How Do the Monsoon Trough and the Tropical Upper-Tropospheric Trough Affect Synoptic-Scale Waves: A Comparative Study

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Abstract

Using the Climate Forecast System Reanalysis, Joint Typhoon Warning Center best track, and Tropical Rainfall Measuring Mission precipitation data, two long-lasting synoptic-scale wave trains in 2004 and 2006 are selected to investigate the atmospheric factors controlling the structures of westward-propagating synoptic-scale disturbances over the tropical western North Pacific. The essential difference between these two wave trains is found in their vertical structures. In 2004, the maximum perturbations occurred from the middle to lower troposphere with an equivalent barotropic structure; however, in 2006, they primarily occurred in the upper troposphere with a prominent tilt regarding height. Distinct configurations of the monsoon troughs, the tropical upper-tropospheric troughs (TUTT), and associated vertical wind shear caused such structural differences. In 2004, the TUTT shifted eastward, creating an easterly sheared environment to confine synoptic-scale waves in the lower troposphere. Then, the monsoon trough enhanced the wave activity through barotropic energy conversion in the lower troposphere. In contrast, while the TUTT shifted westward in 2006, synoptic-scale waves prevailed in the upper troposphere by the environmental westerly shear. Meanwhile, the disturbances developed in the upper troposphere through to the conversion of kinetic energy from the TUTT, exhibiting a top-heavy vertical structure. The coherent movement of the monsoon trough and the TUTT modulate the vertical structure and the development of the synoptic-scale waves.
1. Introduction

Tropical synoptic-scale waves, which are sometimes called tropical-depression (TD)-type waves, are a series of disturbances with alternating cyclonic and anticyclonic perturbations during the boreal summer over the western North Pacific (Wallace and Chang 1969; Chang et al. 1970; Reed and Recker 1971; Takayabu and Nitta 1993; Dickinson and Molinari 2002). A typical synoptic-scale disturbance possesses a wavelength of approximately 2,500 kilometers, a period of 3–8 days, and a propagation direction from southeast to northwest. These disturbances feature a less-tilted vertical structure with the maximum wind perturbation appearing in the mid-lower troposphere (Reed and Recker 1971; Wallace 1971; Lau and Lau 1990; Tam and Li 2006).

The intensity and structure of the synoptic-scale waves are significantly affected by lower-frequency ambient flow in the lower troposphere, i.e., the monsoon trough over the western North Pacific. The kinetic energy of these westward-propagating waves can accumulate in the lower-tropospheric monsoon trough, inducing a shrinkage in the horizontal scale of the wave (Webster and Chang 1988; Sobel and Bretherton 1999; Done et al. 2011). Moreover, lower-tropospheric synoptic-scale disturbances are enhanced through barotropic energy conversions near the monsoon trough where the absolute vorticity exhibits a maximum (Maloney and Hartmann 2001; Wu et al. 2012; Feng et al. 2014).

The vertical structure of these disturbances is regulated by the background vertical wind shear. Observational studies indicated that the perturbation maxima of the synoptic-scale waves occurred in the upper troposphere over the central Pacific and in the mid-lower troposphere over the western Pacific (Reed and Recker 1971; Tam and Li 2006; Serra et al. 2008), and these studies attributed the systematic variation in the vertical structure to the zonal changes in the vertical shear of the basic zonal wind. Theoretically, the zonal vertical wind shear can modulate the phase difference between the barotropic and baroclinic components of the tropical synoptic-scale waves (Wang and Xie 1996), and this shear feeds the mean available potential energy to waves by generating an Ekman pumping-induced heating and meridional heat flux (Xie and Wang 1996). As a result, these waves will be trapped in the lower troposphere in an easterly sheared environment and in the upper troposphere in a westerly sheared environment. Previous studies also demonstrated the critical role of vertical wind shear in the structure and evolution of convectively coupled equatorial waves (e.g., Holton 1971; Molinari et al. 2004; Han and Khoury 2010).

During boreal summer, the distribution of the vertical wind shear over the western North Pacific is controlled by the displacements of the lower-tropospheric monsoon trough and the tropical upper-tropospheric trough (TUTT). A stronger TUTT brought upper-tropospheric westerly anomalies to lower latitudes, and this demonstrates the ability to cause a change in the environmental vertical wind shear in the tropics. Gray (1968) proposed that the TUTT created a large amount of westerly shear in the tropics, subsequently inhibiting the development of tropical cyclones (TCs) over the western North Pacific. However, the impact of the upper-level background circulation on tropical synoptic-scale disturbances has been less discussed; although, some studies revealed the influence of the TUTT on TC formations over the western North Pacific (e.g., Sadler 1976, 1978; Briegel and Frank 1997; Ritchie and Holland 1999; Chen et al. 2008; Wang and Wu 2016). The plausible reason is that the synoptic-scale disturbances propagate northwestward over the equatorial western North Pacific, and the TUTT is usually observed north of 20°N. Observations show that the vertical structure of these synoptic-scale disturbances features a northward tilting with height (Feng et al. 2016). Therefore, the upper-level counterparts of the TD-type waves propagate westward along 20°N over the central Pacific (Feng et al. 2016; see also Fig. 7 in Tam and Li 2006). The TUTT favors upper-tropospheric development of the synoptic-scale disturbances through barotropic energy conversion. These enhanced upper-level perturbations may serve to initiate a lower-tropospheric disturbance through downward momentum transition (Tam and Li 2006).
In summary, the lower-tropospheric monsoon trough, the TUTT, and the background wind shear affect the synoptic-scale wave activity over the western North Pacific, among which the background vertical wind shear is regulated by the other two systems. Nevertheless, few studies discussed how the co-variation of the monsoon trough and the TUTT affect the structure and evolution of synoptic-scale waves. This paper is a comparative study, aiming to investigate the co-effect of the monsoon trough and the TUTT on the synoptic-scale waves. Two long-lasting synoptic-scale wave trains, occurring during 2004 and 2006, were selected for comparison. The present study focused on long-lasting events rather than on those with a short life period based on the consideration that long-lasting events indicate a relatively stable large-scale environment. Additionally, the changes in wave structure and the environment can be identified explicitly. Moreover, examining long-lasting events to understand multiple TC events is beneficial (Schenkel 2016; Hu et al. 2018). The long-lasting wave train in 2004 was reported by Hu et al. (2018), and they found that the long-lasting wave packet induced abnormal multiple TC geneses. They attributed the cause of this event to the abnormal oceanic thermal state in the equatorial region. However, the structure of the disturbances in 2004 was discussed less. This study compares the structures of the disturbances in 2004 and 2006. Then, the possible mechanisms that affect the distinct structures will be investigated.

