A Survey of Statistical Relationships between Tropical Cyclone Genesis and Convectively Coupled Equatorial Rossby Waves[※]

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(Received 27 February 2021; revised 23 September 2021; accepted 9 October 2021)

ABSTRACT

Convectively coupled equatorial Rossby waves (ERW) modulate tropical cyclone activities over tropical oceans. This study presents a survey of the statistical relationship between intraseasonal ERWs and tropical cyclone genesis (TCG) over major global TC basins using four-decade-long outgoing longwave radiation (OLR) and TC best-track datasets. Intraseasonal ERWs are identified from the OLR anomalies using an empirical orthogonal function (EOF) analysis method without imposing equatorial symmetry. We find that westward-propagating ERWs are most significant in four tropical ocean basins over the summer hemisphere and that ERWs exhibit similar northeast-southwest (southeast-northwest) tilted phase lines in the northern (southern) hemisphere, with an appreciable poleward advance of wave energy in most TC basins. The EOF-based ERW indices quantitatively show that ERWs significantly modulate TC genesis. The convectively active (suppressed) phases of ERWs coincide with increased (reduced) TCG occurrences. The TCG modulation by ERWs achieves the maximum where the ERWs propagate through the climatological TCG hotspots. As a result, the total number of TCG occurrences in the TC basins varies significantly according to the ERW phase. The ERW-TCG relationship is significant over the northwestern Pacific Ocean, northeastern Pacific Ocean, and the northern Indian Ocean during the northern summer seasons. In the southern summer season, the ERW-TCG relationship is significant over the southern Indian Ocean, Indonesian-Australia basin, and the southwestern Pacific Ocean. However, ERW activities are weak in the main TC development region of the Atlantic Ocean; and the impact on Atlantic TCG appears to be insignificant.

Key words: tropical cyclone genesis, equatorial Rossby waves, statistical analysis

Citation: Wang, S. G., J. Fang, X. D. Tang, and Z.-M. Tan, 2022: A survey of statistical relationships between tropical cyclone genesis and convectively coupled equatorial Rossby waves. *Adv. Atmos. Sci.*, **39**(5), 747–762, <https://doi.org/10.1007/s00376-021-1089-8>.

Article Highlights:

- Significant equatorial Rossby wave activities are identified over major tropical ocean basins in the summer hemisphere.
- The modulation of tropical cyclone genesis by ERWs is maximized where the ERWs propagate through the climatological TCG hotspots.
- The ERW-TCG relationship is most significant over the northwestern Pacific Ocean but weak over the Atlantic Ocean basin.

1. Introduction

Convectively coupled equatorial Rossby waves (ERWs) propagate westward in lower latitudes (Wheeler and Kiladis, 1999; Yang et al., 2007). The ERWs have significant impact on local weather near the equatorial and subtrop-

ical regions. They bring significant amount of precipitation and modulate tropical cyclone activities. In the present work, we focus on the statistical relationship between ERW and tropical cyclone genesis (TCG). The genesis of a TC refers to the beginning of the life cycle of the TC. It represents the transition from a conglomerate of mesoscale convective systems to a strongly rotating and precipitating vortex. To date, TCG remains one of the most challenging issues in TC research (Camargo et al., 2019; Lee et al., 2020). Numerical weather prediction models and climate models struggle to accurately simulate and predict TCG. As such, TCG is often regarded as the main obstacle toward skillful predic-

[※] This paper is a contribution to the special issue on Predictability, Data Assimilation and Dynamics of High Impact Weather—In Memory of Dr. Fuqing ZHANG.

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tion of TC activities.

Professor Fuqing ZHANG and his collaborators have dedicated significant efforts to advance the science of tropical cyclones (e.g., Sippel and Zhang, 2008; Fang and Zhang, 2010, 2011, 2016; Melhauser and Zhang, 2014; Tao and Zhang, 2014; Poterjoy and Zhang 2014; Shu and Zhang, 2015; Tang and Zhang, 2016). Fang and Zhang (2010) examined the genesis of Hurricane Dolly (2008) with cloud-resolving simulations and advanced diagnostics and concluded that vortex stretching plays a crucial role in TC genesis. Fang and Zhang (2016) took a different approach and explored the genesis of Super Typhoon Megi (2010); they attributed the TC genesis to the interaction of various tropical waves. Shu and Zhang (2015) attributed the genesis of Super Typhoon Haiyan (2013) to the Madden Julian Oscillation (MJO) and mixed Rossby-gravity waves. Tang and Zhang (2016) showed substantial impact of the diurnal radiation cycle on the genesis of Hurricane Edouard (2014). Their advanced numerical simulations, together with careful analysis, added significant insights to the genesis of Edouard.

