Influence of the Monsoon Trough on Westward-Propagating Tropical Waves over the Western North Pacific. Part I: Observations

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ABSTRACT

The present study investigates the possible linkage between the monsoon trough and the interannual variability of the activity of westward-propagating tropical waves (WTW) over the western North Pacific (WNP) during July–November for the period 1979–2007. It is shown that the interannual variability of WTW activity is closely related to the location of the monsoon trough. During the years when the enhanced (weakened) monsoon trough extends eastward (retreats westward), the lower-tropospheric WTW activity is above (below) normal within the southeastern quadrant of the WNP. Furthermore, this study evaluates different wave structures and dynamics of two types of WTWs, equatorial Rossby (ER) waves and mixed Rossby–gravity (MRG)–tropical depression (TD)-type waves, in strong monsoon trough (S-MT) and weak monsoon trough (W-MT) years over the WNP. There is a significant change in the three-dimensional structure as those waves propagate westward to the east of the monsoon trough. For the TD–MRG waves, an apparent transition from MRG waves to off-equatorial TD disturbances is identified in the region of the monsoon trough. For the ER waves, their amplitudes have a faster growth, but their structures and propagation characters have no marked change. Differences in the location of the monsoon trough may lead to an east–west contrast in the WTWs. In a companion study (Part II), diagnostics of energetics and numerical experiments are conducted to explain the observed results in the present study.

1. Introduction

The understanding of tropical waves is fundamental to atmospheric dynamics in the tropics on a wide range of space and time scales. Tropical waves can cause atmospheric oscillations in pressure, temperature, humidity, convection, and winds, with the magnitude of such oscillations being large enough to directly affect tropical weather (Wheeler 2002). Westward-propagating tropical waves (WTWs) are the primary tropical waves over the tropical western North Pacific (WNP) during the Northern Hemisphere summer and fall. Early studies (e.g., Matsumo 1966; Wheeler and Kiladis 1999) suggested that there are three kinds of WTWs over the Pacific: equatorial Rossby (ER) waves, mixed Rossby–gravity (MRG) waves, and tropical depression (TD)-type disturbances. These waves propagate westward along easterly trade winds, and the existence of tropical westerlies over the western tropical Pacific during boreal summer affects the activity of these disturbances (Wallace and Chang 1969; Chen and Weng 1998). Because of their role in modulating tropical weather in the WNP, these disturbances have attracted the attention of researchers.

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The activity of these westward-moving waves plays an important role in modulating tropical cyclones over the WNP (Frank and Roundy 2006; Schreck et al. 2011, 2012, 2013). Previous studies (e.g., Dickinson and Molinari 2002; Ajiyer and Molinari 2003; Fu et al. 2007; Zhou and Wang 2007; Wu et al. 2014a; Kiladis et al. 2009) have shown that a lower-tropospheric MRG wave could subsequently evolve into an off-equatorial synoptic-scale TD-like disturbance as it propagates to the western Pacific, and the disturbance eventually develops into a tropical cyclone. Molinari et al. (2007) showed that tropical cyclogenesis occurred within the region of convection and cyclonic vorticity in the eastern part of the ER wave low. Gall and Frank (2010) and Gall et al. (2010) examined the role of equatorial Rossby waves in tropical cyclogenesis. They found that wave breaking of the ER wave is a mechanism by which vorticity is organized on the scale of a tropical cyclone. Therefore, unraveling the evolution of the WTW structures is a key step in advancing our understanding of the role of tropical waves in tropical cyclogenesis.

The structure and propagation of tropical waves is strongly affected by the interactions with large-scale circulation. Several observational studies have shown that the lower-tropospheric WTWs can experience change in structure and propagation characteristics when they propagate from the central Pacific to the western Pacific (Sobel and Bretherton 1999; Dickinson and Molinari 2002; Yang et al. 2007). A number of studies have been devoted to understanding the influence of mean flow on tropical waves. Webster and Chang (1988) and Chang and Webster (1990, 1995) suggested that the accumulation of wave energy may occur in a confluence zone, and this wave energy accumulation plays a potentially important role in the amplification and scale contraction of waves (Holland 1995; Done et al. 2011). Moreover, Sobel and Bretherton (1999), Sobel and Maloney (2000), and Maloney and Hartmann (2001) found that the large-scale summer circulation over the WNP favors conversion of its energy to wave disturbances, which are an energy source for the evolution of wave disturbances and the formation and maintenance of tropical cyclones. These studies indicated that the background flow in the WNP may provide favorable conditions for the changes of WTWs through the wave–mean flow interaction.