The rest of this paper is organized as follows: Section 2 describes the data and methodology. Section 3 reveals the characteristics of these two long-lasting synoptic-scale wave trains and investigates their differences. Section 4 discusses the critical factors affecting the structures and evolutions of these two wave trains. Finally, conclusions and discussion are presented in Section 5.

2. Data and methodology

2.1 Data

This study used the NCEP Climate Forecast System Reanalysis (CFSR) (Saha et al. 2010) and subsequent CFSR version 2 datasets (Saha et al. 2014), with a horizontal resolution of 0.5° × 0.5° and a temporal resolution of six hours, to map three-dimensional wind fields and their evolutions. The CFSR dataset is considered to be one of the best reanalysis datasets for describing TC activities, providing excellent mean TC intensities and numbers, the highest scores in terms of the climatological frequencies of TC occurrence, excellent hitting rates, and low false alarm rates (Murakami 2014). Best track data from the Joint Typhoon Warning Center (JTWC) are used to provide information about the genesis times and locations of TCs. The first occurrence of a sustained 1-min averaged surface wind reaching 25 knots is defined as the genesis time. In addition, the 3-hourly precipitation rate data derived from the Tropical Rainfall Measuring Mission (TRMM) Multi-satellite Precipitation Analysis (Huffman et al. 2007), which are available at a 0.25° × 0.25° spatial resolution, are also used to represent the rainfall embedded with the synoptic-scale disturbances.

2.2 Methodology

To separate the synoptic-scale perturbations from the original fields, a bandpass-filtering method (Duchon 1979) is adopted. By applying a 3–8-day bandpass filter to the horizontal wind fields, the synoptic-scale wave activities can be isolated from the original wind fields. The environmental winds are adopted by using a 10-day low-pass filter. Thus, the effect of intraseasonal oscillations, e.g., the Madden–Julian oscillation (MJO), is included in the large-scale environment. Note that the wind fields of a mature TC contribute to the variances of 3–8-day waves. Although various TC isolation methods were proposed in previous studies, separating the TC-related wind fields from the original using a completely objective technique remains difficult. However, the general wave characteristics change little except for the wave intensity when the mature TCs are removed using the TC-removing method of Kurihara et al. (1995), as Xu et al. (2013) and Feng et al. (2016) stated. Consequently, no TC-removing algorithm is applied in this study.

2.3 Case selection

This study requires the selection of two long-lasting synoptic-scale wave events with different structures. The wave trains were first identified by the 3–8-day meridional winds at 850 hPa in the Hovmöller diagrams from 2000 to 2015 (figures not shown). The duration of each wave train was calculated after the determination of the development and decay dates using a subjectively defined criterion of 1.5 standard deviations in averaged meridional wind anomalies (Fig. 1). Six events with a duration longer than 30 days were identified occurring in years 2001, 2004, 2006, 2009, 2013, and 2014. Then, longitude-height cross-sections of eddy kinetic energy (EKE) averaged for 10–20°N were drawn. Examining the Hovmöller
diagrams and the longitude-height cross-sections of EKE, the cases in 2004 and 2006 were selected for the following reasons: 1) the wave packets in these two years were less shifted over the western North Pacific (Fig. 2); thereby, the change in the wave structure can be attributed to the variation in large-scale background; 2) their vertical structures differ remarkably, which will be shown later; and 3) the cases in 2004 and 2006 occurred in August and July, respectively, and they involved analogous backgrounds of the western North Pacific summer monsoon. The 2004 event began on 28 July and lasted for 46 days until 11 September, and the 2006 event began on 6 July and ended on 7 August.

3. Differences between the two synoptic-scale wave trains

3.1 Overview

The Hovmöller diagram of the 3–8-day filtered meridional winds at 850 hPa illustrates the overall features of the two synoptic wave trains (Fig. 2). The wave train originated near 120°E due to the combined effects of the Rossby energy dispersion of the preexisting TC “Namtheum” and a series of westward-propagating perturbations (not shown). Nearly no remarkable changes in the amplitude of the synoptic-scale perturbations were observed during late-July and early August. During early August, the wave packet shifted slightly eastward. Then, the wave packet began to intensify remarkably and persisted over the western North Pacific until 11 September, when the amplitude of the wave packet weakened rapidly and drifted eastward to the central Pacific.

In 2006, the origination of this wave packet can be traced back to a westward-propagating MRG-like disturbance near 125°E on 23 June (not shown) (Aiyyer and Molinari 2003; Chen and Huang 2009; Chen and Tam 2012; Wu et al. 2014). Then, the wave packet weakened slightly without a remarkable zonal propagation and remained near 140°E until it drifted eastward and was gradually weakened after early August.

Table 1 shows the wave parameters representing the propagations of each synoptic-scale wave train. The wave train in 2004 featured a typical zonal wavelength of 2,400 km, a period of 4.5 days, a westward phase velocity of −6.2 m s$^{-1}$, and a wave package that shifted eastward with an averaged group velocity of approximately 1.0 m s$^{-1}$. In contrast, the event in 2006 demonstrated an identical zonal wavelength of 2,400 km, a slower phase velocity of −4.5 m s$^{-1}$, an averaged eastward group velocity of 0.5 m s$^{-1}$, and a longer period of 6.2 days. The difference in the zonal group velocity was partly attributed to the background zonal wind, as the background zonal wind speed was 2.54 m s$^{-1}$ in 2004 and is 0.39 m s$^{-1}$ in 2006. More intensive westerly wind caused a more eastward shift of the wave packet in 2004. Moreover, the Doppler-shifted group velocity of equatorial waves is also calculated following the dispersion relation of Matsuno (1966).
The frequency-wavelength relation of the synoptic-scale waves does not match any of the theoretical dispersion relations of dry equatorial wave modes. We speculated that the coupled deep convection modulated the dispersion relation in these synoptic-scale waves.

