Monthly variations of tropical cyclone rapid intensification ratio in the western North Pacific

Xuyang Ge | Donglei Shi | Liang Guan

In this study, the monthly cycle of tropical cyclone (TC) rapid intensification (RI) ratio and its climate controlling factors are investigated. The TC RI ratio is greatest in the late fall season, although both total TC frequency and RI samples are largest in the peak summer season. The environmental conditions are examined to identify the possible controlling factors, including the mean TC locations, the ambient relative vorticity, and the vertical profiles of atmospheric and ocean temperatures. Consistent with previous studies, the lower latitude of TC location and pronounced ambient cyclonic vorticity favor TC RI in the late fall. Moreover, the result suggests that the thermodynamic condition contributes a greater RI ratio. During the late fall season, the outflow layer temperature is much lower, indicating a greater thermodynamic efficacy. Meanwhile, the subsurface ocean condition (i.e., a deeper mixed layer and stronger subsurface thermal stratification) promotes greater RI ratio in October and November.

KEYWORDS
environmental factors, monthly variations, rapid intensification, tropical cyclone

1 | INTRODUCTION

Tropical cyclone (TC) rapid intensification (RI) is a vital process for super typhoon (STY), since the majority of these storms over the western North Pacific (WNP) experience at least one RI process during their life cycles (Ventham & Wang, 2007). Moreover, TC RI is one of the most difficulties in tropical weather forecasting. This is primarily ascribable to the inadequate understanding of internal dynamics and the interaction with the large-scale environment (Bosart, Velden, Bracken, Molinari, & Black, 2000; Ge, 2015; Ge, Guan, & Zhou, 2016; Molinari & Vollaro, 2014; Reasor, Rogers, & Lorsolo, 2013; Shay, Goni, & Black, 2000; Wang & Wu, 2004; Wang & Zhou, 2008). During the RI period, TCs are affected by multi-scale processes that govern the inner-core structure and interactions between the storm and environmental conditions. For instance, the processes include TC interaction with subsurface oceanic thermal conditions (Mei, Xie, Primeau, McWilliams, & Pasquero, 2015; Shay et al., 2000; Wang, Wang, Weisberg, & Black, 2017; Wang, Wang, Zhang, & Wang, 2015), responses to vertical wind shears (VWSs; Ge, Li, & Peng, 2013; Reasor et al., 2013), concentric eyewall replacement cycles (Willoughby, Clos, & Shoreibah, 1982; Kossin & Sitkowski, 2009; Ge et al., 2016), modifications in upper-level outflow (Molinari & Vollaro, 2014), global warming (Sun et al., 2017), and many others.

Recent studies (Gilford, Solomon, & Emanuel, 2017) investigated the possible factors in controlling the monthly variations of TC maximum potential intensity (MPI), and found that the monthly variation of the MPI is a function of environmental conditions. Specifically, the controlling factors reflect the thermodynamic atmosphere–ocean disequilibrium and TC thermodynamic efficacy. As will be shown in Section 3, the ratio of RI samples to the total TC number exhibits a great monthly variation as well. This stimulates us to explore the underlying mechanisms. Hence, the primary objective is to examine the monthly cycle of RI
ratio in the WNP, and to identify the possible contributing factors.

The rest of this paper is organized as the follows: In Section 2, the definition of RI and the dataset are briefly introduced. In Section 3, the monthly variations of RI ratio are presented, and possible environmental conditions are examined in Section 4. Finally, a short summary is given in Section 5.

2 | DEFINITION OF RI AND DATASETS

In this study, the definition of RI follows Wang and Zhou (2008). That is, for each RI sample, the increase of the maximum wind speed satisfies the following criteria: (a) 5 knots in the first 6 hr, (b) 10 knots in the first 12 hr and (c) 30 knots in 24 hr. The criteria (c) is most essential for defining a RI process, while the criteria (a) and (b) are used to precisely identify the onset timing of RI. The RI samples during 1965–2015 are identified. Furthermore, the monthly mean locations for the RI samples are calculated. The comparisons of the atmospheric and oceanic conditions are conducted in the vicinity of these monthly mean locations.

The 6-hourly TC best-track data is obtained from the Joint Typhoon Warning Center (JTWC). The wind fields and temperature are from the National Centers for Environmental Prediction (NCEP) FNL (Final) Operational Global Analysis data, and oceanic data (i.e., sea water temperature and salinity) are from the Global Ocean Data Assimilation System (GOADS). The temperature from Radio Occultation (RO) observations aboard the satellites of COSMIC during 2007–2015 is also used. This global satellite product has a finer vertical resolution, which provides an opportunity to study the outflow layer temperature.