In summer and fall, the large-scale low-level circulation over the WNP features low-latitude westerly monsoonal winds and westerly monsoonal winds (Briegel 2002; Chan and Evans 2002; Tomita et al. 2004). It exhibits strong seasonal (Atkinson 1971; Chia and Ropelewski 2002; Wu et al. 2014b) and interannual variations (Wu et al. 2012). The monsoon trough extends southeastward from the Asian continent through the South China Sea to the western North Pacific during June–September and shifts southward with a closed cyclonic circulation during October and November. The monsoon trough experiences a remarkable east–west displacement following ENSO (Wu et al. 2012; Clark and Chu 2002). Some observational studies have noted that, in the WNP, a strong relationship exists between the monsoon trough change and the longitudinal shift in the location of tropical cyclone (TC) formation (e.g., Wu et al. 2012; Molinari and Vollaro 2013; Cao et al. 2014). As a prominent feature in the tropical WNP, the displacement of the monsoon trough could be also a key large-scale circulation factor for the evolution of WTWs. Chen and Weng (1998) noted that the activity of synoptic-scale disturbances exhibits characteristics similar to the variability in the monsoon trough. Studies indicate that the monsoon trough is a background field favorable for the growth and transformation of wave disturbances (Sobel and Bretherton 1999; Wu et al. 2012, 2014a,b), and WTW activity may be linked to changes in the activity of tropical cyclones over the WNP on interannual time scales (e.g., Roundy and Frank 2004; Dickinson and Molinari 2002; Gall et al. 2010; Gall and Frank 2010), though this connection remains controversial. Gray (1998) found that a westward-propagating wave packet with cloud cluster entering a monsoon trough can bring about an effective wind surge in the vicinity of the trough, and convection becomes rapidly enhanced and causes cyclogenesis. Although these studies partly documented the WTW activity associated with a monsoon trough over the WNP, less attention has been given to the detailed structure and evolution of the WTWs and their interaction with the monsoon trough.

The present study focuses on the statistical representation of the WTW characteristics in the WNP. Our objective is to reveal the detailed three-dimensional structure and evolution of the TD–MRG waves and ER waves during different states of the monsoon trough. The results suggest that differences in the basic state of the monsoon trough can influence the nature of the WTWs that grow on it. In the second part of this paper (Wu et al. 2014, manuscript submitted to J. Climate, hereinafter Part II), we attempt to explore the physical mechanism for the monsoon trough’s modulation of the TD–MRG waves and ER waves in the WNP through diagnosis of the energy conversion and numerical experiments. The present analysis is restricted to the period 1979–2007 to maintain consistency with a former study (Wu et al. 2012). The period after 2007 will be analyzed in a future study for confirmation of operational
predictions of tropical cyclogenesis. The article is arranged into five sections. **Section 2** describes data and the data processing methods. In **section 3**, we present the activity of WTWs in the WNP during the different states of the monsoon trough. The detailed three-dimensional structure and evolution features of TD–MRG waves and ER waves are shown in **section 4**. A summary of results is presented in **section 5**.

### 2. Data and methods

#### a. Data

Two primary datasets are utilized in the present study. The outgoing longwave radiation (OLR) from NOAA polar-orbiting satellites (Liebmann and Smith 1996; http://www.esrl.noaa.gov/psd/data/gridded/data.interp_OLR.html) is used as a proxy for tropical convection. The daily wind fields are obtained from the National Centers for Environmental Prediction (NCEP)–U.S. Department of Energy (DOE) AMIP-II reanalysis (NCEP-2; Kanamitsu et al. 2002; http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html). The apparent heat source Q1 (e.g., Yanai et al. 1973; Yanai and Johnson 1993) is computed from NCEP-2 at 12 standard pressure levels (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa). The values of Q1 are calculated for each successive layer between the standard pressure levels and given at 11 levels from 1000 to 100 hPa. All the datasets are available on a global 2.5° grid and extend from 1979 to 2007. Only the data from July to November are used in this study.

#### b. Methods

Spectral analyses are employed to reveal the distribution of power in the wavenumber–frequency domain of prominent wave types. Conventionally, the wavenumber–frequency analysis is carried out along a full longitude circle (e.g., Wheeler and Kiladis 1999; Hendon and Wheeler 2008). Following Wu et al. (2014a), we reduce the wavenumber–frequency analysis from a global domain to a finite domain, considering the fact that the prominent wave types are primarily confined to the WNP region. This procedure is done in a similar way to the global domain, except that the values of variables are reduced to zero outside the analysis domain. A comparison of the wavenumber–frequency spectra shows that the wave spectrum extracted from the global domain can be different from that based on the WNP domain. The spectral analyses of a finite domain can efficiently extract regional spectral characteristics of zonally propagating waves and provide regional information to the WTWs.

This study employs zonal space–time spectral filtering of all the fields to extract the signals of the westward-propagating TD–MRG waves and ER waves that are dominated by zonal propagation and synoptic frequencies. The spectral filtering techniques based on the zonal global data field are described in detail in Wheeler and Kiladis (1999), Roundy and Frank (2004), and Hendon and Wheeler (2008). The TD–MRG waves cover a period of 2–10 days, with negative zonal wavenumbers ranging from wavenumber 1 to 14, and ER waves that have a 10–48-day period and westward wavenumbers 1–10 that have not been identified with a specific wave type. Here, The TD–MRG wave band is a combined TD-type and MRG wave band. It is hard to separate entirely these two types of higher-frequency waves using this kind of spectral filtering because of their overlapping in the wavenumber–frequency space, and these two distinct types of waves may evolve from one to another or coexist in the same spectral domain while propagating westward. Here, our analysis below does not try to separate them. These filtering bands are similar to those of Wu et al. (2014a) and Frank and Roundy (2006).

To examine the distinctive structures and propagation features of the TD–MRG waves and ER waves between the strong monsoon trough (S-MT) and weak monsoon trough (W-MT) years, a lagged regression analysis is applied for S-MT and W-MT cases, respectively. The statistical significance of the anomalies is estimated using the Student’s t test.