The evolution of these synoptic-scale wave trains was further investigated by analyzing a sequence of synoptic maps (Figs. 3a–f, 4a–f). In both years, series of cyclonic-anticyclonic disturbances propagated northwestward, and this is closely related to the Rossby energy dispersion from mature TCs (Holland 1995; Li and Fu 2006; Fu et al. 2007; Ge et al. 2007, 2008; Krouse et al. 2008). For the 2004 case, after late August 2004, the wave axis was displaced more zonally than before, and the horizontal scales of each disturbance were smaller (Figs. 3a–c). The shrinkage of the zonal wavelength may be associated with the wave energy accumulation by intensified zonal convergence of the ambient flow (Webster and Chang 1988). The convective rainfall embedded in the cyclonic disturbances in 2006 (Fig. 4) was relatively inactive and less organized compared to 2004. Less organized convections in 2006 imply a weaker coupling between the wave circulation and deep convections, which may be attributed to the background vertical wind shear. An easterly shear destabilizes convectively coupled waves and favors a strong convective coupling, whereas a

**Table 1.** General parameters and background zonal wind of the two long-lasting synoptic-scale wave trains. The background zonal wind is calculated as the zonal wind at 850 hPa averaged by 8–15°N, 120–170°E for the 2004 event, and by 8–15°N, 120–160°E for the 2006 event.

| Year | Wavelength (km) | Period (day) | Phase velocity (m s\(^{-1}\)) | Group velocity (m s\(^{-1}\)) | Background zonal wind (m s\(^{-1}\)) |
|------|----------------|--------------|-------------------------------|-------------------------------|-----------------------------------|
| 2004 | 2400           | 4.5          | −6.2                          | 1.0                           | 2.54                              |
| 2006 | 2400           | 6.2          | −4.5                          | 0.5                           | 0.39                              |
The role of the background vertical wind shear will be shown later.

The TC activities are also shown in Figs. 2–4. Generally, seven cyclonic disturbances induced seven TC formations during the active period in 2004, but eight cyclonic disturbances induced six TC formations in the 2006 event. The cyclonic disturbances “F” and “L” in 2006, which exhibit a shrinking zonal scale in the lower troposphere, did not result in a TC development.
3.2 Vertical structures

Figure 5 shows the horizontal distribution of the averaged EKE (defined as $K' = \frac{1}{2}(u'^2 + v'^2)$) at 850-hPa. The region with large EKE values (greater than 2.0 m$^2$ s$^{-2}$) stretched from the East China Sea to the central-western North Pacific with the maximum occurred near Ryukyu Island. The northwest-southeast distribution of EKE indicates a storm track for the synoptic-scale disturbances. In 2006, the regions with large EKE values also stretched southeastward from the ocean near Taiwan to near 142°E (Fig. 5b). The
maximum EKE occurred near 18°N 128°E. The solid red lines in Figs. 5a and 5b denote the averaged axes of the synoptic-scale wave trains. The axis in 2006 was more southwestward than that in 2004, indicating that the synoptic-scale waves occurred more southwestward in 2006.

Vertical structures of averaged EKE are different in these two years, as illustrated in Figs. 5c and 5d. In 2004, the large EKE values appeared from the middle to the lower troposphere from 400 hPa to 925 hPa to the west of 140°E (Fig. 5a), which suggested a bottom-heavy structure. The EKE values gradually decreased from the west to the east in the lower troposphere. In contrast, in 2006, the maximums of the EKE values appear in two layers west of 140°E: one at the lower troposphere, with a maximum near 900 hPa, and the other at the upper troposphere, with a maximum near 200-hPa (Fig. 5b). Two-dimensional E-vectors in the lower and upper troposphere show wave group energy propagated southeastward in the lower and upper troposphere in 2006 (Supplement 1). However, in 2004, the Rossby wave energy dispersion occurred in the middle-lower troposphere, and weaker energy dispersion occurred at the upper-level. Moreover, the low-level disturbances are related to the downward energy propagation from the upper-level in 2006, as indicated by the meridional eddy transport of heat (Supplement 2). These characteristics indicate the different propagation characteristics of the wave trains: in 2004, the disturbances primarily propagated in the middle-lower troposphere; in 2006, the disturbances propagated in the upper and lower troposphere simultaneously.

Figures 6 and 7 show the evolution of the two synoptic-scale waves and their vertical structures using longitude-height maps along the wave train axes.
(the red lines in Figs. 5a, b, defined as a straight line approximately across all the centers of each disturbance in a daily map at 850 hPa corresponding to Figs. 3, 4). Figures 6a–d indicate that most of the disturbances in 2004 exhibited equivalent barotropic vertical structures, with a minimal tilt with height in their relative vorticity and wind components. The maximum amplitudes of these fields occurred from the middle to the lower troposphere. However, by 3 September, the wind fields of anticyclonic disturbance “L” and cyclonic disturbance “M” tilted considerably westward with height (Fig. 6e). In addition, the tilted disturbance “M” possessed a secondary maximum at 200 hPa in the vertical direction, which is different from the preceding disturbances.

The structure in Fig. 7a illustrates the vertical structure in the synoptic-scale waves in 2006. On 04 July 2006, the preceding disturbances “A” and “B” exhibited straight vertical structures with maximum amplitudes in the lower troposphere, which are similar
Nevertheless, the vertical structures of the subsequent disturbances were different. On 12 July, disturbances “D” and “E” demonstrated deep vertical structures of up to almost 100 hPa and a secondary maximum in the wind perturbations in the upper-level (Fig. 7b). Subsequently, the maximum amplitude of the following disturbance “F” appears in the upper troposphere. On 19 July, disturbances “G” and “H” demonstrated a tilting vertical structure, while the southwesterlies in the lower-level corresponded to the upper-level northeasterlies of the upper-level (Fig. 7c). Figures 2b and 4 revealed that the synoptic disturbances were relatively weak on 19 July. In fact, the disturbances in the lower troposphere were weak, but the perturbations were stronger in the upper troposphere than preceding disturbances. Figures 7d–f illustrate a series of disturbances with a relative shallow structure and weak horizontal winds, although these disturbances can be found in the horizontal map at 850 hPa (Fig. 4f).

To show the detailed evolutions of these two synoptic-scale wave trains and their vertical structures,
time-height cross-sections of the relative vorticities and tangential winds over the reference points are shown in Fig. 8. The reference point was first selected as the center point of the axis of the storm track (the red lines in Fig. 5). However, the daily maps show that the axis of the wave train shifts daily in both 2004 and 2006. Therefore, the reference points of 2004 and 2006 are set to vary in latitude following the axis but at a fixed longitude. In 2004, the fixed longitude is set to 145°E, and in 2006, the fixed longitude is set to 130°E. Consequently, the latitudes of the reference points changed between 10–23°N. Comparing these results to those with a reference point at a fixed longitude and latitude, the structures and variations of the wave train change very little, except in magnitude.

