The Linkage between the Eastern Pacific Teleconnection Pattern and Convective Heating over the Tropical Western Pacific

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ABSTRACT

The eastern Pacific (EP) pattern is a recently detected atmospheric teleconnection pattern that frequently occurs during late winter. Through analysis of daily ERA-Interim data and outgoing longwave radiation data for the period of 1979–2011, it is shown here that the formation of the EP is preceded by an anomalous tropical convection dipole, with one extremum located over the eastern Indian Ocean–Maritime Continent and the other over the central Pacific. This is followed by the excitation of two quasi-stationary Rossby wave trains. Departing from the subtropics, north of the region of anomalous convection, one Rossby wave train propagates eastward along the East Asian jet from southern China toward the eastern Pacific. The second wave train propagates northward from east of Japan toward eastern Siberia and then turns southeastward to the Gulf of Alaska. Both wave trains are associated with wave activity flux convergence where the EP pattern develops. The results from an examination of the E vector suggest that the EP undergoes further growth with the aid of positive feedback from high-frequency transient eddies. The frequency of occurrence of the dipole convection anomaly increases significantly from early to late winter, a finding that suggests that it is the seasonal change in the convection anomaly that accounts for the EP being more dominant in late winter.

1. Introduction

The North Pacific is a region where atmospheric low-frequency anomalies, also known as teleconnection patterns, are among the most energetic in the Northern Hemisphere during the winter season. These teleconnections have a far-reaching impact on tropospheric weather and climate over the North Pacific and its surroundings, including the stratosphere. Among these teleconnections, the Pacific–North American (PNA) and western Pacific (WP) patterns are best known and have been extensively studied for the past several decades (e.g., Wallace and Gutzler 1981; Barnston and Livezey 1987; Nigam 2003; Chiang and Vimont 2004; Linkin and Nigam 2008; Wettstein and Wallace 2010). In addition to the PNA and WP, a new teleconnection, known as the eastern Pacific (EP) pattern, was recently found by Athanasiadis et al. (2010, hereafter AWW10). In an empirical orthogonal function (EOF) analysis of daily zonal wind over the wintertime North Pacific sector (0°–87.5°N, 120°E–105°W), it was found that the EP is the second EOF during late winter [February–March (FM)]. [Note that a teleconnection pattern identified by the same name was found by Barnston and Livezey (1987) using a rotated EOF analysis of Northern Hemisphere monthly 700-hPa geopotential height. That pattern is similar but not completely identical to the EP pattern studied here.] AWW10 demonstrated that like the PNA and WP, the EP is closely related to anomalies...
in the jet configuration, storm track, and precipitation over western North America, and should be recognized as one of the main teleconnection patterns over the wintertime North Pacific.

Since the EP pattern has been detected only recently, our knowledge of this pattern is still limited. The goal of this study is to investigate the processes that drive its formation. As will be seen, the EP pattern is closely linked to convective heating over the western tropical Pacific. Section 2 introduces the dataset and the data processing methodology. Section 3 reports the main results of this study. This is followed by the conclusions in section 4.

2. Data and methodology

This study uses unfiltered daily fields archived in the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) dataset (Dee et al. 2011) for the winters (November–March) from 1979 to 2011. The data have a 1.5° × 1.5° horizontal resolution with 23 vertical levels. We use daily 0000 UTC data, rather than monthly or seasonal mean fields, to investigate the temporal variability of teleconnections at their intrinsic subseasonal time scale, as in Feldstein (2000, 2002) and AWW10. As will be shown, the EP pattern is closely linked to convective heating. As a proxy of the occurrence and intensity of convective heating, we will utilize the interpolated outgoing longwave radiation (OLR) data provided by the NOAA/OAR/ESRL Physical Sciences Division (PSD) (Liebmann and Smith 1996). [Although OLR corresponds to the infrared radiation emitted to space at the top of the atmosphere, for the tropics, OLR corresponds to an estimate of the temperature at the tops of clouds. Since low (high) OLR values indicate low (high) cloud-top temperatures, low OLR values correspond to the presence of deep convection and strong convective heating and vice versa for large OLR values.] The seasonal cycle is removed from all daily data. The seasonal cycle is defined by taking the calendar mean of each day from 1979 to 2011, followed by the calculation of a 21-day weighted mean. The daily anomaly is obtained by subtracting the seasonal cycle on that calendar day from the raw daily field at each grid point.