The strength of the TD–MRG (ER) wave activity may be measured by the eddy kinetic energy (EKE):

\[ K^2 = \frac{1}{2}(u^2 + v^2), \]

where an overbar represents a time average during July–November, a prime denotes the TD–MRG (ER) disturbance field, and variables \( u \) and \( v \) represent the zonal wind, the meridional wind, and the kinetic energy, respectively.

### 3. Monsoon trough and interannual variation of wave activity

#### a. Interannual variation of monsoon trough

A monsoon trough index (MTI) based on Wu et al. (2012)’s criterion is defined as the time coefficient of the leading EOF mode to the average latitude (5°–20°N) of the 850-hPa positive relative vorticity in the 100°E–180° longitudinal band for July through November in each year. **Figure 1a** shows this index for each year from 1979 to 2007. Essentially, positive (negative) values of the MTI indicate that cyclonic vorticity in the monsoon trough increases (decreases) over the WNP. The MTI indicates that the years of positive (negative) values are
well synchronized with the eastward extension (westward retreat) of the enhanced (reduced) monsoon trough (Wu et al. 2012). In addition, strong monsoon trough (S-MT) and weak monsoon trough (W-MT) years are defined according to the value of the MTI. The S-MT years with the seven highest values of the MTI include 1982, 1986, 1990, 1991, 1997, 2002, and 2004; and the W-MT years with the seven lowest values of the MTI are 1984, 1988, 1995, 1996, 1998, 1999, and 2007. The monsoon trough index used here (based on the criterion of Wu et al. 2012) is consistent with other definitions used by Molinari and Vollaro (2013). This definition is based on the interannual variability that creates a conceptual separation between the monsoon trough and other tropical disturbances, such as the MJO and tropical waves. The strong and weak monsoon trough years based on this definition are consistent with those based on other definitions (e.g., Molinari and Vollaro 2013). One interesting thing to note is that most of the strong (weak) monsoon trough years chosen in this study are El Niño (La Niña) years. Indeed, the interannual variation of the monsoon trough is a response to ENSO. An El Niño event leads to an increase in relative vorticity over the western central tropical North Pacific and, thus, an eastward extension of the monsoon trough (Wu et al. 2012). Figures 1b and 1c show composites of the 850-hPa winds and OLR for the seven S-MT and seven W-MT years, respectively.

Some noteworthy contrasts of the 850-hPa winds and OLR during S-MT years as opposed to those during W-MT years are obvious from the comparison of Figs. 1b and 1c. The monsoon trough extends eastward during the S-MT but retreats westward during W-MT years. The location of the TC genesis displays a close relation with that of the monsoon trough. More TCs are formed east of 150°E in the S-MT years (Fig. 1b) than in the W-MT years (Fig. 1c). Wu et al. (2012) proposed that an S-MT favors the growth of the wave disturbances and results in more tropical cyclogenesis in the southeast quadrant of the WNP. The time series of the MTI displays year-to-year fluctuations consistent with both TD–MRG wave and ER wave anomalies in the southeast quadrant (5°–20°N, 150°E–180°E) of the WNP (Fig. 1a). The correlation coefficient with TD–MRG (ER) waves is 0.39 (0.61), which
exceeds the 99% confidence level. This indicates that the interannual variations of both the TD–MRG and ER waves are closely related to the location of the monsoon trough. An S-MT event, which is associated with an eastward extension of the monsoon trough, leads to an increase in WTWs to west of the date line.

b. Interannual variability of wave activity and its relationship to the monsoon trough

The wavenumber–frequency spectra of OLR over the WNP domain (0°–20°N, 120°E–180°E) in the S-MT and W-MT years are presented in Figs. 2a and 2b, respectively. Both S-MT and W-MT years capture major spectral peaks in the MRG, TD, ER, MJO, and Kelvin waves, with a similar peak observed in the tropical mean OLR spectrum (e.g., Wheeler et al. 2000; Roundy and Frank 2004). Among the most apparent differences between the S-MT and the W-MT (Fig. 2c) is that the peaks in the OLR spectrum in S-MT years have less power at low wavenumbers for the MRG waves and that the spectral peaks of TDs at low wavenumbers and ER waves at high wavenumbers are broader and have more power.

To examine the interannual variations in the wave activity, the spatial distributions of 1979–2007 mean TD–MRG-filtered EKE are shown in Fig. 3a for July–November. The maximum EKE value appears over the central-to-western Pacific in the upper troposphere (Fig. 3a). In the lower troposphere, the TD–MRG wave activity tends to be enhanced over the region of 120°–160°E. At 850 hPa (Fig. 3c), higher EKE is observed to extend from the tropical central Pacific up to southern China. The displacement of the maximum TD–MRG wave activity follows that of the monsoon trough, roughly centered on 15°N. The pattern of the climatic ER-band EKE (Figs. 3b and 3d) is similar to that in the TD–MRG-band EKE, but the center in the lower
troposphere (600–900 hPa) is particularly strong over the WNP and has a southward shift at 850 hPa (along 10°N).

The distribution of interannual standard deviations in TD–MRG-filtered EKE (Figs. 3a and 3c) shows features that are similar to those of the mean. High values of standard deviation signify a strong interannual variation in the TD–MRG wave activity in the WNP. In the lower troposphere, large standard deviations in TD–MRG-filtered EKE are seen in the monsoon trough region during July–November, and they correspond well with high variability in TD–MRG waves (Fig. 3a) as well as high values of mean TD–MRG waves (Fig. 3c). Compared to that of the TD–MRG wave, the interannual variance of the ER wave (Figs. 3b and 3d) is of a larger magnitude and is shifted to a position centered on 10°N. This location, as well as the maximum value of the climatological ER-filtered EKE, is also similar to the monsoon trough location.

