Sensitivity of Western North Pacific Summertime Tropical Synoptic-Scale Disturbances to Extratropical Forcing – A Regional Climate Model Study

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Abstract

The role of extratropical forcing on the summertime tropical synoptic-scale disturbances (TSDs) in the western North Pacific has been investigated by conducting parallel integrations of the Regional Climate Model. The suite of experiments consists of a control run (CTRL) with European Centre for Medium-Range Forecasts Reanalysis data as boundary conditions, and an experimental run (EXPT) with the same setting, except that signals with zonal wavenumber > 6 were suppressed at the northern boundary (located at 42°N) of the model domain. Comparison between CTRL and EXPT showed that without extratropical forcing, there is weaker TSD activity in the June-to-August season, with reduced precipitation over the TSD pathway. Associated with suppressed TSD, southeastward-directed wave activity is also reduced, leading to less active mixed Rossby gravity waves in the equatorial western Pacific area. Further analysis revealed that extratropical forcing and associated circulation changes can modulate the TSD wavetrain and its coherence structure, in relation to low-level vorticity in the far western North Pacific. In CTRL, west of about 140°E, TSD-related circulation tends to be stronger; in EXPT, vorticity signals and vertical motions are found to be slightly more coherent in the more eastern portion of the TSD wavetrain. The latter enhanced coherency of TSD east of 140°E, from the EXPT simulations, might be due to changes in wave activity transport channeled by modulated upper-level mid-latitude westerlies in EXPT. Our results serve to quantify how extratropical forcing and related general circulation features influence western North Pacific summertime TSD activities. Implications on the understanding of the initiation of TSD are discussed.

Keywords tropical depression-type disturbance; tropical-extratropical interaction; regional climate model; limited-area model


1. Introduction

Summertime tropical synoptic-scale disturbances (TSDs) and their impacts on the tropical weather have interested meteorologists and atmospheric scientists for a long time (Yanai and Maruyama 1966; Nitta 1970; Wallace 1971; Nitta and Takayabu 1985; Dickinson and Molinari 2002). With periods of ~ 3–8 days and wavelengths of ~ 3000 km, TSDs are active over various tropical oceanic areas, including the off-equatorial western North Pacific (Lau and Lau 1990). TSDs in this region originate at approximately 160°E along the equator and propagate westward, reaching the Asian Continent (Nitta and Takayabu 1985; Dickinson and Molinari 2002). These systems are associated with low-level wave troughs (i.e., vorticity maxima),
which in turn are collocated with warm cores in the mid-troposphere at approximately 500 hPa. Cold anomaly is found above and below the warm perturbation, implying a sandwiched vertical temperature structure of TSD (Wallace and Chang 1969; Reed and Recker 1971; Lau and Lau 1990; Tam and Li 2006; Serra et al. 2008). Mechanisms for TSD development include local baroclinic and barotropic instability (Chang et al. 1970; Burpee 1972; Lau and Lau 1992; Maloney and Hartmann 2001; Wu et al. 2014; Feng et al. 2014), whereas these disturbances are also closely related to other equatorial waves, for instance, mixed Rossby gravity (MRG) waves (Matsuno 1966; Takayabu and Nitta 1993; Liebmann et al. 1990; Dunkerton and Baldwin 1995; Chen and Tam 2012). Apart from temperature anomalies, wave troughs are also associated with strong convection and low-level convergence (Reed and Recker 1971; Lau and Lau 1990; Serra et al. 2008).