A variety of tropical wave disturbances may modulate TCG by changing the large-scale environment of the TC-generating mesoscale convective systems. Many authors have documented the impact of these tropical waves on TCG. Prominent examples include the Madden Julian Oscillations (MJO), the Boreal Summer Intraseasonal Oscillation (BSISO), convectively coupled Kelvin waves, equatorial Rossby waves, mixed Rossby-gravity waves, easterly waves in the Atlantic and northeastern Pacific Ocean (Dunkerton et al., 2009; Xu et al., 2013; Wang et al., 2010; Wang, 2012), quasi-biweekly oscillations (QBWO) in the northwest Pacific Ocean (Li and Zhou, 2013; Ling et al., 2016), and monsoon gyres (Wu et al., 2013) and troughs (Wu et al., 2012). The westward-propagating QBWO has also been interpreted as regional ERWs in the northwestern Pacific Ocean by some authors (e.g., Chen and Sui, 2010; Yang et al., 2015), with a period of ~12 days (e.g., Li and Zhou, 2013). Among these waves, the ERWs have a relatively long time scale, ranging from 10, to more than 60 days, with a spectral peak of ~30 days (Wheeler and Kiladis, 1999). As a result, the ERWs represent a potentially significant source of TC genesis on intraseasonal time scales, in addition to the other waves (e.g., MJO/BSISO) documented in previous studies (e.g., Camargo et al., 2009).

The focus of the present study is ERWs. The ERWs are predominantly rotational, producing strong convective and vorticity anomalies in the tropical regions outside the deep tropical belt (5°S – 5°N). Several authors have examined the ERW-TCG relationship in different TC basins by applying various techniques. Bessafi and Wheeler (2006) discussed TC genesis and symmetric ERWs in the southern Indian Ocean based on an empirical orthogonal function (EOF) analysis of symmetric OLR anomalies. Several other authors applied the space-time filtering to study the ERW-TCG relationship in the northwestern Pacific Ocean (Frank and

Roundy, 2006; Schreck III et al., 2011; Chen and Chou, 2014; Chen et al., 2018; Zhao and Wu, 2018; You et al., 2019), and northern Indian Ocean (Landu et al., 2020). The underlying physical mechanisms for the ERW-TCG relationship follow from the general notion that TC genesis is, to a large degree, controlled by large-scale environmental factors (Gray, 1979; Emanuel and Nolan, 2004; Tippett et al., 2011) associated with ERWs. Molinari et al. (2007) analyzed observations in detail and suggested that ERWs may increase tropospheric humidity, enhance lower level cyclonic vorticity, and trigger mesoscale convective systems, all of which favor TC genesis. Several other case studies conducted cloud-resolving numerical experiments with advanced data assimilation techniques and added further evidences to this physical mechanism (e.g., Gall et al., 2010; Shu and Zhang, 2015; Fang and Zhang, 2016; Yang and Wang, 2018).

While the ERW-TCG relationship is generally accepted, the preceding literature review indicates that authors have been using different methodologies or datasets to compute ERWs or TCG in different basins. The inconsistency makes it difficult to compare the ERW-TCG relationship from these studies quantitatively. To address this issue, we conduct a survey on the statistical relationship between ERWs and TCG over the tropical ocean basins around the globe using a consistent analysis technique. We plan to examine the ERW-TCG relationship using four-decade-long observational records of outgoing longwave radiation (OLR) and TC best-track datasets over global tropical ocean basins following the same analysis protocol. As shown later, the ERW-TCG relationship is robust over most TC basins based on the EOF characterization of ERWs, but the degree of its modulation varies.

The remainder of this article is structured as follows. Section 2 describes the data and methodology. Results are presented in section 3. Discussion and the conclusions of this study are presented in section 4.

2. Data and Methodology

This study uses a daily-interpolated OLR dataset from the NOAA polar-orbiting satellites (Liebmann and Smith, 1996) with a horizontal resolution of 2.5° . In low latitudes, OLR variability is dominated by variation of the cloud top height while surface temperature plays a minor role. For this reason, OLR is widely used as a proxy for deep convection in tropical and subtropical regions. The main advantages of using the OLR dataset are threefold. (1) Its record started from 1979, and it is much longer than other convective variables available at the global scale; (2) OLR measurements are highly accurate without requiring the inversion of a radiative transfer model; (3) the global OLR datasets are available in near real time. The International Best Track Archive for Climate Stewardship (IBTrACS, Knapp et al., 2010, 2018) is used to analyze TC genesis. The timing and locations of TC genesis are defined as the first occurrence

of a one-minute surface maximum sustained wind speed that exceeds 17 m s^{-1} . We use the IBTrACS and OLR datasets from 1979 to 2019, which comprises 41 years of record.

We employ the EOF analysis technique (Lorenz, 1956) to identify temporal and spatial characteristics of ERWs from the OLR anomalies in the regions of interest. The EOF analysis is one of the simplest machine learning methods to extract spatial modes, and the associated time variability is based on the maximum variance principle. We outline the details of our EOF analysis as follows. Band-pass filtering technique is applied to OLR anomalies to extract westward, 10–60 day signals before the EOF analysis. Because of this westward filtering, the remaining signals contain no standing or eastward information [see discussion in Wang (2020) for the MJO case]. As a result, the westward filtering distinguishes the ERWs from eastward intraseasonal oscillations (the MJO/BSISO). The propagation characteristics of the ERWs may also be verified by examining the lag correlation from the reconstructed signals (see Wang et al., 2018).