The EP is defined as the second EOF (EOF2) of the daily 250-hPa zonal wind for the same North Pacific domain as in AWW10 (see section 1) during FM. The variance explained by EOF2 is 9.2%. The data are weighted by the square root of cosine of latitude in the EOF analysis. We use EOF analysis to define the EP pattern in order to be consistent with AWW10. Later, as we will find, the EP pattern is also obtained from self-organizing map (SOM) analysis (Kohonen 2001). The standardized principal component time series of the second EOF will be used as an index for the amplitude of the EP pattern. This time series will be referred to as the EP index. To examine if the tropical heating pattern associated with the EP is a pattern that occurs in nature, we apply SOM analysis to the OLR anomalies over the entire equatorial band between 30°N and 30°S. One of the major advantages of SOM analysis compared with EOF analysis is that the SOM patterns more closely resemble the daily fields (i.e., the pattern correlations between the SOM patterns and the daily fields are typically greater than that between EOF patterns and the daily fields, and are not very sensitive to the domain size). Another advantage is that each day is associated with a particular SOM pattern (i.e., the SOM pattern with the smallest Euclidean distance with the daily field). This enables one to assign an occurrence frequency for each SOM pattern. SOM analysis is particularly useful for examining a continuum of spatial patterns (Kushnir and Wallace 1989; Franzke and Feldstein 2005). For daily fields that are well described by a continuum, the occurrence frequency and intrapattern Euclidean distances for the SOM patterns varies gradually. On the other hand, if EOF analysis was preferable, the occurrence frequency would have to peak and the intrapattern Euclidean distances would have to be a minimum for the SOM patterns that most closely resemble the EOF. As is discussed in Johnson and Feldstein (2010), this typically does not occur for atmospheric quantities. SOM analysis has been widely used in the meteorological community during recent years (e.g., Johnson et al. 2008).

Several of the calculations performed in this study use linear regression. The method that we use to evaluate statistical significance for linear regression follows that of Kosaka et al. (2012), where the number of effective degrees of freedom $N_{\text{eff}}$ corresponds to

$$N_{\text{eff}} = \frac{N}{1 + 2 \sum \left(1 - r_x/N[r_x(\tau)r_y(\tau)]\right)},$$

(1)

where $N$ is length of time series $X$ and $Y$, and $r_x$ and $r_y$ are the corresponding autocorrelation functions for $X$ and $Y$, with a lag of $\tau$ days.

3. Results

Figures 1a–d show the zonal wind field at 250 hPa, a cross section of the zonal wind along 150°W, and the geopotential height fields at 250 and 1000 hPa, respectively, for FM, regressed against the EP index. The 250-hPa zonal wind is characterized by a tripole pattern...
in the jet exit region of the eastern Pacific. The EP assumes a wave train spatial structure with a northeast–southwest tilt. Compared to the PNA and WP (see AWW10), the EP is located farther eastward. The tropical center of the EP is confined to the upper troposphere, and is weaker than the subtropical and midlatitude centers. The subtropical and midlatitude centers are deep in the vertical direction, extending from the surface to the lower stratosphere with a maximum perturbation wind speed of approximately 10 m s\(^{-1}\) (Fig. 1b). Because of its northeast–southwest tilt, west of the date line, the positive phase of the EP represents a strengthened and extended subtropical jet with a reduced zonal wind speed south of the subtropical jet, and to the east of the date line, the EP corresponds to a poleward shift of the eddy-driven jet. For the negative EP phase, each of these features is reversed.