It is also well recognized that summertime TSDs are related to the genesis and development of tropical cyclones (TCs) (Maloney and Hartmann 2001; Schreck et al. 2012; Feng et al. 2014). Under favorable conditions, TCs can be formed due to TSD-related intense diabatic heating (Frank and Roundy 2006) and strong low-level convergence (Heta 1991). TSD is responsible for approximately 10–20% of TC genesis cases in the western North Pacific, and the number even reaches ~60% in the north Atlantic (Chen et al. 2008). The difference in the proportion of TSD-induced TC formation between two ocean basins is due to the presence of the monsoon gyre, to which approximately 70% of TC formation in the western North Pacific region is related (Chen et al. 2008). Apart from TCs, TSD can interact with/be influenced by circulation patterns on various timescales, such as the Madden–Julian Oscillation (Madden and Julian 1971; Maloney and Hartmann 2001; Cho et al. 2004) and El Niño–Southern Oscillation (ENSO) (Wu et al. 2014). TSD itself can also contribute to approximately 10% of the variability of western tropical Pacific summertime precipitation (Lubis and Jacobi 2015). A better understanding of factors affecting TSD is hence essential for improving tropical weather forecasts, seasonal climate predictions, and climate projections over the tropical to subtropical area.

Based on reanalysis data, Tam and Li (2006) found that TSD can be triggered by wave activity from extratropical regions to the tropics. Particularly, intrusion of upper-level potential vorticity from the mid-latitudes can lead to the formation of vorticity perturbations over the tropics, which later develop into TSD through wave energy dispersion. Later climatological analyses (Fukutomi and Yasunari 2014; Archambault et al. 2015) and single-case studies (Molinari and Vollaro 2012) also point out that the extratropical westerly flow can play a role in the formation and the strengthening of tropical easterly waves. Therefore, extratropical forcing can be an important source of TSD variability. In this study, we revisit the problem by assessing the contribution of extratropical wave activity to TSD over the western North Pacific, on the basis of regional atmospheric model experiments. By comparing results from parallel integrations with and without mid-latitude synoptic-scale waves in the model environment, the influence of extratropical forcing on TSD, including its associated tropical circulation impacts, can be inferred. Note that hitherto, very few studies have explored such tropical–extratropical linkage using numerical models. The rest of this paper is organized as follows. Section 2 outlines the model setup and experimental design. Section 3 gives the impacts of extratropical wave activity on the variability and structure of TSD, as well as the summertime tropical circulation over the western Pacific. Section 4 presents the discussion and conclusion.

2. Model experiments and data used

To test the sensitivity of TSD to extratropical forcing, numerical experiments were conducted using the Regional Climate Model system 3 (RegCM3), which is a compressible, grid-point model (Giorgi et al. 1993a, b; Pal et al. 2007). The model was maintained by the Earth System Physics section of the International Centre for Theoretical Physics. It was run at the horizontal resolution of 60 × 60 km, with 18 sigma levels, within the domain of 13°S–42°N and 80°E–160°W (see Fig. 1). Model parameterization schemes used included the modified Emanuel Scheme for cumulus convection (Emanuel 1991; Chow et al. 2006), Pal scheme (Pal et al. 2000) for large-scale precipitation, Biosphere-atmosphere transfer scheme (Dickinson et al. 1993) for land surface processes, Holtslag scheme (Holtslag et al. 1990) for planetary-boundary-layer processes, and NCAR CCM3 radiative scheme (Kiehl et al. 1998).

Under such a setting, model integrations were conducted for the April-to-September period. Six-hourly data from the European Centre for Medium-Range Forecasts Reanalysis (ERA-40; Uppala et al. 2005), as well as the climatological monthly mean National Oceanic and Atmospheric Administration Optimum Interpolation Sea Surface Temperature (Reynolds et al. 2002), were used as initial and lateral boundary
conditions. Reanalysis data from seven selected years, namely, 1983, 1984, 1986, 1990, 1992, 1993, and 2001 were chosen. Note that these years were not major El Niño nor La Niña events; they were chosen to avoid seasons during which there were prominent east–west shifts of synoptic-scale activities due to major ENSO events (Sobel and Maloney 2000). Based on these seven years, six-hourly climatology of dynamic and thermodynamic variables (i.e., wind, temperature, and humidity), at all vertical levels, was first obtained and then filtered to retain signals with zonal horizontal scales equivalent or larger than those for wavenumber 6 (i.e., zonal wavenumber |k| within 0–6). The spatially smoothed data were used as lateral boundary conditions for model integrations. At the northern boundary, transients (i.e., deviations from the seasonal mean) with |k| > 6, archived from each prescribed year, were superimposed onto the derived climatology, and the model was integrated for each of these years. In the control experiment (hereinafter referred to as CTRL), transients were set to their original amplitudes; in the other experiment (referred to as EXPT), transients with only 10% of their original amplitudes were then superimposed onto the climatology (see Table 1). For the eastern, western and southern boundaries in both experiments, no transient signals were imposed. Consequently, the only difference between the two experiments is the magnitude of |k| > 6 perturbations at the northern boundary. For each April-to-September period, five ensemble integrations were conducted, with the n-th-ensemble member initiated by treating the n-th day in April as the initial condition for the 1st of April. By comparing the difference between EXPT and CTRL, we can evaluate how extratropical synoptic-scale forcing from the northern boundary of the domain affects the regional climate, focusing on the behavior of TSD. In this study, only data from June to August (JJA) were used for all analyses.