3 | MONTHLY VARIATION OF RI RATIO

Figure 1 displays the monthly distribution of total TC numbers, RI samples, and the RI ratio over WNP, respectively. In the WNP basin, TCs mainly occur during the tropical season (i.e., from July to November). Therefore, we mainly focus on this period in this study. For the monthly cycle of TC occurrence, its peak appears in August. The majority of the RI samples occur during July to November, with the peaks in September and October. This is consistent with Wang and Zhou (2008). However, the ratio of RI sample shows a distinct monthly cycle. Specifically, the maximum ratio occurs in October (30%), whereas a minimum appears in August (17%). This indicates that, in August, the total number of TCs is the greatest, but the RI ratio is the lowest. On the contrary, the TC number is the lowest in late fall season, whereas the RI ratio is the highest. This result also agrees with previous results (Holliday & Thompson, 1979; Wang & Zhou, 2008). Provided that different evolution features of TC number and RI ratio, a question here is what is responsible for such discrepancies. TC RI is largely modulated by both atmospheric and oceanic conditions. Wang and Zhou (2008) suggested that three climate factors likely contribute to the monthly variations of the ratio of RI, including: (a) TC formation location, (b) the mean low-level meridional shear vorticity and (c) VWS. In short, they mainly focused on the large-scale dynamic impacts, but the thermodynamic factors are not explored. To this end, we will further examine the atmospheric and oceanic conditions.

4 | CONTROLLING FACTORS

1. Latitude impact

Figure 2 displays the monthly mean location of TC genesis and the onset of RIs. In general, the genesis regions show similar monthly migrations, but shift southeastward compared to the RI regions. Obviously, the mean location of RI samples in the late fall season (October and November) is located southward compared with the peak summer season. Specifically, the mean latitude in the late fall is about 12°N, which is about 6° southward shift than its counterpart (July–August). This is attributable to the monthly migration of the monsoon trough. It is realized that TC intensification rate largely depends on the latitude effect (Demaria & Pickle, 1988; Li et al., 2012; Smith, Kilroy, & Montgomery, 2015). Physically, lower the latitude, smaller is the Coriolis parameter and thus is the inertial stability. A smaller inertial stability favors a rapid establishment of boundary inflow layer, leading to a faster vortex spin-up (Li et al., 2012; Smith et al., 2015). By this reasoning, the latitude effect partly accounts for the monthly cycle of TC RI ratio.
2. Low-level atmospheric vorticity

The atmospheric conditions are important factors in influencing TC formation and intensification. Both the low level relative vorticity and VWS are believed to have a great impact on the TC intensification. In the current study, the VWS is obtained by calculating the differences between 200 and 850 hPa wind fields averaged within a radius of 500 km centered at the mean region of RI onset. It is found that the monthly variations of VWSs are very limited (Figure 3), and thus contributes a little to the monthly variations of RI ratio. Next, the latitudinal distribution of relative vorticity averaged between 120 and 160°E is examined in Figure 4. It is worth mentioning that the origin represents the mean location of onset of RI samples in each calendar month. Apparently, there exists significant cyclonic vorticity around the mean location. Notice that the ambient relative vorticity is the greatest in November, whereas it is the smallest in July and August. Close examinations show that the zonal wind associated with the monsoon trough has a maximum (minimum) cyclonic shear in November (August). This large-scale dynamic condition favors a greater RI ratio in late fall season as well.

3. Tropospheric temperature profile

TC MPI and intensification is closely associated with the ambient temperature and moisture profile (Bister & Emanuel, 1998; Emanuel, 1989; Ge et al., 2016; Gilford et al., 2017; Stovern & Ritchie, 2016; Wang, Camargo, Sobel, & Polvani, 2014). As shown in Figure 3, the greatest MPI occurs in September, indicating that the evolution of MPI is different from the monthly variations of the RI ratio. The thermodynamic efficiency of the TC MPI is mainly determined by sea surface temperature (SST) and the outflow layer temperature. The seasonal cycles of SST in the WNP shows a single annual maxima and a single annual
minima, which coincide with boreal summer or winter. Nevertheless, the outflow layer temperature influenced by planetary wave-forcing from the lower stratosphere plays an important role in the monthly cycle of TC MPI. Namely, tropical lower stratospheric temperature exhibits a strong monthly cycle, minimizing in the boreal winter and maximizing in the boreal summer. As a result, TC MPI maxima shift from peak summer season to September in WNP. Given the importance of the outflow layer temperature, we compare the vertical profiles of atmospheric temperature between the boreal summer and late fall seasons. In each calendar month, the temperature profiles are obtained by averaging within a radius of 500 km centered at the mean location. For simplicity, the temperatures in July and August (named as JA) are averaged to represent the thermodynamic condition for the peak summer season, and the mean temperature in October and November (named as ON) signifies the late fall season. To minimize the uncertainties in the NCEP/FNL reanalysis dataset, the RO/COSMIC dataset is also used. Figure 5 compares the vertical profiles of temperature in JA and ON, respectively. Of a particular interest is that marked differences occur in the lower stratospheric temperature between JA and ON. The averaged outflow layer temperature is nearly 3–4 K warmer in JA than that in ON. Further examination shows that there are no significant differences in SST around the mean RI location between JA and ON. As such, the TC MPI thermodynamic efficacy is higher in ON, and thus a potential for rapid intensification. Previous idealized numerical simulations (Wang et al., 2014; Stovern & Ritchie, 2016) demonstrated that a lower outflow temperature is apt to a RI and thus a stronger TC.