Figure 8a reveals a westward-propagating wave train with a nearly barotropic structure from 11 August to 2 September 2004. The maximum relative vorticity occurs in the middle troposphere, and the maximum of the cross wind usually occurs slightly below that of the relative vorticity. After 2 September, when the wave train began to dissipate, the disturbance demonstrated a more westward tilt with height. Meanwhile, the upper-level perturbation developed, and the lower-level perturbation quickly weakened.

Figure 8b shows the detailed vertical structure of the synoptic wave train in 2006. The most notable signal of the wave propagation is found before 23 July. The wave propagates in the lower troposphere from 6 to 12 July, in which the vertical structure is similar to those disturbances in 2004. After 13 July, the wave packet begins to propagate in the upper troposphere from until 19 July and in both the upper and lower levels from 19 July to 7 August. The lower-tropospheric disturbances are relatively weak from 13 to 19 July 2006 (Figs. 2b, 4), but the upper-level perturbations are prominent during the same period (Fig. 8b). After 31 July, the low-level signal, regard-
less of the relative vorticity or the cross wind, is weak. However, the wave activity is still identifiable from the fluctuations in the relative vorticity at the upper troposphere. Although Fig. 2b shows a low-level wave train after 13 July, Fig. 8b indicates that the primary propagation occurred in the upper troposphere during this period. The next section will discuss what factors caused the different vertical wave structures in these two events.

4. Possible mechanisms

4.1 The role of monsoon trough

The lower-tropospheric monsoon trough features remarkable zonal convergence and shear, inducing most tropical cyclogenesis over the western North Pacific (Ritchie and Holland 1999; Molinari and Vollaro 2013; Feng et al. 2014; Wu et al. 2015; Huangfu et al. 2017). Figure 9 shows the background horizontal flow at 850 hPa averaged for each active period of the synoptic-scale wave trains. In general, the monsoon troughs in these two cases are both stronger than the climatology (see Figs. 1g, h in Molinari and Vollaro 2013). The monsoon trough in 2006 was located slightly west of that in 2004. Furthermore, the weaker westerlies to the south of the trough line in 2006 indicate that the trough was relatively weaker than that in 2004.

Energetic diagnostics are usually adopted to quantitatively represent the influence of the monsoon trough on the synoptic-scale eddies. In this study, the EKE growth is calculated using barotropic energy conversions, which describes the kinetic energy transport from the mean flow to the transient eddies via barotropic instabilities. The calculation of barotropic energy conversions follows the form derived by Maloney and Hartmann (2001):

$$\frac{\partial K'_\text{baro}}{\partial t} = -u' \frac{\partial v'}{\partial y} - u' \frac{\partial v}{\partial x} - u \frac{\partial v'}{\partial x} - v \frac{\partial v}{\partial y}.$$ 

In Eq. (1), $K'$ represents the EKE, $(u', v')$ represent the zonal and meridional components of the 3-8-day filtered winds, and $(\bar{u}, \bar{v})$ represent the zonal and meridional components of the lower-frequency horizontal winds obtained using a 10-day low-pass filter. The terms on the right side of Eq. (1) represent the conversion of the kinetic energy from the mean kinetic energy to EKE through the meridional shear of the zonal basic flow, the zonal shear of the meridional basic flow, the zonal convergence of the zonal basic flow, and the meridional convergence of the meridional basic flow.

Figures 9a and 9b describe the horizontal distributions of the kinetic energy tendencies caused by barotropic energy conversion averaged from 28 July to 11 September 2004 and from 6 July to 7 August 2006, respectively. The EKE increased in most of the regions along with the monsoon trough, except for small regions to the south of Japan during 2004 (Fig. 9a). In contrast, in 2006, the EKE only grew in the relatively narrower region near the axis of the monsoon trough (Fig. 9b). A positive EKE tendency occurred more eastward during 2004 than during 2006. The barotropic conversions may be an important reason for the different horizontal distributions of the synoptic-scale wave activities between 2004 and 2006. As Table 2
The averaged EKE growth rate reached $8.56 \times 10^{-5}$ m$^2$ s$^{-3}$ in the lower troposphere in 2004, which is more than twice that in 2006. This is primarily caused by the conversion from the meridional shear of the zonal mean flow. The monsoon trough features a horizontal convergence between the monsoonal westerlies and trade easterlies. This study traces the zonal shift of the monsoon trough in the time-longitude diagrams by the intersection of westerlies and easterlies. Figures 10a and 10b show Hovmöller diagrams of 10-day low-pass filtered 850-hPa zonal winds averaged from 5°N to 20°N during the active periods of these two wave trains. Before late-July 2004, the weak westerly flow was generally located to the west of 125°E, indicating that the monsoon trough was located over the South China Sea (Fig. 10a). On 31 July, the westerlies abrupt intensified and extended eastward to near 155°E. During August, the eastern end of the westerlies was sustained near 150°E, providing favorable zonal convergence and shear for the maintenance of the synoptic-scale wave packet. Figure 10b shows the zonal variations of the zonal basic flow from 15 June to 15 August 2006. The eastern end of the wave train remained near 135°E with west-east oscillation during July, which may be associated with the intraseasonal oscillations, e.g., the Madden Julian oscillation (Supplement 3). The break-off of the low-level wave activity during 12–16 July is associated with the westward retreat of the monsoon trough. After 01 August, the westerlies shifted eastward to 145°E, which established a region of zonal convergence. However, the low-level disturbances did not develop (Fig. 8b), despite the remarkable zonal

Table 2. Contributions to the growth of the EKE at 850 hPa through barotropic energy conversions averaged over 10–20°N, 135–145°E ($10^{-5}$ m$^2$ s$^{-3}$).

| Year | $\frac{\partial K'}{\partial t}$ | $u'v'\frac{\partial u}{\partial y}$ | $u'v'\frac{\partial v}{\partial x}$ | $u'^2\frac{\partial u}{\partial y}$ | $v'^2\frac{\partial v}{\partial x}$ |
|------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|
| 2004 | 8.56                            | 10.60                           | -3.22                           | 5.72                            | -4.54                           |
| 2006 | 2.56                            | 4.72                            | -0.84                           | 3.43                            | -4.75                           |

Fig. 10. Time-longitude section of 10-day low-pass-filtered zonal wind (contour; m s$^{-1}$) and growth rate of EKE through barotropic energy conversion (shaded, unit: $10^{-5}$ m$^2$ s$^{-3}$) at 850 hPa averaged for 5–20°N (a) from 28 July to 11 September 2004; (b) from 6 July to 7 August 2006.
convergence during August 2006. This result suggests that the monsoon trough is not the only factor affecting the development of the synoptic-scale disturbances.