After the westward filtering, the OLR anomalies are computed in two steps. First, we extract the OLR daily climatology from the raw OLR data as the sum of the first three harmonic components. Second, we subtract the climatology from the previous step. Through this type of filtering, the leading pair of EOFs are designed to represent propagating signals. They are degenerate and inseparable; the two EOFs represent the same propagating mode with the explained variance as the sum of the two. Any linear combination of the leading pair yields a logical and mathematically consistent representation of the spatial modes (Wang, 2020). One consequence of this property is that the EOF representation is non-unique. For the sake of physical interpretation, it is convenient to rearrange the EOFs and principal components (PCs) using a rotational transformation, such as phase one corresponding to nascent convection at the east end of each region. The amplitude is derived based on the standardized principle time series associated with the two EOFs, defined as $\text{Amp} = \sqrt{\text{PC1}^2 + \text{PC2}^2}$. The phase angle between the two PCs varies from 0 to 360° and is grouped into eight phases in 45-degree intervals.

As shown below, ERWs are active in several distinct ocean basins. The spatial and temporal characteristics of ERWs in these regions vary to some degree (e.g., Yang et al., 2007). Considering the regional variations of ERWs, we analyze regional ERWs separately by applying the EOF analysis to the individual activity centers for each ERW. The two PC time series corresponding to the two leading EOFs are normalized to have standard deviation of 1. The EOF technique is applied to westward-propagating OLR anomalies at the frequency bands of 10–60 days. We focus on those regions in the summer hemispheres. This dependence on seasonality breaks equatorial symmetry in the theoretical solutions of equatorial ERW modes. Because of this, we do not distinguish between symmetric and antisymmetric modes, while at the same time acknowledging that ERWs often refer to theoretical symmetric shallow water modes in the

beta plane (Matsuno, 1966; Kiladis et al., 2009; Fuchs-Stone et al., 2019; Emanuel, 2020). Hence, the “ERWs” discussed in the present study extend beyond the conventional notion of ERWs as an equatorial symmetry mode.

The analysis domains for EOF are chosen such that it is sufficiently large to cover major ERW activities but small enough to allow distinctive and coherent propagating modes to emerge. The spatial domain of the EOFs for the ERWs in the boreal summer seasons are (0° – 30°N , 40° – 180°E) for the northwestern Pacific and Indian Ocean basins; (0° – 30°N , 20° – 140°W) for the northeastern Pacific Ocean and Atlantic Ocean basins; the EOFs for ERWs in the austral summer seasons are (0° – 30°S , 45° – 145°E) in the southern Indian Ocean and Indonesian-Australia basins, and (0° – 30°S , 120°E – 120°W) in the southeastern Pacific Ocean basin.

The statistical significance of the TCG occurrences as a function of discrete ERW phases is assessed with the Chi-Square (χ^2) goodness-of-fit test (Chapter 5, Wilks, 2011). χ^2 is computed as:

$$\chi^2 = \sum_{i=1}^8 \frac{[\# \text{Observed}(i) - \# \text{Expected}(i)]^2}{\# \text{Expected}(i)},$$

where $\# \text{Observed}(i)$ denotes the observed number of TCG events in phase i , and $\# \text{Expected}(i)$ indicates the expected number of TCG events. Here, the null hypothesis, H_0 , is taken as the TCG and is evenly distributed, and $\# \text{Expected}(i)$ is the average number of TCG events for all eight ERW phases. H_0 is rejected if the probability of χ^2 (P -value) is exceedingly small (e.g., 0.05), indicating that the observed TCG occurrences are statistically significant as a function of the ERW phases. This system contains seven degrees of freedom, i.e., the total number of phases minus one.

3. Results

3.1. Overview of the intraseasonal ERWs

Figure 1 shows the space-time spectra (Wheeler and Kiladis, 1999) of the OLR anomalies between 25°S and 25°N in all seasons. At intraseasonal time scales (10–100 days), the MJO and BSISO show the most prominent spectral peaks at eastward wavenumbers 1–5, consistent with previous studies. The ERW continuum shows distinct spectral peaks at westward wavenumbers 2–5 and frequencies ranging from a few days to 100 days. The rest of this subsection briefly reviews the intraseasonal ERWs, focusing on their seasonal and geographical characteristics.

We identify the centers of ERW activities (Fig. 2) based on standard deviation of OLR anomalies at the ERW band (defined as zonal wavenumbers 1 to 10), frequencies less than that of the ERW, and equivalent depths less than 90 meters (area bounded by the closed red curves). Because our interest lies in the TCG-ERW relationship, we focus on

Fig. 1. Space-time spectra of all-season symmetric OLR between 25°S–25°N. The $n = 1$ ERW peaks in the westward wavenumbers 3–4 and frequencies of $\sim 0.02 \text{ d}^{-1}$. The red closed curve indicates regions selected for ERWs: zonal wavenumbers from -10 to -1 , ER frequencies $< 0.1 \text{ d}^{-1}$ with an equivalent depth of 90 m. The black curves denote equivalent depth for the theoretical dispersion relation (for three waves, equatorial Rossby waves, Kelvin waves, and inertial gravity waves) are 8 and 90 m respectively.