To depict wave activities and forcing with extratropical origins, eddy heat transport at 850 hPa and also 200 hPa E-vectors from CTRL and EXPT were computed. Following Trenberth (1986), the E-vectors are defined as \[ \vec{E} = \left(\frac{\overline{\theta u'}}{\overline{\theta u}}, -\frac{\overline{\theta u'}}{\overline{\theta u}}\right) \], with prime denoting 3–8 day bandpass filtered values, and overbar seasonal and ensemble averaging. Results for heat transport \[ \overline{T' u'} \], with an asterisk denoting deviations from the seasonal mean, and E-vectors from the two experiments are given in Fig. 2. It can be clearly seen that there are strong southeastward-directed 200 hPa E-vectors at the northern boundary in CTRL, but not

![Fig. 1. Model domain for RegCM3 simulations [13°S–42°N, 80°E–160°W] and topography therein (shading; units: m).](image)

| Settings of the atmospheric model experiments. The only difference between CTRL and EXPT experiments lies in the magnitudes of wavenumber > 6 transients imposed at the northern boundary. See text for details. |
|-----------------|-----------------|-----------------|
| **CTRL** | **EXPT** |
| eastern, western and southern boundaries | wavenumber = 0–6 climatology | wavenumber = 0–6 climatology |
| northern boundary | wavenumber = 0–6 climatology + wavenumber > 6 transients | wavenumber = 0–6 climatology + 10% of wavenumber > 6 transients |
| sea surface temperature | climatological SST | climatological SST |
| number of simulations | 7 years × 5-member ensemble | 7 years × 5-member ensemble |
in EXPT (see Fig. 2a). This indicates southward wave energy dispersion from the north in CTRL only, consistent with the experimental setups. The 850 hPa northward heat transport is also stronger in CTRL near the northern boundary, especially over the continental area (see Fig. 2b). Expectedly, a stronger presence of mid-latitude synoptic-scale waves in CTRL serves to transport more heat to the north in the model environment. Overall, the above confirms that upper-level extratropical forcing from the northern boundary, in relation to mid-latitude synoptic-scale disturbances, is strongly suppressed in EXPT in comparison with CTRL because of our experimental design.

3. Results

To understand how synoptic-scale forcing might influence the mean circulation in the western North Pacific region, the background 850 hPa wind, precipitation from the CTRL and EXPT, as well as their difference, are given in Figs. 3a, 3c, and 3e. For CTRL, there is a stronger easterly wind branch over [20–30°N, 120–180°E], and a stronger northerly branch over the South China Sea. There is more heat transported to the north in CTRL at approximately 30°N at 850 hPa (see Fig. 2b). The 850 hPa mean temperature on the continent and that over the northern part of the domain is also lower in CTRL (Fig. 2a), which makes the meridional and land–sea temperature (pressure) gradient more negative (positive) in the low levels. This results in low-level wind field changes due to the differences in pressure gradient between CTRL and EXPT. Enhanced precipitation by approximately 2–4 mm day$^{-1}$ is found within [0–20°N, 100–160°E] in CTRL. Noteworthily, enhanced precipitation is collocated with typical pathways of TSD. To identify signals of TSD, bandpass filtered 850 hPa vorticity was computed (Lau and Lau 1990). Figures 3b, 3d, and 3f show the variance of the 3–8 day filtered low-level vorticity. For CTRL, it can be seen that the vorticity variance is generally larger over [5–25°N, 100–150°E], indicating stronger synoptic-scale wave activities. This supports the notion that extratropical forcing plays a role in modulating western North Pacific summertime synoptic-scale disturbances. Note that the seasonal mean precipitation is also enhanced in the same area with enhanced wave activities in CTRL, consistent with the fact that these systems are important in bringing rainfall in the region (see Fig. 3e).