4.2 The role of TUTT

The upper-tropospheric circulation patterns averaged for the active periods of the wave trains in 2004 and 2006 are shown in Fig. 11. The subtropical anticyclonic flow in association with the subtropical high was located more westward in 2006 (Fig. 11b) than in 2004 (Fig. 11a). A much stronger TUTT was found in 2006 than in 2004, with a southwest-northeast orientation. The EKE growth through the barotropic energy conversion is also shown in Figs. 11a and 11b. In 2006, a large amount of kinetic energy was converted from the mean flow into eddies near the western end of the TUTT due to the westward intrusion of the TUTT (Fig. 11b). However, no remarkable barotropic energy conversion can be found in the upper troposphere in 2004.

Figure 12 illustrates the evolutions of the TUTT-related zonal winds during each active period of the synoptic-scale waves. In 2004, easterlies prevailed over the western Pacific Ocean through the summer until mid-September (Fig. 12a). Thus, the TUTT did not impact the western North Pacific through July and August. The region near 140°E, where the low-level disturbances developed, was dominated by negative EKE tendencies, suggesting that the upper-tropospheric disturbances were inhibited by the large-scale mean flow. However, in 2006, the TUTT-related westerlies reached 150°E, which indicated the westward penetration of the TUTT (Fig. 12b). Affected by the TUTT, a large amount of kinetic energy was converted from the mean flow into synoptic-scale eddies at approximately 150°E. This conversion occurred primarily through the meridional convergence of the meridional basic flow and the meridional shear of the zonal basic flow (Table 3). The westward penetration of TUTT favors the development of upper-level perturbations through barotropic energy conversion in 2006.

4.3 Vertical wind shear

The background vertical wind shear speed is defined as 

\[ S = \sqrt{(U_{200} - U_{850})^2 + (V_{200} - V_{850})^2}, \]

and the zonal vertical shear is defined as 

\[ S_z = U_{200} - U_{850}, \]

and the meridional vertical shear is defined as 

\[ S_m = V_{200} - V_{850}. \]

In both 2004 and 2006, a negative \( S_z \) (easterly shear) appeared in the southwestern ocean of the western North Pacific, while a positive \( S_z \) (westerly shear) appeared in the central Pacific (Figs. 13a, b). The regions with the negative or positive \( S_z \) correspond to the area of large value of the wind shear speed \( S \). Regions with weaker vertical wind shears \( S \) spread from northwest to southeast in both years. The distinction is that a stronger westerly shear (negative \( S_z \)) shifted further westward in 2006 due to the westward intrusion of the TUTT.

Figure 14 shows the variation of environmental total wind shear speed, zonal wind shear, and meridional wind shear affecting the synoptic-scale disturbances. During the period of the 2004 event, the zonal wind shear was negative, which meant that the easterly shear prevailed (Fig. 14a). As a result, the synoptic-scale disturbances were confined to the lower tropo-
Fig. 12. Time-longitude section of 10-day low-pass-filtered zonal wind (contour; m s$^{-1}$) and growth rate of EKE (shaded; 10$^{-5}$ m$^2$ s$^{-3}$) at 200 hPa averaged for 15–20°N (a) from 28 July to 11 September 2004; (b) from 6 July to 7 August 2006.

Table 3. Contributions to the growth of the EKE at 200 hPa through barotropic energy conversions averaged over 10–20°N, 135–145°E (10$^{-5}$ m$^2$ s$^{-3}$).

| Year | $\partial K'/\partial t$ | $u'v'/\partial y$ | $u'u'/\partial x$ | $u'\partial u'/\partial x$ | $v'^2/\partial y$ |
|------|------------------|-----------------|-----------------|--------------------------|-----------------|
| 2004 | $-5.00$          | $-0.72$         | $-0.28$         | $-1.26$                  | $-2.72$         |
| 2006 | $-0.85$          | $-1.20$         | $1.25$          | $-3.29$                  | $2.38$          |

sphere due to the easterly sheared environment (Fig. 8a). After 5 September, the zonal wind shear gradually increased from negative to positive (Fig. 14a), which was corresponding to the westward penetration of the TUTT (Fig. 12a). Meanwhile, the low-level disturbances gradually decayed, and wind perturbations appeared in the upper troposphere (Fig. 8a).

The vertical wind shear varied with remarkable fluctuations in 2006. Before 15 July, the negative zonal shear $S_z$ denotes that the easterly shear prevailed, which confined synoptic-scale waves in the lower troposphere in these disturbances (Fig. 14b). The vertical structure of these disturbances was quite similar to that in 2004 (Fig. 8). From 15 to 19 July, the sign of the zonal wind shear became positive, and a weak westerly shear prevailed during these days (Fig. 14b). Simultaneously, disturbances appeared mainly in the upper troposphere (Fig. 8b). After 19 July, the zonal wind shear fluctuated between $-2.0$ and $-12.0$ m s$^{-1}$. Without a stable vertical shear type of background circulation, the structure of the disturbances was less organized during this period (Fig. 8b).

The eddy baroclinic conversion $(-R/P \cdot \omega' t')$, which is related to the background vertical wind shear, was also calculated following previous work (Lau and Lau 1992; Au-Yeung and Tam 2018). Caused by a stronger zonal wind shear, the vertical-integrated eddy baroclinic conversion rate was greater in 2004 than that in 2006 (Supplement 4). Combining the effects of the background vertical wind shear and convection-induced upward motion of the warm air, the eddy baroclinic conversion is an important reason that induced
Fig. 13. Total vertical wind shear speed (contour; m s\(^{-1}\)) and vertical shear of zonal wind (shaded; m s\(^{-1}\)) during (a) 28 July to 11 September 2004; (b) 6 July to 7 August 2006.