To discuss the relationship between wave activity and the monsoon trough, a correlation analysis is carried out. The vertical cross section of the correlation coefficient between the TD–MRG-filter EKE and MTI is shown in Fig. 4a. There is a strong positive correlation at the lower level located at about 130°–170°E, but with a negative correlation at the upper level. At 850 hPa (Fig. 4c), the TD–MRG wave displays a significant positive correlation east of 130°E over the tropical Pacific, while a negative correlation is located over the South China Sea region and east of the Philippines. For the ER wave (Figs. 4b and 4d), an in-phase relationship is seen between the upper and lower level, and positive correlation covers a large area over the tropical WNP south of ~20°N at 850 hPa.

Previous studies have shown that tropical wave development is dependent on large-scale background flow, and the monsoon trough may be one of the key variables that favor the development or amplification of tropical waves in the WNP. Here, we confirm that TD–MRG waves and ER waves are strongly connected to the geographic extension and interannual evolution of the monsoon trough in the WNP, especially in the lower troposphere. The interannual variation in the location and intensity of WTWs may be anticipated with changes in the monsoon trough condition. In the next section, we focus on investigating the relationship between the interannual variability of the monsoon trough condition and the interannual variability of the three-dimensional structure and evolution of the TD–MRG waves and ER waves in the WNP, respectively.

4. Monsoon trough–dependent WTWs in WNP

a. TD–MRG waves

For further confirmation of the influence of the monsoon trough on the TD–MRG waves, Fig. 5 shows distributions of the composite mean wave-filtered EKE variance during July–November in different states of the
monsoon trough in WNP for the TD–MRG band and the difference between the S-MT and W-MT years. In both cases, the maximum EKE value appears over the central-to-western Pacific in the upper troposphere (Figs. 5a,b). Overall (particularly in the upper troposphere), the TD–MRG-band EKE is weaker during S-MT than W-MT years. The EKE distribution indicates that the TD–MRG wave activity tends to extend downward and develop in the lower troposphere over the region of 130°E–180°. Compared to the W-MT years, waves tend to be enhanced in the lower-to-middle troposphere and suppressed in the upper troposphere to the east of 130°E in S-MT years (Fig. 5c). At 850 hPa (Figs. 5d and 5e), the distribution of variance confirms that EKE anomalies associated with TD–MRG waves often propagate along the near-equatorial region toward the date line and then move northwestward along the monsoon trough. Maximum EKE variance in this band occurs during July–November in the location of the monsoon trough region of the WNP. Accompanying the eastward extension of the monsoon trough (Figs. 5d and 5e), the maximum EKE value in lower levels extends further eastward during S-MT years than during W-MT years. A large area of positive and statistically significant difference in TD–MRG waves covers the tropical central Pacific to the WNP (centered in the region of the monsoon trough extension) (Fig. 5f). As a result, the EKE at 850 hPa is higher and the low-middle-level waves are more active during S-MT years than during W-MT years.

To examine the structure and evolution of the TD–MRG waves in the WNP during S-MT and W-MT years, the filtered OLR and wind fields are regressed onto the time series of the filtered OLR at the reference point (7.5°N, 165°E) with a lagged time from day −2 to day +2. At 850 hPa (Figs. 6a,b), there is a clear wave train with a northwest–southeast orientation. The waves propagate westward and northwestward from the equatorial eastern Pacific to the western North Pacific. Note that the amplitude of the TD–MRG waves is greater during S-MT than during W-MT years. Another distinctive characteristic is seen in the TD–MRG wave evolution. During S-MT years, there is a clear transition of an equatorial MRG wave to a northwest–southeast-oriented TD-type wave train. A statistically significant MRG wave signal (cyclonic disturbance A) moves westward on day −2 over the central Pacific, with two gyres centered on and symmetric about the equator and an antisymmetric OLR signal. Once the waves reach the eastern part of the monsoon trough, the waves move northwestward, and the convection appears to catch up with the vortex and becomes in phase with the vortex center on day 0. This disturbance A displays a TD wave structure by the time its center moves off the equator to about 5°N. The anticyclonic disturbance B, on the equator farther east, has an MRG structure at this time. Two days later, the disturbance B also shows a fairly clear TD-type signature. The transition process is remarkably similar to that in observations in Dickinson and Molinari (2002).