4. Subsurface oceanic condition

The changes in surface and subsurface ocean condition influence TC intensification (e.g., Price, 1981; Shay et al., 2000; Vincent, Emanuel, Lengaigne, Vialard, & Madec, 2014). For instance, the oceanic cooling effect leads to a reduction of air-sea flux, by which inhibits TC intensification. With this regard, we compare the oceanic conditions during the period of interest. Figure 6 compares the vertical profiles of the climate subsurface ocean temperature and salinity between JA and ON. The SST is insignificantly warmer in JA than that in ON. Nevertheless, the vertical water temperature gradient within the subsurface layer (i.e., up to the depth of 50 m) is smaller in ON. In other words, during the late fall season, the vertical profile of subsurface ocean water temperature is less sharpened within the mean RI region. The sharper the subsurface ocean temperature gradient, the stronger is the cooling effect. Mei et al. (2015) proposed that the seasonal mean lifetime peak intensity of WNP TCs are controlled by the upper ocean temperatures, which is in agreement with our findings. Furthermore, a stronger salinity gradient at the subsurface
(above the depth of 90 m) is obvious in ON. This implies a greater vertical stratification in the subsurface ocean, which is unfavorable for upwelling. Hence, both a deeper mixed layer and stronger subsurface thermal stratification contributes to a greater RI ratio in October and November.

The oceanic cooling effect is a function of TC attributes (i.e., intensity, size, and propagation speed). Under similar oceanic conditions, the slower the TC movement and the larger the TC size, the stronger is the oceanic cooling effect. By this reasoning, the monthly variations of mean TC movement speeds are compared. There are no significant differences in the monthly propagation speeds (not shown). In short, the variation of thermodynamic stratification partially contributes to the monthly variations of RI ratio in the WNP.

5 | SUMMARY

In this study, the monthly variations of TC RI ratio in the WNP is examined. It is revealed that the ratio of RI samples is greater in the late fall season (October–November), although the total number is largest during the peak summer season. This suggests that there are probably different environmental conditions controlling the TC genesis and RI process. The monthly cycle of RI ratio is likely determined based on both atmospheric and subsurface ocean conditions, including the mean TC locations, the ambient relative vorticity, and both atmospheric and oceanic temperature profiles. During the late fall season, the outflow layer temperature is lower, indicating a greater thermodynamic efficacy. Meanwhile, the subsurface ocean conditions (i.e., a deeper mixed layer and stronger subsurface thermal stratification) denotes a weaker cooling effect.

Although the possible climate factors in controlling the monthly variations of RI ratio are proposed, the relative role of these factors are not clear. Furthermore, the long-term (i.e., inter-annual and inter-decadal) variability of TC RI ratio is not investigated yet. Recent studies (Sun et al., 2017; Wang et al., 2015, 2017) pointed out that the large-scale climate modes, such as El Niño-Southern Oscillation (ENSO), Pacific-North America pattern (PNA) and North Atlantic Oscillation (NAO), have impacts on the RI via the modulation on the large-scale dynamic and thermodynamic environment. These issues await further study.

ACKNOWLEDGEMENTS

This work is jointly sponsored by the National Science Foundation of China (grant nos. 41575056, 41730961, and 41775058) and Key Basic Research Program of China (grant no. 2015CB452803).

REFERENCES

Bister, M., & Emanuel, K. A. (1998). Dissipative heating and hurricane intensity. Meteorology and Atmospheric Physics, 65, 233–240.

Bosart, L. F., Velden, C. S., Bracken, W. E., Molinari, J., & Black, P. G. (2000). Environmental influences on the rapid intensification of hurricane Opal (1995) over the Gulf of Mexico. Monthly Weather Review, 128, 322–352.

Demaria, M., & Pickle, J. D. (1988). A simplified system of equations for simulation of tropical cyclones. Journal of the Atmospheric Sciences, 45, 1542–1554.

Emanuel, K. A. (1989). The finite-amplitude nature of tropical cyclogenesis. Journal of the Atmospheric Sciences, 46, 3431–3456.

Ge, X. (2015). Impacts of environmental humidity on concentric eyewall structure. Atmospheric Science Letters, 16, 273–278.