Fig. 14. Total vertical wind shear speed (black curve; m s\(^{-1}\)), vertical shear of meridional wind (blue; m s\(^{-1}\)), and vertical shear of zonal wind (red; m s\(^{-1}\)) between 200 hPa and 850 hPa during (a) 28 July to 11 September 2004; (b) 6 July to 7 August 2006. Calculations are averaged in a 5° × 5° rectangle around the reference points in Fig. 8.
a stronger wave packet in 2004. More examinations should be performed to investigate the potential effect of baroclinic conversion in the vertical structure of the synoptic-scale disturbances.

5. Conclusions and discussion

5.1 Conclusions

Using the CFSR reanalysis data, JTWC best tracks, and TRMM 3B42 precipitation data, the current work performed a comparative study investigating the large-scale atmospheric factors controlling the structures and evolutions of the synoptic-scale disturbances over the western North Pacific during the boreal summer. Two long-lasting synoptic-scale wave trains from 28 July to 11 September 2004 and from 6 July to 7 August 2006 were selected in this study. Although some differences are found in the lower troposphere, the essential difference between these two cases is in the upper-level. In 2004, the maximum wind perturbations occurred in the middle-lower troposphere with an equivalent barotropic structure in 2004. However, although wind perturbations can be found in the lower troposphere in 2006, the maximum perturbations appear in the upper troposphere.

The reason for the distinct vertical structures in these two events was attributed to the different displacements of the monsoon trough, the TUTT, and the vertical wind shear. In 2004, the stronger monsoon trough with the intensified low-level cyclonic flow extended further east and persisted for more than one month. The TUTT retreated to the central North Pacific, and the upper-level circulation was dominated by a southwestward flow. Modulated by the monsoon trough and the TUTT, an easterly shear environment prevailed over the western North Pacific, which favored wave development in the lower troposphere. The lower-tropospheric disturbances further developed under the effects of the strong monsoon trough. In contrast, in 2006, the monsoon trough also existed with a weaker strength and an intraseasonal variation. Moreover, the TUTT was intensified and penetrated further westward. Affected by this configuration of the upper-level TUTT and the lower-level monsoon trough, the regions with enhanced synoptic-scale disturbance activity were dominated by a westerly shear or a very weak shear environment. While the low-level disturbance development was inhibited by westerly shear, the upper-level TUTT supported the kinetic energy for the growth of upper-tropospheric perturbations, especially during 13–19 July. Consequently, the lower-tropospheric wind anomalies in those disturbances of 2006 were weaker, but the upper-level perturbations were stronger than those in 2004.

5.2 Discussion

The above results suggest that the development of a synoptic-scale disturbance is controlled by the co-effect of the vertical wind shear and the large-scale horizontal ambient flow. While environmental easterly shear confines waves in the lower troposphere and suppresses upper-level perturbations, an intensified lower-tropospheric monsoon trough favors low-level development through barotropic energy conversion. While environmental westerly shear confines waves in the upper troposphere, upper-level disturbances develop due to the favorable large-scale environment when the TUTT penetrates westward. The co-effect of the monsoon trough, the TUTT, and the vertical wind shear can intensify a synoptic-scale wave packet in either the lower or the upper troposphere. However, the origin of the synoptic-scale perturbation may not be related to the monsoon trough or the TUTT. For example, the westward-propagating upper-level disturbances can be transported downward while penetrating an easterly shear environment from the east (Tam and Li 2006; Zhou and Wang 2007; Feng et al. 2016). While these enhanced disturbances propagate into the western North Pacific, low-level perturbations may be induced by downward energy transport to serve as a low-level initial perturbation. A preliminary examination suggested that the low-level disturbances may have originated from the upper-level perturbations via downward wave propagation in 2006. An in-depth diagnosis of the vertical wave propagation may aid in answering this question.

As the most important precursor for TC genesis, a synoptic-scale disturbance is usually identified from lower-tropospheric wind perturbation. However, the present study shows that those disturbances with an upper-level maximum, in which the structures are similar to those in 2006, may be misidentified as weak conventional TD-type disturbances with maximum amplitudes occurring in the lower troposphere. The disturbances with a wind anomaly maximum in the upper troposphere exhibit a lower TC genesis efficiency (Ge et al. 2013). The misidentification of synoptic-scale disturbance may be one of the reasons why not all long-lasting events are associated with multiple TC events.

The current results emphasized the critical role of the TUTT in regulating the vertical structure of the synoptic-scale disturbances over the western North Pacific. However, the interactions between the TCs and synoptic-scale waves, including the TC forma-
tions in the wave packet and the energy dispersions from TCs, are not discussed in this study. A mature TC can strengthen synoptic disturbances in the wake (Li et al. 2006; Ge et al. 2010). The best track data show that the TCs in 2004 were stronger compared to those in 2006, which may be another reason why the disturbances in 2004 were stronger than those in 2006.

Another remaining issue is the question of how these two wave packets persist for more than one month. The role of the thermal state of the equatorial ocean, e.g., the El Niño–Southern Oscillation (ENSO) event, exerts a significant influence on the interannual variabilities of the monsoon trough and tropical waves over the western North Pacific (e.g., Wu et al. 2014). The year 2004 was a developing year for the central Pacific El Niño. In contrast, despite some arguments regarding the type of El Niño in 2006 (Yu and Kim 2013), the year 2006 was suggested to be a developing year for a weak conventional eastern Pacific El Niño (e.g., Chen and Tam 2012). The ENSO does affect the synoptic-scale activity during boreal summer by regulation of the large-scale circulation, e.g., the monsoon trough and the TUTT. The role of the interannual or interdecadal background should be discussed in revealing the large-scale control of the long-lasting synoptic-scale events, as indicated by Hu et al. (2018).

**Supplements**

Supplement 1 shows E-vectors at 850 hPa averaged (a) from 28 July to 11 September 2004; (b) from 6 July to 7 August 2006; (c) and (d) are identical to (a) and (b) except for at 200 hPa. Red lines denote the averaged wave axis as in Fig. 5.

Supplement 2 shows vertical cross-section of $v^t$ (contour, $K$ m s$^{-3}$) averaged for 10–20°E and (a) from 28 July to 11 September 2004; (b) from 6 July to 7 August 2006.

Supplement 3 shows the real-time multivariate MJO index on a MJO phase diagram during (a) from 1 July to 30 September 2004, and (b) from 1 June to 31 August 2006.