Table 1 confirms the wave transition as wave disturbances propagate through the monsoon trough. Wavelength shrinks from 4900 to 3800 km and the phase speed of the TD–MRG wave derived from OLR keeps roughly
at 9.6 m s⁻¹, while the phase speed of the TD–MRG wave derived from wind decreases from 9.6 to 6.4 m s⁻¹ between lags −2 and 2 days. The propagation speed of tropical waves could be strongly affected by the interaction with atmospheric circulation. Han and Khouider (2010) found that the westerly shear tends to decelerate westward-moving tropical waves via inducing a Doppler shift. So the background flow of the monsoon trough has a strong impact on the slowdown process of waves. Moreover, the convection also plays significant roles in modulating the propagation speed of tropical waves so that slower propagation would be expected in regions of higher convective activity (with broad convective), such as in the western Pacific (Dias and Pauluis 2009, 2011). Therefore, the waves must slow down. However, the wave structure change with convergence in a cyclonic vortex leads to the speed of wave convection being larger than the phase speed of the wave eddy as discussed further below. The decelerated propagation leads to a shift of the negative OLR anomalies from lying 7.5–8° east of the cyclonic vortex center to near the cyclonic vortex center (Fig. 6a). This suggests the transition of an MRG wave to a TD-type wave train along the monsoon trough from day −2 through day +2. In contrast, such a transition is vague during W-MT years (Fig. 6b). The longitudinal zone for transition is shifted to the east, consistent with the eastward extension of the monsoon trough during S-MT years (Chen and Weng 1998; Clark and Chu 2002). Both the cyclonic vorticity (Wang and Chan 2002) and the zonal confluence (Kuo et al. 2001; Tam and Li 2006) of the monsoon low-level flow during the S-MT years may contribute to the growth of the TD perturbations. The zonal scale of the TD–MRG waves is contracted as they propagate westward. Because of the westward retreat of

The vertical structure of the TD–MRG waves during S-MT and W-MT years is shown in Figs. 6c and 6d. In these figures, the meridional wind field is regressed onto the filtered OLR time series at 7.5°N, 165°E. A wave train is clearly identified throughout the troposphere. The maximum amplitude of the meridional wind is located in the middle troposphere (near 400–600 hPa). A marked difference between S-MT and W-MT years lies in the tilting of the vertical axis of the meridional wind field. It tends to be barotropic during S-MT years (Fig. 6c) and becomes tilting eastward with height during W-MT years (Fig. 6d). This tilt with height is consistent with the vertical structure of the classical MRG waves shown in previous analyses (e.g., Wheeler et al. 2000; Kiladis et al. 2009). This difference of vertical structure for different monsoon trough years is a response to the transformation from MRG to TD waves.

To further clarify the wave transition, Figs. 7a and 7c (Figs. 7b and 7d) show longitude–time diagrams of the 850-hPa EKE of TD–MRG waves along 20°N (2.5°N) for S-MT and W-MT years, obtained by regression on the OLR variation of the TD–MRG band at 20°N, 120°E (2.5°N, 175°W). Figures 7b and 7d show that the TD–MRG waves propagate from the eastern Pacific and terminate over the WNP. In comparison, the wave decay occurs a few thousand kilometers farther to the east during S-MT than W-MT years. The new waves form over the WNP, and their phase speed is estimated to be somewhat slower than the original waves (Figs. 7a,c), consistent with the differences in Fig. 6 and Table 1. This suggests that the TD–MRG waves may be affected by the monsoon trough.
Fig. 6. Lagged regression of the TD–MRG-band-filtered 850-hPa wind (vector; m s$^{-1}$) and OLR (shaded) fields (regressed based on the filtered OLR time series at 7.5°N, 165°E) during the TC season for (a) S-MT years and (b) W-MT years. (c), (d) Plot of a longitude–height cross section of the regressed meridional wind anomaly averaged over the latitudes of 5°–10°N for S-MT and W-MT years. Only wind and OLR signals that exceed the 95% significance level are shown. The shaded interval for OLR is 1.5 W m$^{-2}$ with negative shaded blue.
The wave transition in the monsoon trough region is an interesting feature of the Pacific TD–MRG waves. To illustrate the wave transition, Fig. 8 shows the lagged regression of anomalies of the 850-hPa streamfunction, column-integrated Q1, and 850-hPa divergent wind vectors for both the S-MT and W-MT years. The streamfunction anomalies during both the S-MT (Fig. 8a) and W-MT (Fig. 8b) years indicate a pronounced east–west-oriented wave train in the WNP from the date line to east of the Philippines, with the anomalies aligned along the axes of the climatological mean monsoon trough, consistent with the wave train in Fig. 6. While wave structures at the lower levels can act as a forcing agent for convection, the convection itself can modify the structures of the theoretical modes (Yang et al. 2009). As expected, the Q1 anomaly pattern is very consistent with the OLR anomaly (Fig. 6) as well as the 850-hPa divergent wind fields. This vertical structure suggests a response to the convective heating. This figure also highlights a transition of the horizontal structure of the streamfunction to a northeast-to-southwest tilt in the vicinity of the monsoon trough. During the structure change, the centers of divergent wind and Q1 are organized in correspondence with the center of the streamfunction. Lower-tropospheric wave structures with convergence in a cyclonic vortex center tend to result in the convection filling the cyclonic vortex center in the monsoon trough region and the lower-level wave showing a clear TD-type disturbance structure. In contrast, this structure change during W-MT years is relatively weak and has a westward movement of 15°–20° longitude while the monsoon trough retreats westward. This wave transition from one to another is thought to be aided by barotropic deformation in the confluence and shear region of the monsoon trough, where the horizontal structure of a northeast-to-southwest tilt and the amplitude increasing may be strengthened by a positive feedback between the wave

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<th>Table 1. Synoptic-wave characteristics for a disturbance of TD–MRG waves propagating through the monsoon trough at lags −2, 0, and 2 days.</th>
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<tr>
<td><strong>Disturbance A (S-MT years)</strong></td>
</tr>
<tr>
<td>Lag = −2 days</td>
</tr>
<tr>
<td>Phase speed (vortex; m s⁻¹)</td>
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<tr>
<td>Phase speed (OLR; m s⁻¹)</td>
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<td>Wavelength (km)</td>
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<td>Vorticity and OLR center mean distance (°lon)</td>
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![Fig. 7. A longitude–time diagram of EKE (m² s⁻²) of 850-hPa winds for the TD–MRG waves along 20°N associated with the OLR variation of the TD–MRG waves at the base point 20°N, 120°E during the TC season for (a) S-MT years and (b) W-MT years. (b),(d) As in (a),(c), but along 2.5°N and with base points at 2.5°N, 175°W.](http://journals.ametsoc.org/jcli/article-pdf/28/18/7108/4053692/jcli-d-14-00806_1.pdf)
and the monsoon trough. This is supported by other studies (e.g., Dickinson and Molinari 2002; Kiladis et al. 2009; Wu et al. 2014a) and will be addressed further in Part II.