E-vectors are also used to examine the 850 hPa 3–8 day westward propagating ($-20 \leq \vec{k} \leq -4$) wave activity and its related Rossby wave energy dispersion (see Fig. 4). Both CTRL and EXPT give southward and southeastward pointing E-vectors over the region of [5–25°N, 110–150°E], where a large 850 hPa vorticity variance is found (see Fig. 3). In both experiments, E-vectors are directed to the south to southeast in this area, consistent with the NE-SW tilted structure of the synoptic-scale eddies (Lau and Lau 1990). Note that the magnitude of E-vectors is approximately 30 % stronger in CTRL, especially for the zonal (eastward) component. This clearly indicates that there are stronger wave-like disturbances in the low levels in CTRL compared with EXPT. Indeed, the difference in E-vectors between the two experiments (see Fig. 4c) suggests that in CTRL, there is stronger southward Rossby wave energy dispersion toward more tropical latitudes. Such a difference can have an impact on
equatorial wave activities over the western Pacific sector, as will be subsequently shown. Conversely, it is noteworthy that synoptic-scale disturbances are not entirely suppressed in EXPT; in terms of their variance, the suppression is only approximately 20–30%, although extratropical waves have only 10% of their original magnitude (hence ~1% of variance). This suggests that summertime TSDs in the western North Pacific can exist, regardless of the presence of extratropical forcing from the north.

Chen and Tam (2012) reported a mechanism by which TSD can excite MRG waves. To depict activities of equatorial waves, the wavenumber–frequency spectra (Wheeler and Kiladis 1999; Au-Yeung and Tam 2018) of the symmetric component of the 850 hPa meridional wind, averaged over 13°S–13°N, were plotted in Fig. 5. Westward propagating MRG signals in CTRL were found to be significantly stronger than those in EXPT by more than 25% (about 0.1 in the logarithm plot), implying stronger MRG waves over the equatorial western Pacific in CTRL. This is also in accordance with Dunkerton and Baldwin (1995), who reported the transformation of MRG waves to TSD on the basis of observations. Overall, more active TSD and associated MRG waves in the tropical western Pacific are seen in CTRL compared to EXPT, and this is solely due to stronger forcing from extratropical locations in the experimental design.

Here, we further extract the wave-like signals associated with the synoptic-scale disturbances, using the results from empirical orthogonal function (EOF) analyses based on the bandpass filtered westward propagating 850 hPa vorticity signals over [0–30°N, 100–160°E] (see Section 2). Figure 6 shows the
standardized spatial pattern of the first and the second leading EOFs, based on data from all ensemble members from both CTRL and EXPT. It can be seen that the structures of all EOFs are wave-like, consistent with the waveform of TSD reported in previous studies (e.g., Lau and Lau 1990; Reed and Recker 1971). The fraction of domain-integrated variance explained by two leading EOFs is similar (approximately 10%), and two patterns are in quadrature in space. These two leading EOFs, together with their principal component
(PC) time series, thus represent westward propagating wave signals in the low-level vorticity. Wave trains in both CTRL and EXPT have similar structures and wavelengths. They both have the strongest amplitudes within 110–140°E and a NE-SW tilted structure west of 150°E. The wave starts to exhibit the NE-SW tilted structure in a more western location in CTRL than that in EXPT. The less tilted structure of eddies in CTRL over 120–140°E may increase the zonal propagation of wave activity and further affect wave energetics (Lau and Lau 1992; Trenberth 1986).