the respective summer seasons. Figure 2 shows the standard deviation of OLR anomalies during the northern summer (May–October) in panel a and southern summer (November–April) seasons in panel b. The ERW standard deviation during May–October (Fig. 2a) reaches $\sim 20 \text{ W m}^{-2}$ in the summer (northern) hemisphere, but only $\sim 10 \text{ W m}^{-2}$ in the winter (southern) hemisphere. During November–April (Fig. 2b), the ER standard deviation is also significantly higher in the summer (southern) hemisphere. As a result, ERWs display a stark contrast between the summer and winter hemispheres, in that the ERWs are significantly more active in summer, recalling that the northern hemisphere summer runs from May through October and the southern hemisphere summer spans November through April. The ERWs, in general, maximize over the tropical oceanic areas where convective activities are climatologically prevalent, most notably the ITCZ regions in the tropical Pacific and Indian Oceans. Two distinct centers of ERW activities are present in the boreal summer seasons: the northwestern Pacific Ocean and northeastern Pacific ITCZ region, denoted as regions A and B, respectively, as shown in Fig. 2a. The zonal extent of both areas is broad, spanning more than 120 degrees in longitude; however, the meridional extent is confined to about 30 degrees in latitude. Figure 2b shows that

Fig. 2. The standard deviation of OLR anomalies (W m^{-2}) associated with ERWs (zonal wavenumbers 1 to 10, periods of 10 to 90 days, equivalent depths less than 90 m, see red box in Fig. 1) in the extended boreal winter season (April to November), and summer season (November to April). Black/gray boxes indicate the area for the EOF analysis in different basins: A for the northwestern Pacific and northern Indian Ocean, B for the Atlantic and northeastern Pacific Ocean, C for the southwestern Pacific Ocean, and D for the southern Indian and Indonesia–Australia basin.

during the austral summer season (TC season in the southern hemisphere), ERWs maximize in two regions: the southwestern Pacific Ocean and the southern Indian Ocean denoted as C and D, respectively. Although the two are seemingly connected by visual inspection, as will be discussed later, it is more convenient to analyze the ERW-TCG relationship in the two TC basins separately. Other significant ERW activity regions include those in the off-TC seasons or those regions with little TC activities (e.g., the southern Atlantic). The enhanced ERW activities in the summer hemisphere from our analysis differ from Wheeler and Kiladis (1999, their Figs. 7g and 7h), which imposed symmetry for ERWs in compliance with Matsuno's theory for tropical waves.

The enhanced ERW activities in these regions suggest a potential impact of ERWs on TC activities, following the theoretical argument that enhanced low-level vorticity from ERWs may favor TC genesis. While the ERW-TCG relationship has been documented in some regions, as reviewed in the introduction section, to the best of the authors' knowledge, no prior studies have systematically examined whether or not there are similar ERW-TCG relationships in the eastern Pacific Ocean, Gulf of Mexico, Atlantic Ocean, southwestern Pacific Ocean, and Indonesian-Australia Ocean basins. As shown in Fig. 1, the wavenumber 3–4 ERWs propagate through these basins in the extended boreal summer seasons. Given their sizable spatial and temporal extent, the ERW-TCG relationship may be significant

across different TC basins, or there may be considerable regional dependence. In the following subsections, we will perform a statistical analysis of ERWs and TCG in these regions using the same EOF analysis method and further quantify the ERW-TCG relationship as a function of the ERW phases.

3.2. The ERW-TCG Relationship over Major TC basins

3.2.1. Northwestern Pacific Ocean and Northern Indian Ocean

We first examine the ERWs in the northwestern Pacific Ocean and the northern Indian Ocean. Figure 3 shows the two leading EOFs in this region, which explain 8.6% and 8.5% of the total variance in the selected spectral band, respectively. Collectively, they represent a single westward propagating mode that explains 17.1% of the total variance. The error bars of the explained variances are well separated from the other modes (not shown), indicating that they are statistically significant according to North's EOF testing criteria for eigenvalue separation (North et al., 1982). The first two EOFs represent one single propagating mode, and they are indistinguishable both statistically and dynamically. Both EOFs extend from the Indian Ocean to the dateline (Fig. 3). The EOF patterns approach near-zero values near the boundaries, indicating that the EOF analysis domain is adequate. The EOFs show maximum values around 15°N, with near-zero values approaching the equator (0°–5°N) and

Fig. 3. EOF analysis of OLR anomalies (shaded; westward-propagating, 10–60 days periodicity, May to November) in region A. Panels (a) and (b) show the spatial patterns of EOFs 1 and 2, respectively. For reference, the smoothed climatology of TCG in the western Pacific (green contours, unit: number of TCG events per year) and northern Indian Ocean (magenta contours) are shown in both panels. The TCG climatology is obtained by counting TCG within every 2-degree box subject to the Gaussian filtering. Explained variances for EOF 1 and 2 are 8.6% and 8.5%, respectively. The dashed boxes indicate the regions where TCG is computed in these two basins, respectively.

subtropical latitudes (30°N), broadly consistent with Matsuno's theory for ERWs. Both EOF patterns display the southwest-northeast tilt phase lines in the northwestern Pacific Ocean and East Asia. Zonal wavelengths are ~ 100 degrees of longitude at lower latitudes (e.g., 60° – 160°E at 5°N for EOF2) and much shorter in higher latitudes (e.g., 60 degrees at 20°N), indicating significant wave dispersion. The signs of the EOFs are chosen such that northwest

propagation is present, consistent with known ERW dynamics. These two leading EOFs maximize in the open ocean, specifically within the main development region for TCs in this basin. Notably, the ERW convective signals reach many coastal regions, including southern China and South Asia, suggesting a potentially significant impact on these coastal areas.