The vertical structures of TD–MRG divergence and Q1 are shown in Fig. 9. Lower-tropospheric TD–MRG divergence and Q1 display an eastward tilt from the surface to an elbow around 600 hPa to the east of the monsoon trough, and its divergence center in the lower levels leads slightly the maximum Q1 center at mid-levels. Once the TD–MRG wave reaches the region of monsoon trough, it exhibits large differences in the vertical structure of divergence, with opposite sign in the lower and middle-upper troposphere. Consistently, the Q1 has the same sign throughout the troposphere. Associated with this structure of Q1 and convergence, the TD–MRG vertical motion extends throughout the troposphere and coincides with the Q1 structure (Figs. 10a,b).

At longitudes of the cyclonic vortex center (Figs. 10c,d), enhanced Q1 heating is observed at day 0 over the latitudes of about 0°–10°N from the surface up to about 100 hPa. During the S-MT years, the strong Q1 heating is located in the middle troposphere (near 300–600 hPa), and significant upward motion penetrates throughout the whole troposphere along the Q1 heating (Fig. 10c). In contrast, the signals are weaker, and the maximum center is located in the upper troposphere (near 200–400 hPa) during W-MT years (Fig. 10d). Lower-level convergence and associated upward motion and convective heating (Q1) provide further evidence for the systematic evolution of TD–MRG wave structures in the monsoon trough region. The lower-tropospheric wave structure change associated with the monsoon trough may play an important reorganizing role for convection. The region of upward motion and convective heating is in phase with the region of low-level convergence and middle-upper-tropospheric divergence (Fig. 9a), which favors the development of a barotropic structure for the TD wave. In contrast, lower-tropospheric convergence slightly leads the position of convective heating, which leads to its vertical tilted
ascent signal during W-MT years. The current research suggests that the convection is close to being in phase with the vertical motion and low-level convergence in the monsoon trough region. On the other hand, the heating associated with this convection is one of the most important energy sources for the waves (Yasunaga and Mapes 2014). Therefore, the lower-tropospheric wave dynamical structure is further forced by such vertical structures of convection. A barotropic structure of the TD wave appears to be forced by interaction between the wave structure and convection in the monsoon trough region.

For the westward-propagating vortex and convection of the MRG waves, combined with monsoon trough convergence, linear theory would predict that their phase speeds would become increasingly slower, their amplitude would increase, their scale would shrink (wavelength shorten), and their structure would display a southwest–northeast tilt because of the forced nature of the monsoon trough mean flow (see Part II). This wave structure change (scale shrink and horizontal tilt) tends to lead to convergence in the cyclonic vortex region of the lower troposphere. While the vertical motion is in phase with the convergence at the lower levels, it shifts to a cyclonic vortex center ahead of the convection. The enhanced and shifted Q1 heating associated with vertical motion change leads the convection to catch up with the vortex, and, thus, the convection tends to fill the cyclonic vortex center. At last, the wave appears to have a TD-type disturbance structure: the maximum Q1 heating is in the middle troposphere, there is significant upward motion, and the convection is organized in correspondence with a cyclonic vortex. As the TD disturbances provide an important source for tropical cyclogenesis, this explains why WNP TCs are more active within the southeastern quadrant of the WNP during the S-MT than during the W-MT.

b. ER waves

The previous section showed the structure and evolution of TD–MRG waves in the WNP. In this section, we provide a similar evaluation of the ER waves and assess the statistical relationship between the ER waves and the monsoon trough. Figure 11 presents the distributions of the ER wave-filtered EKE variance during July–November during different states of the monsoon trough in the WNP and the difference between the S-MT years and W-MT years. As with the TD–MRG waves, the maximum EKE appears over the central-to-western Pacific in the upper troposphere (Figs. 11a,b). At 850 hPa, the major occurrence of the ER wave is located north of the equator in and near the expected position of the monsoon trough in the WNP and the difference between the S-MT years and W-MT years. As with the TD–MRG waves, the maximum EKE appears over the central-to-western Pacific in the upper troposphere (Figs. 11a,b). At 850 hPa, the major occurrence of the ER wave is located north of the equator in and near the expected position of the monsoon trough in the WNP and the difference between the S-MT years and W-MT years.

FIG. 9. Vertical profiles of the regressed TD–MRG heat source Q1 ($10^{-3} \text{W m}^{-2}$; shading) and divergence ($10^{-7} \text{s}^{-1}$; contours) of the regressed horizontal wind anomalies for (a) S-MT and (b) W-MT years along 7.5°N.
variation of the monsoon trough location (Figs. 11c,f). Compared to that of the TD–MRG wave (Fig. 5), it displays a larger magnitude and smaller horizontal tilt.