For geopotential height field, the EP takes on a wave train form with a northeast–southwest tilt centered at 55°N, 168°W and at 37°N, 145°W, extending throughout the entire troposphere (Figs. 1c and 1d). In contrast, the PNA is a north–south dipole over the North Pacific in the upper troposphere, and a monopole centered south of the Aleutian Islands at the surface, while the WP is a northeast–southwest dipole throughout the entire troposphere (Linkin and Nigam 2008).

How does the EP develop? As a starting point, we examine wave activity fluxes (Takaya and Nakamura 2001). Since these fluxes are parallel to the group velocity, the wave activity flux indicates the direction of energy propagation. Figure 1c suggests that the wave activity associated with the EP propagates in a south–eastward direction. This apparently suggests that the EP's energy source is located in the Bering Sea, but time-lagged regression of the wave activity fluxes, shown in Fig. 2, indicate otherwise.

As can be seen, well before the formation of the EP, up until lag –13 days (Fig. 2b), weak wave activity fluxes are present over eastern Asia and the northeastern Pacific. From lag –13 to lag –10 days, over the northeastern Pacific, there is a dipolelike geopotential height anomaly pattern that develops (as expected, this pattern is in quadrature with the zonal wind tripole that defines the EP pattern), which corresponds to the incipient EP pattern [hereafter EP incipient (EPI)], while over eastern Asia, during the same time interval a weak but noticeable wave train is clearly seen. Departing from the source region centered at 27°N, 105°E, as evidenced by wave activity flux vectors (lag –13 days), the wave train propagates eastward along the East Asian jet over the North Pacific. When this wave train reaches the EPI area, there is wave activity flux convergence, and the EP anomaly pattern amplifies. This coincides with the wave train splitting into two branches, as wave activity fluxes continue both northward and southward. Then, between day –10 and day –7, over eastern Asia, the wave train

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FIG. 1. Regressions on the second principal component (PC2) for (a) the 250-hPa zonal wind anomaly, (b) the zonal wind anomaly in pressure–latitude cross section through 150°W, and geopotential height at (c) 250 and (d) 1000 hPa. Red (blue) contours show positive (negative) values and the arrows in (c) indicate wave activity fluxes. In (a) and (b), black dashed contours denote the climatology of the zonal wind. The contour interval is 2 m s\(^{-1}\) for regression of the anomalous zonal wind anomaly, 10 m s\(^{-1}\) for the climatology of the zonal wind, and 20 (10) m for regression of geopotential height at 250 (1000) hPa. Shading denotes statistically significant values at the 99% confidence level based on a Student's \(t\) test.
turns predominately northward from east of Japan toward eastern Siberia, and then gradually develops into a mature wave train, referred to as the East Asian wave train (EAW), which plays an important role in the EP formation process. This wave train continues to propagate eastward and then southeastward. Over the Gulf of Alaska, the southeastward wave activity flux associated with the EAW meets with the northward-branching wave activity flux discussed above (i.e., there is wave activity flux convergence) and additional enhancement of the EP anomalies. This is followed by the retrogression and rapid development of the negative geopotential height anomaly over western Canada, which becomes the northern center of the EP pattern, along with eastward movement and further development of the southern EP center. Consistently, the EAW is observed to undergo downstream development from day −7 to day 0, as the upstream anomalies over the western Pacific decay and those over the eastern Pacific strengthen. At day −4, the EAW reaches its maximum amplitude, after which the EAW weakens and the EP continues to undergo rapid growth reaching its maximum amplitude by day 0. Then, the EP weakens gradually and disappears after day +5.