To analyze the difference in the strength of waves between CTRL and EXPT, time series of the combined magnitude of the two leading PCs are computed. The combined magnitude at time \( t \) (= day from June 1) is calculated by first computing \( y(t) = \sqrt{PC_1(t)^2 + PC_2(t)^2} \), where \( PC_1 \) and \( PC_2 \) are the magnitude of the two leading EOFs, for each season and each ensemble member. Finally, for each calendar day, the combined magnitude is found by averaging \( y(t) \) over all ensemble members and seasons. Results for the two experiments are shown in Fig. 7. It can be seen that during most of the days within the JJA season, CTRL gives stronger disturbances in this region than EXPT, with a difference reaching approximately 20%. This is also consistent with the previous result that MRG activities are approximately 25% stronger in CTRL (see Fig. 5). This indicates that extratropical forcing contributes to the formation and hence part of the variability of TSD, although the forcing is not a necessary condition for the waves to exist.

To obtain TSD-related circulation patterns, EOF reconstruction and regression techniques are used. Particularly, the two abovementioned leading EOFs for the 850 hPa vorticity were first used to reconstruct the vorticity time series at the reference point of (15°N, 120°E), which is a location with strong signals of TSD (see Figs. 3, 6). To obtain TSD-related circulation maps for a particular field, the following linear regression method is used:

\[
\phi_{reg}(x, y) = [\phi(t, x, y) - \bar{\phi}(x, y)] \times \zeta(t, 120°E, 15°N) / \sigma_{\zeta} 
\approx r(\phi, \zeta) \times \sigma_{\phi}(x, y),
\]

where \( \phi \) is the field of interest, \( \zeta \) is the EOF-filtered 850 hPa vorticity, \( \sigma \) is its temporal standard deviation, \( r \) is the correlation, bar denotes the time average, \( x \) and \( y \) denote the longitude and latitude of the datapoint. Figures 8a and 8b give the regression maps for precipitation and 850 hPa vorticity onto the (15°N, 120°E)
values of EOF-filtered vorticity. The structure of the regressed waves is consistent with spatial patterns of the first two leading EOFs shown in Fig. 6 for both CTRL and EXPT, which suggests that the regression method can extract the targeted signals. Values of both regressed precipitation and 850 hPa vorticity are larger in CTRL than EXPT (see Figs. 8c, d), which is consistent with Fig. 7, indicating that TSD is stronger in the low levels in CTRL because of the presence of extratropical forcing.

Meteorological variables at different vertical levels are further regressed onto $\tilde{\zeta}$, to the vertical structure of TSD from the model experiments. Figure 9 gives the vertical cross-sections of the regressed vorticity and temperature along the TSD wavetrain (see black solid lines in Figs. 8a, b). Prominent vorticity perturbations with deep vertical extent can be seen, with westward tilted structure (see Lau and Lau 1990). West of approximately 150°E, the temperature field is strongest in the 500–400 hPa layer, with anomalies almost in phase with those of the vorticity. At the 900–850 hPa levels, however, anomalous positive (negative) vorticity tends to be more collocated with cold (warm) perturbations. These circulation features, as seen in both model experiments, are consistent with the observed structure of TSD (Tam and Li 2006; Au-Yeung and Tam 2018). The differences of vorticity perturbations between the CTRL and EXPT vertical cross-sections are also computed (see Fig. 9c). The low-level vorticity perturbations in CTRL are significantly larger than those in EXPT west of approximately 140°E. Again, this points to the fact that extratropical forcing leads to stronger wave amplitudes in the more western part of the TSD wavetrain. However, it is also noteworthy that signals are found to be stronger in EXPT east of 140°E from the surface to approximately 500 hPa. In this region over the more eastern part of the wavetrain, stronger vortices anomalies can be seen in EXPT (see box region in Fig. 9c). It seems that wave activity in this domain is more influenced by the modulated circulation and background climate because of the presence of extratropical synoptic-scale forcing from the northern boundary.