Supplement 4 shows the vertical cross-section of eddy baroclinic conversion rate (contour, $10^{-3}$ m$^2$ s$^{-3}$) along the thick straight line in Figs. 5a and 5b averaged (c) from 28 July to 11 September 2004; (d) from 6 July to 7 August 2006.

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**References**

Aiyer, A. R., and J. Molinari, 2003: Evolution of mixed Rossby–gravity waves in idealized MJO environments. J. Atmos. Sci., 60, 2837–2855.

Au-Yeung, A. Y. M., and C.-Y. Tam, 2018: Dispersion characteristics and circulation associated with boreal summer westward-traveling mixed Rossby–gravity wave–like disturbances. J. Atmos. Sci., 75, 513–533.

Briegel, L. M., and W. M. Frank, 1997: Large-scale influences on tropical cyclogenesis in the western North Pacific. Mon. Wea. Rev., 125, 1397–1413.

Chang, C. P., V. F. Morris, and J. M. Wallace, 1970: A statistical study of easterly waves in the western Pacific: July–December 1964. J. Atmos. Sci., 27, 195–201.

Chen, G., and R. Huang, 2009: Interannual variations in mixed Rossby–gravity waves and their impacts on tropical cyclogenesis over the western North Pacific. J. Climate, 22, 535–549.

Chen, G., and C.-Y. Tam, 2012: A new perspective on the excitation of low-tropospheric mixed Rossby–gravity waves in association with energy dispersion. J. Atmos. Sci., 69, 1397–1403.

Chen, T.-C., S.-Y. Wang, M.-C. Yen, and A. J. Clark, 2008: Are tropical cyclones less effectively formed by easterly waves in the western North Pacific than in the North Atlantic? Mon. Wea. Rev., 136, 4527–4540.

Dickinson, M., and J. Molinari, 2002: Mixed Rossby–gravity waves and western Pacific tropical cyclogenesis. Part I: Synoptic evolution. J. Atmos. Sci., 59, 2183–2196.

Done, J., G. Holland, and P. Webster, 2011: The role of wave energy accumulation in tropical cyclogenesis over the tropical North Atlantic. Climate Dyn., 36, 753–767.

Duchon, C. E., 1979: Lanczos filtering in one and two dimensions. J. Appl. Meteor., 18, 1016–1022.

Feng, T., G. H. Chen, R. H. Huang, and X. Y. Shen, 2014: Large-scale circulation patterns favourable to tropical cyclogenesis over the western North Pacific and associated barotropic energy conversions. Int. J. Climatol., 34, 216–227.

Feng, T., X.-Q. Yang, W. Zhou, R. Huang, L. Wu, and D. Yang, 2016: Synoptic-scale waves in sheared background flow over the western North Pacific. J. Atmos. Sci., 73, 4583–4603.

Fu, B., T. Li, M. S. Peng, and F. Weng, 2007: Analysis of tropical cyclogenesis in the western North Pacific for 2000 and 2001. Wea. Forecasting, 22, 763–780.

Ge, X., T. Li, and X. Zhou, 2007: Tropical cyclone energy dispersion under vertical shears. Geophys. Res. Lett., 34, L23807, doi:10.1029/2007gl031867.

Ge, X., T. Li, Y. Wang, and M. S. Peng, 2008: Tropical
cyclone energy dispersion in a three-dimensional primitive equation model: Upper-tropospheric influence. J. Atmos. Sci., 65, 2272–2289.

Ge, X., T. Li, and M.S. Peng, 2010: Cyclogenesis simulation of Typhoon Papiroon (2000) associated with Rossby wave energy dispersion. Mon. Wea. Rev., 138, 42–54.

Ge, X., T. Li, and M. S. Peng, 2013: Tropical cyclone genesis efficiency: Mid-level versus bottom vortex. J. Trop. Meteor., 19, 197–213.

Gray, W. M., 1968: Global view of the origin of tropical disturbances and storms. Mon. Wea. Rev., 96, 669–700.

Han, Y., and B. Khouider, 2010: Convectively coupled waves and storms. J. Atmos. Sci., 67, 2913–2942.

Holland, G. J., 1995: Scale interaction in the western Pacific monsoon. Meteor. Atmos. Phys., 56, 57–79.

Holton, J. R., 1971: A diagnostic model for equatorial wave disturbances: The role of vertical shear of the mean zonal wind. J. Atmos. Sci., 28, 55–64.

Hu, K., J. C. L. Chan, G. Huang, G. Chen, and W. Mei, 2018: A train-like extreme multiple tropical cyclone genesis event in the northwest Pacific in 2004. Geophys. Res. Lett., 45, 8529–8535.

Huangfu, J., R. Huang, W. Chen, T. Feng, and L. Wu, 2017: Interdecadal variation of tropical cyclone genesis and its relationship to the monsoon trough over the western North Pacific. Int. J. Climatol., 37, 3587–3596.

Huffman, G. J., R. F. Adler, D. T. Bolvin, G. Gu, E. J. Nelkin, K. P. Bowman, Y. Hong, E. F. Stocker, and D. B. Wolff, 2007: The TRMM multisatellite precipitation analysis (TMPA): Quasi-global, multiyear, combined-sensor precipitation estimates at fine scales. J. Hydroeteor., 8, 38–55.

Krouse, K. D., A. H. Sobel, and L. M. Polvani, 2008: On the wavelength of the Rossby waves radiated by tropical cyclones. J. Atmos. Sci., 65, 644–654.

Kurihara, Y., M. A. Bender, R. E. Tuleya, and R. J. Ross, 1995: Improvements in the GFDL hurricane prediction system. Mon. Wea. Rev., 123, 2791–2801.

Lau, K.-H., and N.-C. Lau, 1990: Observed structure and propagation characteristics of tropical summertime synoptic scale disturbances. Mon. Wea. Rev., 118, 1888–1913.

Lau, K.-H., and N.-C. Lau, 1992: The energetics and propagation dynamics of tropical summertime synoptic-scale disturbances. Mon. Wea. Rev., 120, 2523–2539.

Li, T., and B. Fu, 2006: Tropical cyclogenesis associated with Rossby wave energy dispersion of a preexisting typhoon. Part I: Satellite data analyses. J. Atmos. Sci., 63, 1377–1389.

Li, T., X. Ge, B. Wang, and Y. Zhu, 2006: Tropical cyclogenesis associated with Rossby wave energy dispersion of a preexisting typhoon. Part II: Numerical simulations. J. Atmos. Sci., 63, 1390–1409.