Ge, X., Guan, L., & Zhou, S. (2016). Impacts of initial structure of tropical cyclone on secondary eyewall formation. Atmospheric Science Letters, 17, 569–574.

Ge, X., Li, T., & Peng, M. (2013). Effects of vertical shears and midlevel dry air on tropical cyclone developments. Journal of the Atmospheric Sciences, 70, 3859–3875.

Gilford, D. M., Solomon, S., & Emanuel, K. A. (2017). On the seasonal cycles of tropical cyclone potential intensity. Journal of Climate, 30, 6085–6096.

Holliday, C. R., & Thompson, A. H. (1979). Climatological characteristics of rapidly intensifying typhoons. Monthly Weather Review, 107, 1022–1034.

Kossin, J. P., & Sitkowski, M. (2009). An objective model for identifying secondary eyewall formation in hurricanes. Monthly Weather Review, 137, 876–892.

Li, T., Ge, X., Peng, M., & Wang, W. (2012). Dependence of tropical cyclone intensification on the Coriolis parameter. Tropical Cyclone Research and Review, 1, 242–253.

Mei, W., Xie, S. P., Primeau, F., McWilliams, J. C., & Pasquero, C. (2015). Northwestern Pacific typhoon intensity controlled by changes in ocean temperatures. Science Advances, 1(4), e1500014. https://doi.org/10.1126/sciadv.1500014.

Molinari, J., & Vollaro, D. (2014). Symmetric instability in the outflow layer of a major hurricane. Journal of the Atmospheric Sciences, 71, 3739–3746.

Price, J. F. (1981). Upper ocean response to a hurricane. Journal of Physical Oceanography, 11, 153–175.

Reasor, P. D., Rogers, R., & Lorsolo, S. (2013). Environmental flow impacts on tropical cyclone structure diagnosed from airborne Doppler radar composites. Monthly Weather Review, 141, 2949–2969.

Shay, L. K., Goni, G. J., & Black, P. G. (2000). Effects of a warm oceanic feature on hurricane opal. Monthly Weather Review, 128, 1366–1383.

Smith, R. K., Kilroy, G., & Montgomery, M. T. (2015). Why do model tropical cyclones intensify more rapidly at low latitudes? Journal of the Atmospheric Sciences, 72, 1783–1804.

Stovem, D. R., & Ritchie, E. A. (2016). Simulated sensitivity of tropical cyclone size and structure to the atmospheric temperature profile. Journal of the Atmospheric Sciences, 73, 4553–4571.

Sun, Y., Zhong, Z., Li, T., Yi, L., Camargo, S. J., Hu, Y., … Shi, J. (2017). Impact of ocean warming on tropical cyclone track over the western north pacific: A numerical investigation based on two case studies. Journal of Geophysical Research, 122, 8617–8630.

Ventham, J. D., & Wang, B. (2007). Large scale flow patterns and their influence on the intensification rates of western North Pacific tropical storms. Monthly Weather Review, 135, 1110–1127.

Vincent, E. M., Emanuel, K. A., Lengagne, M., Vialard, J., & Madec, G. (2014). Influence of upper ocean stratification interannual variability on tropical cyclones. Journal of Advances in Modeling Earth Systems, 6, 680–699.

Wang, B., & Zhou, H. (2008). Climate variation and prediction of rapid intensification in tropical cyclones in the western North Pacific. Meteorology and Atmospheric Physics, 99, 1–16.

Wang, C., Wang, X., Weisberg, R. H., & Black, M. L. (2017). Variability of tropical cyclone rapid intensification in the North Atlantic and its relationship with climate variations. Climate Dynamics, 49, 3627–3645.

Wang, S., Camargo, S. J., Sobel, A. H., & Polvani, L. M. (2014). Impact of the tropopause temperature on the intensity of tropical cyclones: An idealized study using a mesoscale model. Journal of the Atmospheric Sciences, 71, 4333–4348.

Wang, X., Wang, C., Zhang, L., & Wang, X. (2015). Multidecadal variability of tropical cyclone rapid intensification in the western North Pacific. Journal of Climate, 28, 3806–3820.
Wang, Y., & Wu, C. C. (2004). Current understanding of tropical cyclone structure and intensity changes—a review. *Meteorology and Atmospheric Physics, 87*, 257–278.

Willoughby, H. E., Clos, J. A., & Shoreibah, M. G. (1982). Concentric eye walls, secondary wind maxima, and the evolution of the hurricane vortex. *Journal of the Atmospheric Sciences, 39*, 395–411.

**How to cite this article:** Ge X, Shi D, Guan L. Monthly variations of tropical cyclone rapid intensification ratio in the western North Pacific. *Atmos Sci Lett.* 2018;19:e814. [https://doi.org/10.1002/asl.814](https://doi.org/10.1002/asl.814)