Figure 12 shows the evolution from day $-6$ to day $+6$ of the horizontal and vertical structure of ER waves regressed onto the time series of the ER-filtered $-\text{OLR}$ at the reference point (7.5°N, 165°E). The pair of cyclones and the accompanying significant convective signal propagate westward at a phase speed of around 4.8 m s$^{-1}$ (Fig. 12a). OLR anomalies are located to the east of the circulation cell. The structures and phase speeds of the observational ER waves are in good agreement with the theoretical structure of an $n = 1$ ER wave. During S-MT years, the vortex’s phase speed reduces from the eastern portion of the monsoon trough, and the convection’s speed keeps at 4.8 m s$^{-1}$ (Table 2). The convection (maximum negative OLR anomalies) shifts from lying 10° east of the cyclonic vortex center on day $-6$ to 5° east of the cyclonic vortex center on day $+6$. Such phase speed change may be related to the growth of the ER wave amplitude and the shrinkage of the wavelength. Unlike the TD–MRG waves, the ER waves do not show a shift in the structure. Vertical cross sections of the ER waves are shown in Fig. 12c. The ER waves have a barotropic structure from the middle to lower troposphere (from 300 to 1000 hPa) and features opposite anomalies at the upper level.

Figures 12b and 12d show the evolution of ER waves during W-MT years. In contrast to the S-MT years, the ER waves during W-MT years become weaker, and the traveling vortex is located farther west and north on day $-6$. From day 0 to $+6$, the vortex center maximum lies somewhat farther north of the convection signal, along 17.5°N, but the OLR signals associated with ER waves are still along 7.5°N and propagate westward at phase speeds of around 4.8 m s$^{-1}$. Unlike S-MT years, the vortex signals propagate at nearly the same phase speed of around 3.2 m s$^{-1}$ from day $-6$ to $+6$, but the wave does not appear to slow down in the monsoon trough region. As the ER waves move westward, their wavelengths also contract, and the maximum negative OLR anomalies shift from lying 20° to 10° east of the cyclonic vortex center. The vertical structure of the wave is not so apparent on day $-6$. This discrepancy is possibly because the ER wave train is more poleward before it reaches the monsoon trough region. However, in the later stages from day 0 to $+6$, the intensified ER waves display a clear barotropic structure at the middle and lower troposphere.

In contrast to the TD–MRG waves, a longitude–time diagram of ER-band 850-hPa EKE regressions along 7.5°N (not shown) indicates that the ER waves display broadly similar features in S-MT and W-MT years, but the ER waves in S-MT years begin to enhance from more eastward locations (about east of the date line). This confirms that the structures have no significant shift.
as ER waves propagate westward through the monsoon trough. Those two westward-moving waves show very different behaviors resulting from their interaction with the monsoon trough. In particular, the growth rates of the ER waves are much larger than those of the TD–MRG waves. In the lower troposphere, the ER waves extract more barotropic energy from the background flow of the monsoon trough than do the TD–MRG waves; this can be attributed to the difference in the amplitudes of those waves as the wave amplitudes grow larger. It is interesting to note that the ER waves show a smaller horizontal tilt. This may be attributed to the broad spatial scale (e.g., Kiladis et al. 2009) and convection forcing associated with more baroclinic energy from the basic flow of the monsoon trough. Those traits will be further examined in Part II and further study.

Figure 13 shows the evolution of the 850-hPa streamfunction, integrated Q1, and 850-hPa divergent wind in both the S-MT and W-MT years. The streamfunction centers show a clear wave train, consistent with wind in Fig. 12. The wave circulation cells move westward with the Q1 and divergence anomaly centers, which coincides with OLR centers, to take their place to the east of the circulation centers. Vertical cross sections of the integrated Q1 and divergence anomalies are shown in Fig. 14. The divergence displays an eastward tilt at the lower levels with a secondary divergence maximum center near 900 hPa, and an opposite sign appears at upper levels with a maximum divergence center near 200 hPa. This divergence signal is accompanied by the Q1 anomalies that have a slight eastward tilt with height and the same sign through the whole troposphere. In response to the Q1 and divergence, a significant vertical motion penetrates throughout the troposphere along the Q1 heating with a slight eastward shift with height to 300 hPa (Figs. 15a,b). At longitudes of the cyclonic vortex center (Figs. 15c,d) associated with ER waves, the Q1 heating maximum is observed over the latitudes of 0°–10°N through the troposphere. During the S-MT years, strong Q1 heating is located in the middle troposphere (near 300–600 hPa), and obvious upward motion is seen along the Q1 heating (Figs. 15a,c). In contrast, the Q1 heating and the upward motion are of the same sign through the troposphere, but the signals are weaker, and the maximum center is located in the middle-to-lower troposphere (near 200–700 hPa) during W-MT years (Figs. 15b,d). The lower-tropospheric divergence structure ahead of the middle-tropospheric convective heating may cause the eastward shift and vertical tilt of the convective heating in the monsoon trough region, leading to the eastward movement of the convection and deeper convection. This implies that the lower-tropospheric background flow of the monsoon trough could provide favorable background conditions for redevelopment of the lower-tropospheric ER waves, which could help organize the wave’s horizontal and vertical coupling structures.