Figure 4 shows the phase composite of the OLR anom-

Fig. 4. Composite of OLR anomalies (shaded, W m^{-2}) and TCG occurrences (dots) in the northern Indian Ocean and northwestern Pacific oceans in the 8 ERW phases. Black dots denote TCG in the northwestern Pacific Ocean basin and magenta dots for TCG in the northern Indian Ocean basin. The smoothed TCG climatologies in the northwest Pacific (same in all the panels, green contours, $0.5, 1 \text{ yr}^{-1}$, as in Fig. 3) and northern Indian Ocean basin (magenta contours, $0.1, 0.4 \text{ yr}^{-1}$) are shown for reference. The number of days in each phase is marked in the top right corner of each panel.

alies and TCG occurrences for each individual phase. By phase compositing, we group variables according to the eight ERW phases when the amplitudes of the normalized PCs exceed 0.75. The spatial pattern and magnitudes are robust whether or not filtering is used for computing the OLR anomalies. The OLR anomalies (shaded colors in Fig. 4) reach more than 20 W m^{-2} in both the convectively suppressed and active regions. Northwestward propagation may be inferred from the OLR anomalies, which may also be confirmed by lag correlation analysis (Wang et al., 2018). The ERWs show a predominant westward propagation with appreciable poleward components in other basins and the summer hemispheres. Nevertheless, poleward propagation is absent in the canonical shallow water theory of equatorial waves.

We consider the days when ERW amplitudes (computed as the amplitude of two ERW normalized PCs) are greater than 0.75 for TCG occurrences. This condition is met on approximately 75% of the total summer days (as derived empirically from its statistical distribution). About 20%–30% of the total TCG events occurred when the ERW amplitude was less than 0.75, and the influence by these weak ERWs is considered minor hence excluded. As the westward propagating ERWs sweep through the northwestern Pacific Ocean, TCG is significantly modulated in both the active and suppressed ERW phases. The maximum climatological area for TCG (green contours in 7.5° – 20°N , 125° – 155°E) collocates with the positive (suppressed) OLR anomalies in phases 8 and 1, and negative (convective) OLR anomalies in phases 3–5. The most significant negative ERW-OLR anomalies (convective phase) occur in phases 3–5, corresponding to higher TCG occurrences, while the positive ERW-OLR anomalies in phases 1 and 8 correspond to lower TCG occurrences within the suppressed phases. The phase composite shows that convectively active OLR anomalies of ERWs increase TCG occurrences, while suppressed anomalies decrease TCG. As a result, ERWs modulate the locations of TCG. The spatial patterns of the OLR anomalies (Fig. 4) resemble those in Zhao and Wu (2018), while the total number of ERWs with respect to ERW phases differs from those in their Fig. 2c. This difference may be attributed to the use of different variables to identify ERWs, differing lengths of record, or the technical details within the EOF analysis (i.e., precipitation was used in that study, the observational period was 1998–2012, and upon a smaller analysis domain).

In addition to the modulation of TCG locations, ERWs also change the total number of TCGs, because the spatial extent of the TC basin is much smaller than the wavelength of the ERWs. Figure 5a shows that TCG in both basins varies smoothly with respect to the ERW phases. Tropical cyclone genesis (TCG) is computed for the regions (0° – 30°N , 120° – 180°E) in the western Pacific Ocean basin, and (0° – 30°N , 50° – 100°E) in the northern Indian Ocean basin (as denoted by the dashed boxes in Fig. 3). The number of TCG events in the northwestern Pacific basin is about 100

in the most active ERW phase from 1979–2019, and the number of events is less than 50 in the most suppressed phase. The average number of TCG events is 73 for each phase or, equivalently, 1.78 per year for each phase. The active phases of ERW increase TCG by nearly 30% (100 in phase 4 compared to the all-phase average of 73). The P -value for the number of TCG as a function of eight ERW phases is 6.8×10^{-8} from Eq. (1), indicating that the TCG-ERW in this region is very unlikely to occur at random.

The OLR anomalies due to the ERW are significantly weaker in the northern Indian Ocean compared to the northwestern Pacific Ocean (Fig. 2). In the Bay of Bengal, where a local TCG maximum is present, the OLR anomalies are $\sim 5 \text{ W m}^{-2}$, much less than that in the northwestern Pacific. Tropical cyclone genesis (TCG) in the northern Indian Ocean is much less frequent. Nevertheless, modulation of TCG by ERWs in this region is still statistically significant, despite that both signals are weaker. The magenta dots in Fig. 5 mark TCG in this region. Phase 8 corresponds to enhanced convective activities in the Bay of Bengal, where climatological TCG peaks locally in this region. The opposite occurs in phase 4. The number of TCG in this region

Fig. 5. The number of TCG events (1979–2019) for each phase in region A for the northern Indian Ocean and northwestern Pacific Ocean basins. (a) Number of TCG events in each ERW phase in the northwestern Pacific and northern Indian Oceans, and (b) the TCG count anomalies normalized by the number of years. The average number of TCG in each phase is 10, 73 over the period, or 0.25, 1.79 yr^{-1} , in the Indian Ocean and western Pacific basin, respectively. The label of the right axis denotes the normalized TCG count anomalies by decade. The P -values for the number of TCG events in the two basins are 6.8×10^{-8} and 0.06, respectively.

show a smooth transition from phase 1 to 8 (Fig. 5b), ranging from six in phase 4 to ~20 in phase 8 during 1979–2019. The average number of TCG events is ~11 in each phase, or 0.25 per year, in the Indian Ocean. TCG occurs most frequently in phase 8 (21), which is 80% higher than the overall average. The P -value for the number of TCG is also small (0.06), which is marginally significant.

