Sci. 48: 60–62.

Rao GV, MacArthur PD. 1994. The SM/I estimated rainfall amounts of tropical cyclones and their potential in predicting the cyclone intensity changes. Mon. Weather Rev. 122: 1568–1574.

Shu HL, Zhang QH, Xu B. 2013. Diurnal variation of tropical cyclone rainfall in the western North Pacific in 2008–2010. Atmos. Oceanic Sci. Lett. 6: 103–108.

Steranka J, Rodgers EB, Gentry RC. 1984. The diurnal variation of Atlantic Ocean tropical cloud distribution inferred from geostationary satellite infrared measurements. Mon. Weather Rev. 112: 2338–2344.

Su C-H, Lau K-M, Takayabu YN, Short DA. 1997. Diurnal variations in tropical oceanic cumulus convection during TOGA COARE. J. Atmos. Sci. 54: 639–655.

Tian B, Soden BJ, Wu X. 2004. Diurnal cycle of convection, clouds, and water vapor in the tropical upper troposphere: Satellites versus a general circulation model. J. Geophys. Res. 109: D11101, doi: 10.1029/2003JD004117.

Wang Y, Druyan I, Gentry RC. 1986. The relationship between satellite measured convective bursts and tropical cyclone intensification. Mon. Weather Rev. 114: 1539–1546.

Wu QY, Ruan Z, Chen D, Lian T. 2015. Diurnal variations of tropical cyclone precipitation in the inner and outer rainbands. J. Geophys. Res. 120: 1–11, doi: 10.1002/2014JD022190.

Yang GY, Slingo J. 2001. The diurnal cycle in the Tropics. Mon. Weather Rev. 129: 784–801.

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