5. Summary and discussion

This study examined the monsoon trough–dependent structure and evolution of two types of westward-propagating synoptic disturbances: TD–MRG waves and ER waves in the WNP using the observed OLR field and NCEP-2 datasets over the 1979–2007 period. It is found that both TD–MRG and ER wave activity have a distinct signal of interannual variations, which is closely related to the location and intensity of the monsoon trough. During the S-MT (W-MT) years when the monsoon trough extends eastward (retreats westward), more WTW activity occurs to the east (west) of 140°E. Meanwhile, those westward-moving waves show very different structures and relationships with the convection in the S-MT and W-MT years, responding to the
Fig. 12. As in Fig. 6, but for the ER waves.
changes in the monsoon trough variability in this region. When the monsoon trough enhances and extends eastward to around 160°E, the TD–MRG and ER wave activities are above normal along the monsoon trough region at the lower troposphere over the most part of the WNP, and slightly fewer wave activities tend to occur near the south of the Philippines and the South China Sea. There is evidence that the MRG and ER waves in the WNP are often modified by their dynamical response to the monsoon trough, but different waves have been found to have different evolutions.

The TD–MRG mode exhibits a clear MRG-to-TD-type wave transition during S-MT years. In contrast, the transition is not clear, and the TD-type disturbance is much weaker during the W-MT years. They show an equivalent barotropic structure during S-MT years, but tilt eastward with height during W-MT years. This wave transition could be because the circulation pattern of TD–MRG waves appears contracted and displays horizontal tilts as they move westward and interact with the low-level flow of the monsoon trough and the divergent wind center begins to fill the vortex center. This

<table>
<thead>
<tr>
<th>Disturbance A (S-MT years)</th>
<th>Disturbance C (W-MT years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lag = −6 days</td>
<td>Lag = 0 days</td>
</tr>
<tr>
<td>Phase speed (vortex; m s⁻¹)</td>
<td>4.8</td>
</tr>
<tr>
<td>Phase speed (OLR; m s⁻¹)</td>
<td>4.8</td>
</tr>
<tr>
<td>Wavelength (km)</td>
<td>8200</td>
</tr>
<tr>
<td>Vorticity and OLR center mean distance (° lon)</td>
<td>10</td>
</tr>
</tbody>
</table>

FIG. 13. As Fig. 8, but for the ER waves.
divergence change causes a shift of the vertical motion and convective heating into the vortex center, and the wave structure displays characteristics of TD-type disturbances. It is suggested that the monsoon trough may play an important role in the wave transition, which then reorganizes the structure of convective and dynamical fields and modulates the spatial relationships between them through the interaction with TD–MRG waves. The wave transition tends to slow the phase speed as waves propagate westward through the monsoon trough but leads to an increase in the wave amplitude and a reduction of wave-scale contraction. The evidence in this research and in Dickinson and Molinari (2002) suggests that, in many cases, TD waves may have their origin in MRG waves, though not all TD wave trains are related to the transition of MRG waves. Previous studies show roughly 40% of TD waves, which are associated with tropical cyclogenesis, have been referred to as a transition of MRG waves to off-equatorial TD-type disturbances (Fu et al. 2007).

In contrast to the TD–MRG waves, the amplitude of the ER waves appears significantly dependent on the state of the monsoon trough, which is similar to that shown in TD–MRG waves. The influence of the monsoon trough on ER waves varies between different states. However, unlike the TD–MRG waves, their coupling structures between convection and the circulation have no marked change as waves propagate westward through the monsoon trough and the wave circulation cells move westward with the vertical motion and convective heating anomaly centers to take their place to the east of circulation centers. The vertical structure of the ER mode shows a slight eastward tilt with height during both S-MT and W-MT years. It is interesting to note that the ER amplitude has a stronger increase as waves interact with the monsoon trough than TD–MRG waves, but the wave transition is actually more significant in the TD–MRG waves. This suggests that the wave amplitude increase is not sufficient for the wave transition, as the mechanism for evolution of waves may be different in the two different wave structures.

Observations presented in this paper provide an insight into the structure and evolution of the TD–MRG and ER waves as they propagate westward through the monsoon trough. It is suggested that the monsoon trough favors redevelopment of the wave disturbances, which plays an important role in reorganizing the wave structure. Previous studies (Sobel and Bretherton 1999; Done et al. 2011; Kuo et al. 2001) have suggested that the waves propagating through a wave accumulation zone with confluent background flow shrink, slow down,
and become more energetic in the lower troposphere. The WNP background flow favors the disturbances extracting barotropic energy from the background flow (Maloney and Dickinson 2003; Serra et al. 2008). Our accompanying work indicates the importance of the lower-level convergence and cyclonic environmental shear of the monsoon trough for growth and reorganizing of lower-level wave disturbances in the WNP region through a positive feedback between the wave growth and horizontal structure based on the barotropic energy conversion. During the S-MT years when the monsoon trough extends eastward, more energy extracted from the lower-level background flow of the monsoon trough leads to an apparent transition from MRG waves to off-equatorial TD disturbances and a faster growth of ER amplitudes in the monsoon trough region. Those different influences of the monsoon trough on TD–MRG and ER waves can be attributed to different structures of two waves. In contrast, such a transition from MRG to TD waves and an amplitude growth of ER waves are vague during W-MT years, which could be because of the weak energy conversion and westward retreat following the monsoon trough. In addition to future documentation of the dynamics of WTWs, it will be important to explore in detail in Part II how the basic state of the monsoon trough modulates WTW variability over the WNP.

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