3.2.2. Northeastern Pacific Ocean, Gulf of Mexico, and the Atlantic Ocean

Figure 6 shows the two leading EOFs in the northeastern Pacific Ocean and the Atlantic Ocean. These two EOFs explain 6.9% and 6.8% of the total variance, respectively. The EOFs show alternating patterns from the Atlantic to the Gulf of Mexico/Caribbean sea and continuing westward to the northeastern Pacific Ocean. Their amplitudes are relatively weak in the Atlantic Ocean (east of 60°W). The EOFs show northeast-southwest tilted phases, consistent with the EOFs in the western Pacific (Fig. 3). The first EOF (EOF1) peaks around (15°N, 100°W). The ERW signals are nearly non-existent south of 10°N in the northeastern Pacific Ocean. The climatology of TCG (green contours) in the northeast Pacific Ocean shows a local maximum in the northeastern Pacific ITCZ region, with the maximum values reaching nearly 2–3 TCG per year at (15°N, 105°W). The main development region (MDR) in the Atlantic Ocean has TCG occurrences ~0.2–0.5 per year.

Figure 7 shows the composite of the OLR anomalies and TCG over the eight ERW phases in this region. The am-

plitude of the OLR anomalies reaches ~15 W m⁻², notably weaker than those in the northwestern Pacific basin (Fig. 4). Consistent with the EOF pattern (Fig. 6), the ERW is the strongest in the eastern Pacific ITCZ region, moderate in the Gulf of Mexico/Caribbean sea, and nearly non-existent in the Atlantic Ocean. More TCG occurrences are found in phases 3–5, as the ERW convective phases sweep through the climatological hotspots in the northeastern Pacific region. Modulation of TCG by ERWs appears to be weak in the Atlantic MDR region. Figure 8 shows the total number of TCG in the northeastern Pacific Ocean (0°–30°N, 140°W to the Pacific coast of northern America) and the Gulf of Mexico (15°–30°N, 82°–88°W). Modulation of TCG by ERWs in the Atlantic MDR (the Caribbean Sea and the Atlantic Ocean) is insignificant. Consistent with the spatial patterns (Fig. 7), the total number is higher in phases 3–5, with more than 70 TCG events from 1979–2019, and it is the lowest in phase 8 with 30 TCG in the same period. On average, the number of TCG events in each phase is 60 from 1979 to 2019 or, equivalently, 1.5 per year in each phase. Phase 3 has the highest number of TCG events (85), ~30% more than the average (60); phase 8 has the lowest number of TCG events (30), ~50% reduction from the mean. The P -value is exceedingly small (1.4×10^{-6}), indicating that the TCG distribution as a function of the ERW phases is unlikely to occur by chance.

3.2.3. Southwestern Pacific Ocean

Figure 9 shows the two leading EOFs in the southwest-

Fig. 6. EOF patterns of OLR anomalies (westward-propagating, 10–60 days periodicity) in the Atlantic and northeastern Pacific Oceans (region B in Fig. 2a). Panels (a) and (b) show the spatial patterns of EOFs 1 and 2, respectively. For reference, the smoothed climatologies of TCG in the northeast Pacific Ocean, the Gulf of Mexico, Caribbean Sea, and western Atlantic Ocean (green contours; units: number of TCG events per year) are shown in both panels. Explained variances of EOFs 1 and 2 are 6.9% and 6.8%, respectively.

Fig. 7. Composites of OLR anomalies (W m^{-2} , shaded) in the eight ERW phases. Dots: TCG in the Atlantic and northeastern Pacific Oceans. The smoothed TCG climatology (same in all the panels, green contours, $0.2, 1 \text{ yr}^{-1}$, as in Fig. 6) is shown for reference. The number of days in each phase is marked in the top left corner of each panel.

ern Pacific Ocean basin (region C in Fig. 2). Both EOFs display northwest-southeast tilted structures. Explained variances for EOF 1 and 2 are 11.0% and 10.7%, respectively. Unlike the EOFs in region A (Fig. 2a), the zonal wavelength remains nearly the same across different latitudes. Tropical cyclone genesis (TCG) is mainly found in the region ($10^{\circ}\text{--}20^{\circ}\text{S}$, $150^{\circ}\text{E}\text{--}150^{\circ}\text{W}$).

Composites of the OLR anomalies and TCG over the

ERW phases in this region are displayed in Fig. 10. The poleward shift of the wave troughs and crests indicates a poleward dispersion of wave energy. The amplitude of the OLR anomalies reaches $\sim 20 \text{ W m}^{-2}$, notably weaker than those in the northwestern Pacific basin but stronger than those in the northeastern Pacific basin. Enhanced TCG and convective ERW anomalies tend to coincide, while the TCG occurrence in the suppressed ERW anomaly area is significantly

less. Figure 11 shows that the total number of TCG events (within the region 0° – 30° S, 140° E– 90° W) is largest in phase 5, where the northwest-southeast tilted convective ERW anomalies are located within the longitudes 170° E– 160° W. The lowest TCG occurrence is found in phase 1, when suppressed ERW anomalies are present.