We have also computed the anomalous pressure velocity ($\omega'$) and its correlation with $\tilde{\zeta}$, at different levels from CTRL and EXPT, as well as the difference in $\omega'$ between the two experiments (Fig. 10). Compared with Fig. 9, over the western part of the wavetrain, rising motion (subsidence) is seen to coincide with positive (negative) vorticity at approximately 500 hPa. Consistent with Fig. 9, anomalous rising/sinking west of 140°E have larger amplitudes in CTRL compared with EXPT, whereas stronger and more coherent vertical motion is found in EXPT at more eastern locations. The latter is clearly related to stronger coherence between $\omega'$ and $\tilde{\zeta}$ in that part of the TSD wavetrain. To summarize, although TSD becomes stronger overall in CTRL due to extratropical forcing, within 140–180°E the wavetrain is slightly stronger and with a more coherent circulation structure in EXPT.

Figure 11 gives CTRL minus EXPT JJA mean 200hPa geopotential height and circulation, with zonal wind contours included to indicate the presence of the tropical upper tropospheric trough (TUTT). The upper-level Asian high is stronger in EXPT under a warmer troposphere because of less northward eddy temperature transport at low levels (see Fig. 2). Consequently, there is a stronger zonal flow north of 20°N in EXPT. Enhanced westerlies and stronger stationary wave features might channel more wave activity and trigger TSD at the TUTT locations (e.g., Sadler 1967; Tam and Li 2006; Feng et al. 2020; Wang et al. 2020). Additionally, the TUTT axis is located more to the southeastern in EXPT (not shown, but see Fig. 11).
This and more active TUTT cells under stronger upper-level northeasterlies might also enhance low-level TSD in the western North Pacific (Wen et al. 2018; Guo and Ge 2018), which is probably why there are more coherent TSD signals east of 140°E in EXPT.

Additional analysis based on E-vectors associated with TSD formation is also conducted, following Tam and Li (2006), to evaluate the role of upper-level wave activity intrusion responsible to TSD. Anomalous 200 \( \mathbf{u} \)- and \( \mathbf{v} \)-wind related to TSD are first computed on the basis of lag regression onto the reference TSD index \( \hat{\zeta} \). The \( \mathbf{u}^2 \), \( \mathbf{v}^2 \) and \( \mathbf{u}^2 \mathbf{v} \) covariance maps and hence the corresponding TSD-related E-vectors are then found based on \( \mathbf{u}' \) and \( \mathbf{v}' \) regression for negative lags (Fig. 12). (Negative lag regression maps are considered because of their better depiction of triggering processes of TSD.) It can be seen that there are strong southward-pointing E-vectors west of 140°E, coming from the northern boundary, in CTRL. The E-vectors near the northern boundary indicate wave activities to the south from the extratropics; this makes a stronger wave signal of TSD west of 140°E in CTRL. Conversely, compared with CTRL, stronger southward directed E-vectors are seen in [10–25°N, 150–160°E] in EXPT. The E-vector characteristics in the latter further support our proposed mechanism of more extratropical influence that favors TSD formation east of 140°E, in relation to TUTT changes.

4. Discussions and conclusion

The sensitivity of western North Pacific summertime TSD to extratropical forcing has been examined, by
comparing a Regional Climate Model control run in boreal summer, with an experimental run having the same setting except with suppressed synoptic-scale variability at its northern boundary. Consistent with the above model design, in JJA southward directed upper-level E-vectors and low-level northward heat transport north of approximately 25°N was found to be reduced in the experimental run, compared with the control run. This means that in the former, mid-latitude synoptic-scale disturbances are indeed suppressed, with implications on the regional circulation within the model environment. Further comparison between the two experiments revealed that in the presence of extratropical forcing, there is stronger TSD activity over the western Pacific between 10°N and 25°N. Along the propagation path of TSD, precipitation is also intensified. Also, stronger southeastward low-level Rossby wave energy dispersion associated with TSD was found, leading to more active MRG waves at 850 hPa over the equatorial western Pacific sector. Hence, via the triggering of TSD, forcing related to extratropical synoptic-scale activity can lead to stronger tropical disturbance/MRG waves, as well as their associated rainfall in summertime western North Pacific.