Malone, E. D., and D. L. Hartmann, 2001: The Madden–Julian oscillation, barotropic dynamics, and North Pacific tropical cyclone formation. Part I: Observations. J. Atmos. Sci., 58, 2545–2558.

Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. J. Meteor. Soc. Japan, 44, 25–43.

Molinari, J., and D. Vollaro, 2013: What percentage of western North Pacific tropical cyclones form within the monsoon trough? Mon. Wea. Rev., 141, 499–505.

Molinari, J., D. Vollaro, and K. L. Corbosiero, 2004: Tropical cyclone formation in a sheared environment: A case study. J. Atmos. Sci., 61, 2493–2509.

Murakami, H., 2014: Tropical cyclones in reanalysis data sets. Geophys. Res. Lett., 41, 2133–2141.

Reed, R. J., and E. E. Recker, 1971: Structure and properties of synoptic-scale wave disturbances in the equatorial western Pacific. J. Atmos. Sci., 28, 1117–1133.

Ritchie, E. A., and G. J. Holland, 1999: Large-scale patterns associated with tropical cyclogenesis in the western Pacific. Mon. Wea. Rev., 127, 2027–2043.

Sadler, J. C., 1976: A role of the tropical upper tropospheric trough in early season typhoon development. Mon. Wea. Rev., 104, 1266–1278.

Sadler, J. C., 1978: Mid-season typhoon development and intensity changes and the tropical upper tropospheric trough. Mon. Wea. Rev., 106, 1137–1152.

Saha, S., S. Moorthi, H. Pan, X. Wu, J. Wang, S. Nadiga, P. Tripp, R. Kistler, J. Woollen, D. Behringer, H. Liu, D. Stokes, R. Grumbine, G. Gayno, J. Wang, Y. Hou, H. Chuang, H. H. Juang, J. Sela, M. Iredell, R. Treadon, D. Kleist, P. Van Delst, D. Keyser, J. Derber, M.Ek, J. Meng, H. Wei, R. Yang, S. Lord, H. van den Dool, A. Kumar, W. Wang, C. Long, M. Chelliah, Y. Xue, B. Huang, J. Schemm, W. Ebisuzaki, R. Lin, P. Xie, M. Chen, S. Zhou, W. Higgins, C. Zou, Q. Liu, Y. Chen, Y. Han, L. Cucurull, R. W. Reynolds, G. Rutledge, and M. Goldberg, 2010: The NCEP climate forecast system reanalysis. Bull. Amer. Meteor. Soc., 91, 1015–1057.

Saha, S., S. Moorthi, X. Wu, J. Wang, S. Nadiga, P. Tripp, D. Behringer, Y. T. Hou, H. Y. Chuang, M. Iredell, M. Ek, J. Meng, R. Yang, M. P. Mensez, H. van den Dool, Q. Zhang, W. Wang, M. Chen, and E. Becker, 2014: The NCEP climate forecast system version 2. J. Climate, 27, 2185–2208.

Schenkel, B. A., 2016: A climatology of multiple tropical cyclone events. J. Climate, 29, 4861–4883.

Serra, Y. L., G. N. Kiladis, and M. F. Cronin, 2008: Horizontal and vertical structure of easterly waves in the Pacific ITCZ. J. Atmos. Sci., 65, 1266–1284.

Sobel, A. H., and C. S. Bretherton, 1999: Development of synoptic-scale disturbances over the summertime tropical northwest Pacific. J. Atmos. Sci., 56, 3106–3127.

Takayabu, Y. N., and T. Nitta, 1993: 3–5 day-period disturbances coupled with convection over the tropical Pacific Ocean. J. Meteor. Soc. Japan, 71, 221–246.

Tam, C.-Y., and T. Li, 2006: The origin and dispersion characteristics of the observed tropical summertime synoptic-scale waves over the western Pacific. Mon.
Wallace, J. M., 1971: Spectral studies of tropospheric wave disturbances in the tropical western Pacific. *Rev. Geo-
phys.*, 9, 557–612.
Wallace, J. M., and C. P. Chang, 1969: Spectrum analysis of large-scale wave disturbances in the tropical lower
troposphere. *J. Atmos. Sci.*, 26, 1010–1025.
Wang, B., and X. Xie, 1996: Low-frequency equatorial
waves in vertically sheared zonal flow. Part I: Stable waves. *J. Atmos. Sci.*, 53, 449–467.
Wang, C., and L. Wu, 2016: Interannual shift of the tropical
upper-tropospheric trough and its influence on tropical
cyclone formation over the western North Pacific. *J.
Climate*, 29, 4203–4211.
Webster, P. J., and H.-R. Chang, 1988: Equatorial energy
accumulation and emanation regions: Impacts of a
zonally varying basic state. *J. Atmos. Sci.*, 45, 803–
829.
Wu, L., Z. Wen, R. Huang, and R. Wu, 2012: Possible
linkage between the monsoon trough variability and the tropical cyclone activity over the western North
Pacific. *Mon. Wea. Rev.*, 140, 140–150.
Wu, L., Z. Wen, T. Li, and R. Huang, 2014: ENSO-phase
dependent TD and MRG wave activity in the western
North Pacific. *Climate Dyn.*, 42, 1217–1227.
Wu, L., Z. Wen, and R. Wu, 2015: Influence of the monsoon
trough on westward-propagating tropical waves over the western North Pacific. Part I: Observations. *J.
Climate*, 28, 7108–7127.
Xie, X., and B. Wang, 1996: Low-frequency equatorial
waves in vertically sheared zonal flow. Part II: Unsta-
bile waves. *J. Atmos. Sci.*, 53, 3589–3605.
Xu, Y., T. Li, and M. Peng, 2013: Tropical cyclogenesis in
the western North Pacific as revealed by the 2008–09
YOTC data. *Wea. Forecasting*, 28, 1038–1056.
Yu, J.-Y., and S. T. Kim, 2013: Identifying the types of
major El Niño events since 1870. *Int. J. Climatol.*, 33,
2105–2112.
Zhou, X., and B. Wang, 2007: Transition from an eastern
Pacific upper-level mixed Rossby-gravity wave to a
western Pacific tropical cyclone. *Geophys. Res. Lett.*, 34, L24801, doi:10.1029/2007GL031831.