3.2.4. Southern Indian Ocean and Indonesian-Australia basin

The southern Indian Ocean and the Indonesian-Australia basin are interconnected. Our tests indicate that it is pos-

Fig. 8. The number of TCG events (1979–2019) for each phase in region B, the northeastern Pacific Ocean, the northwest Caribbean sea, and the Gulf of Mexico. The average number of TCG events over 1979–2019 is 60 (or 1.5 yr^{-1}) for each phase. The label of the right axis denotes the normalized TCG count anomalies by decade. The P -value is 1.4×10^{-6} .

sible to perform EOF analysis for ERWs in the individual basins separately or combine them with one set of EOFs. For the sake of simplicity, we opt to use one pair of the EOFs in the following discussion without any further rigorous justification.

Figures 12 shows the two leading EOFs in the southern Indian Ocean (region D). Explained variances for EOF 1 and 2 are 7.5% and 7.4%, respectively. The phase lines tilt northwest-southeast, as in the southwestern Pacific Ocean (Fig. 9). The EOF patterns show the maximum values around 10° – 15° S, with near-zero values near the equator and subtropical latitudes ($\sim 30^{\circ}$ S). As the ERWs propagate westward, the wave crests/troughs shift southward, again indicating poleward energy dispersion. TCG occurrences are nearly zonally distributed over the ITCZ region.

Figure 13 shows that the location of convective ERW anomalies and high TCG occurrences coincide in the southern Indian Ocean; vice versa, dry anomalies coincide with low TCG. Convective anomalies in phases 4–5 coincide with the climatological TCG maximum and correspond to clustered TCG occurrences; opposite behavior may be found in phases 1 and 8. The results are broadly consistent with Bessafi and Wheeler (2006), which shows that TCG is concentrated in the region of ERW convection and cyclonic circulation cells. Over the Indonesian-Australia basin, the clustering of TCG is most significant over phase 1 when ERW convection is active over this region. Conversely, TCG occurrences achieve the minimum in phases 4 and 5 when suppressed ERW occurs in this region.

The number of TCG events is computed for the region (0° – 30° S, 140° E– 120° W) for the southern Indian Ocean

Fig. 9. EOF analysis of OLR anomalies (westward-propagating, 10–60 days periodicity) in region C (southwestern Pacific Ocean). Panels (a) and (b) show the spatial patterns of EOFs 1 and 2, respectively. Explained variances for EOFs 1 and 2 are 11.0%, 10.7%, and they are statistically indistinguishable. For reference, a smoothed climatology of TCG over this region (green contour, units: number of TCG events per year) is shown in both panels.

Fig. 10. Composite of OLR anomalies in the eight ERW phases (shaded). Dots: TCG events in the southern Pacific Ocean. The smoothed TCG climatology (same in all the panels, green contours, $0.15, 0.4 \text{ yr}^{-1}$) in the southwestern Pacific Ocean is shown for reference.

(Fig. 9) and (0° – 30° S, 45° – 110° E) for the Indonesian-Australia basin. Figure 14a shows that phase 4 has the highest TCG occurrences during 1979–2019, reaching ~ 70 in the southern Indian Ocean, and the lowest count (20) occurs in phase 1. The total number of TCG events over the Indonesian-Australia basin (Fig. 14b) is highest in phase 1 (30) and lowest in phase 5 (4). The P -values of TCG for the Indian and Indonesian-Australia basins are exceedingly small (2.3

$\times 10^{-7}$ and 2.9×10^{-7} , respectively), indicating that the TCG distribution, as a function of the ERW phase, cannot be attributed to randomness.

4. Summary

Atmospheric circulation systems are instrumental in setting the stage for tropical cyclone genesis. On intraseasonal

time scales, eastward propagating waves (BSISO/MJO) and westward propagating ERWs play significant roles in tropical cyclone genesis (TCG). This study focuses on the latter and presents a global survey of the statistical relationship between ERW and TCG over major TC basins. Following the same analysis protocol, intraseasonal ERWs are consistently identified by applying the same EOF analysis method to the westward propagating OLR anomalies. Equatorial symmetry is not imposed in the process of identifying ERWs.

Fig. 11. Left: The number of TCG events (1979–2019) for each phase in region C, southern Pacific basin. The label of the right axis denotes the normalized TCG count anomalies by decade. The average number of TCG events in each phase is 30.0 over the period (or 0.73 yr⁻¹) in the southwest Pacific Ocean. The *P*-value of TCG events as a function of phase is 0.036.