The TSD-related anomalous circulation from the two experiments was further examined, by regressing various meteorological variables onto the standardized EOF-filtered low-level vorticity at 850 hPa over the equatorial western Pacific sector. Hence, via the triggering of TSD, forcing related to extratropical synoptic-scale activity can lead to stronger tropical disturbance/MRG waves, as well as their associated rainfall in summertime western North Pacific.

The TSD-related anomalous circulation from the two experiments was further examined, by regressing various meteorological variables onto the standardized EOF-filtered low-level vorticity at (15°N, 120°E). Compared with the experimental run, TSD wavetrain from the control run tends to be stronger west of 140°E, with stronger vortices and more vigorous anomalous vertical motion. Nevertheless, in the eastern portion (140–180°E) of the TSD wavetrain, relatively
Fig. 10. Same as Fig. 9, except for the pressure velocity (shading; units: $10^{-5}$ Pa s$^{-1}$, downward as positive) and its correlation coefficient with EOF-filtered 850 hPa vorticity at (15°N, 120°E) (contour). Dashed areas indicate where the differences are statistically significant above the 90% confidence level based on Student’s $t$-test. The green box in (c) indicates the region where anomalous vertical motion from EXPT is stronger than CTRL.

Fig. 11. CTRL minus EXPT JJA mean 200 hPa geopotential height (shading; units: m) and wind (arrows; see upper right for scale). Green (orange) solid line indicates the zero line of 200 hPa zonal wind in CTRL (EXPT).
Fig. 12. 200 hPa E-vector (arrows; units: m$^2$ s$^{-2}$) associated with 850 hPa EOF-filtered vorticity at (15°N, 120°E) summed for negative lags (~8 day to -1 day) only, and the meridional component (shading; units: m$^2$ s$^{-2}$). See text for details.

stronger and more coherent wave signals exist in the experimental run. Further inspection showed that in the latter run, the upper-level Asian high is stronger due to less northward eddy heat flux; it is possible that the strengthened anticyclone might channel more mid-latitude wave activity to TUTT, which excites more TSD. Diagnostics based on E-vectors associated with TSD formation support this view. Apart from the modulated upper-level flow, changes in the low-level mean circulation might also affect TSD characteristics. Additional diagnostics revealed that besides changes in the strength and structure of eddies, modulation in the low-level wind pattern also contributes to stronger barotropic energy conversion between background wind and TSD eddies (figures not shown). More studies can be conducted to better understand the intricate nature of such wave–mean flow interaction associated with TSD eddies.

This work helps to quantify the role of extratropical wave activity in determining TSD variability. In terms of low-level vorticity (MRG-related ψ-wind) anomalies over the far western north (equatorial) Pacific, suppression of extratropical forcing leads to ~20–25% reduction of their variance in the model environment. Although not the major contributor to the climatological mean activity, equatorward wave energy dispersion due to extratropical systems might still influence TSD amplitudes from year to year. This type of forcing can be sensitive, for instance, to the structure and location of the subtropical jet. More studies need to be carried out to better understand how mid-latitude forcing influences the sub-seasonal to seasonal TSD activity or in fact other related tropical systems such as TCs and how such forcing might depend on the background circulation.

Finally, it is worth mentioning that the Regional Climate Model can well capture the circulation structure of TSD, including their connection with low-level MRG waves through equatorward energy dispersion. It is possible that other equatorial wave types (for instance n = 1 equatorial Rossby waves, which are also important for TC genesis) might also be affected by mid-latitude forcing. Further observational and modeling studies will be conducted in this direction. Numerical experiments, similar to those in this study, can be designed for examining the sensitivity of these equatorial waves to various components of the extratropical circulation. This can be a useful method for studying this aspect of the tropical–extratropical linkage in the atmospheric general circulation.
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