As a result, the morphology of ERWs departs markedly from theoretical $n = 1$ ERW wave dynamics. The results are summarized as follows:

(1) In the summer hemisphere, four regions of heightened ERW activities are identified: the northwestern Pacific Ocean and the northern Indian Ocean during the northern summer, and the southwestern Pacific Ocean and the southern Indian Ocean during the southern summer. Westward propagating ERWs, derived from the EOF analysis of intraseasonal OLR anomalies in various TC basins, exhibit northeast-southwest tilted phase structure in the northern hemisphere and a corresponding southeast-northwest tilt in the southern hemisphere. The nature of the tilted structures emerging from these observations are consistent with observational analyses in previous studies (e.g., Kiladis and Wheeler, 1995; Yang et al., 2007; Kiladis et al., 2009). The meridional tilted structure is absent in canonical equatorial Rossby wave theory (Matsuno, 1966) but in qualitative agreement with the westward propagating moisture mode proposed by Fuchs-Stone et al. (2019) and Emanuel (2020). Both studies couple Rossby wave dynamics with moisture, surface fluxes, and cloud-radiation feedback. Nevertheless, there is still a significant discrepancy between the westward propagating moisture mode theory and our observations that show ERWs having a spectral peak at zonal wavenumbers 3–4. Moisture mode theory predicts the most unstable mode at wavenumber 1 or 2. This discrepancy suggests a need to further improve the theoretical understanding of convectively coupled ERWs.

(2) Across the four regions of heightened ERW activi-

Fig. 12. Spatial patterns of the two leading EOFs of OLR anomalies (westward-propagating, 10–60 days periodicity) in region D (southern Indian Ocean). Panels (a) and (b) show the spatial patterns of EOFs 1 and 2, respectively. The smoothed TCG climatology in the southern Indian Ocean and Indonesian-Australia basin is shown for reference. Explained variances for EOFs 1 and 2 are 7.5% and 7.4%, respectively.

Fig. 13. Composite of OLR anomalies (shaded) and TCG occurrences (dots) for the eight ERW phases over the southern Indian Ocean. The smoothed TCG climatology (same in all the panels, green contours, 0.2, 0.6 yr^{-1} , as in Fig. 12) in the southern Indian basin is shown for reference. The number of days in each phase is marked in the top left corner of each panel.

ies in the summer hemispheres, the EOF patterns for the ERWs display their maximum values around 10–15 degrees in both hemispheres, with vanishing values near the equator and subtropical latitudes (~ 30 degrees). As the ERWs propagate westward, the wave crests/troughs shift poleward in most basins, suggesting wave propagation toward high latitudes.

Nevertheless, ERWs originating from the Atlantic Ocean show no poleward propagation, while the poleward propagation is maximized in the northeast Pacific Ocean along the ITCZ region. The OLR anomalies associated with ERWs reach amplitudes of $\sim 20\text{--}30 \text{ W m}^{-2}$ over all the TC basins. The strongest ERW signals are found in the northwest-

Fig. 14. The number of TCG events (1979–2019) for each phase in (a) the southern Indian Ocean and (b) the Indonesian-Australian basin. The P values are 2.3×10^{-7} and 2.9×10^{-7} , respectively, for the number of TCG events in panels (a) and (b).

ern Pacific Ocean ($\sim 25\text{--}30 \text{ W m}^{-2}$). The ERWs are weaker in the northeastern Pacific Ocean, and the amplitude is $\sim 15 \text{ W m}^{-2}$.

(3) The ERW activities are quantified using EOF-based ERW indices. It is shown that ERWs modulate the timing and location of TCG in several major TC basins, consistent with previous studies. The ERW-TCG relationship is significant over the northwestern and northeastern Pacific Ocean and the northern Indian Ocean during the northern summer season and over the southern Indian Ocean and southwestern Pacific Ocean during the southern summer season. Climatological ERW activities are weak in the main development region of the Atlantic Ocean, and their direct impact on TCG appears to be insignificant in these regions. The ERW convectively active phases coincide with increased TCG occurrences, while the ERW suppressed phases coincide with much reduced TCG occurrences. To the best of the authors' knowledge, impacts of ERWs (or the lack thereof) on TCG in the eastern Pacific, southwestern Pacific, Atlantic, and Indonesian-Australia Oceans, as well as the Gulf of Mexico/Caribbean sea, have yet to be documented in prior peer-reviewed literature.

(4) Modulation of TCG by ERWs maximizes where the ERWs propagate through the climatological TCG hotspots in tropical ocean basins. As a result, the total numbers of TCG events in different basins are modulated by ERWs by 30%–50% relative to the climatological TCG numbers

across these basins, with the notable exclusion of the Atlantic main development region.

The above results indicate that the TCG-ERW relationship is robust across several major TC basins where ERW activities are significant. The scientific understanding of this statistical relationship, however, remains limited. We hypothesize that ERWs generate significant low-level vorticity and convergence anomalies, which may modulate the likelihood of TC precursors and the transition probability of these precursors to TCs (Hsieh et al., 2020). As a result, ERWs alter the large-scale environment to either support or suppress TC genesis.

The present study has focused on the statistical relationship between the TCG and ERWs assuming that this relationship is linearly independent from the influence of other tropical waves. Incorporating the synergistic effects of ERWs with other tropical waves may help improve the understanding of the climatological distribution of TCG in time and space. We suggest that our results may have significant implications for predicting TC activities at extended lead times. A better understanding of ERWs and other prominent tropical intraseasonal oscillations may lead to improved prediction of TC genesis and subsequent development over global TC basins.

Acknowledgements. The authors dedicate this study to the memory of Prof. Fuqing ZHANG, an inspiring mentor, a dear friend, and a dedicated scientist. Prof. ZHANG made fundamental contributions to the sciences of weather prediction and tropical cyclones. This work is supported by NSFC (Grant No. 41875066). We also thank the three anonymous reviewers for their thoughtful and constructive comments, which significantly improved the clarity and interpretation of this work.

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