Further examination suggests that there is a positive feedback from high-frequency transient eddies (eddies with a period less than 10 days) to the low-frequency EP that also plays an important role in the growth of the EP pattern. The high-frequency transient eddies are obtained by applying a 101-point Lanczos filter (Duchon 1980).
with a nominal frequency cutoff period of 10 days to the data. Figure 3 shows the lagged regressions of the zonal wind and the divergence of the pseudo $E$-vector $\left(y^2 - u^2, -v'u\right)$ at 250 hPa onto the EP index, where the primes denote the high-frequency (period less than 10 days) components of the wind field. As is shown in Hoskins et al. [1983, Eq. (A8) of their paper], the local zonal wind tendency is approximately proportional to the convergence of the $E$ vector. As can be seen in Fig. 3, an $E$-vector divergence first appears over the northeastern Pacific at about lag $-7$ days, after the EP pattern is established. The $E$-vector divergence strengthens systematically until lag $-2$ days, and projects onto the EP pattern. This behavior suggests that the after the excitement of the EP pattern by the wave trains, further amplification of the EP pattern occurs via a feedback with the high-frequency transient eddies. Such behavior is reminiscent of that for the PNA teleconnection pattern (Feldstein 2002; Franzke et al. 2011). Furthermore, an examination of the low-frequency (period greater than 10 days) component to the $E$ vector finds that these eddies further reinforce the EP pattern prior to lag day 0, but that they contribute to the decay of the EP pattern after day 0 (not shown).

The above result shows that the EP formation is closely related to the propagation of wave trains from
eastern Asia. Then, we ask, how are these wave trains excited? In view of the fact that the origin of these wave trains are in the subtropics near the Maritime Continent, where convective heating has its largest variability, it is plausible that the wave trains are excited by convective heating anomalies over this region (e.g., Sardeshmukh and Hoskins 1988; Nitta 1987; Zheng et al. 2013). To test this hypothesis, lagged regressions of the OLR anomalies against the EP index are calculated (see Fig. 4). As can be seen, a dipolelike anomaly pattern is observed, which indicates enhanced convective heating over the eastern Indian Ocean–Maritime Continent and reduced convective heating over the central tropical Pacific. This signal appears as early as lag day $-20$. Between lag day $-16$ and lag $-2$ days, the OLR anomalies retain their intensity, with the negative OLR extending slightly southward and eastward. After lag day 0, the OLR anomalies gradually weaken. [To evaluate how well OLR describes convective heating, we have regressed the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) precipitation against the EP index. A very similar dipole anomaly as that for OLR is found over the tropical Indian Ocean and western Pacific (not shown).] The time period of the maximum convective heating overlaps with the growth and persistence of the wave trains, and suggests that the positive EP may be driven by enhanced convection over the Maritime Continent and reduced convection over the central tropical Pacific, followed by wave train propagation toward the EP region, and vice versa for the negative EP, with the wave train anomalies taking on the opposite sign.

Next, we address the questions of 1) whether the EP pattern occurs in response to dipole anomalies in tropical convection of the form shown in Fig. 4, and 2) whether the EP pattern is more prominent during late winter because of an increase in the frequency of these dipole convection anomalies from early to late winter. For this purpose, we apply SOM analysis to the OLR field. In Fig. 5, we show 20 SOM patterns. [We choose 20 SOM patterns as a compromise between accuracy and convenience. As discussed in Johnson et al. (2008), as the number of SOM patterns is increased, the average of the pattern correlations between the SOM pattern with the minimum Euclidean distance for a given day and the daily field for the corresponding day becomes larger [see Table 1 of Johnson et al. (2008)]. An analysis with different numbers of SOM patterns (e.g., 16 and 25 SOM patterns) yields essentially the same results as those presented below (not shown). However, if too many SOM patterns are chosen, it is not convenient to display these patterns and the number of days that each SOM pattern is observed becomes too small.] To address the first question, the following procedure is performed. We
first calculate the pattern correlation between each of the 20 OLR SOM patterns and the OLR regressed against the EP index at lag 7 days. This time lag was chosen because it takes about one week for a Rossby wave train excited by tropical convection to reach the middle and high latitudes (Hoskins and Karoly 1981). We identify those pattern correlations with a value greater (less) than 0.6 (−0.6) with the positive (negative) EP phase. For the positive phase, SOM patterns 1, 6, and 11 were selected and for the negative phase patterns 10, 15, and 20 were chosen. We identify OLR–SOM events by using the following criteria: first, an event for a particular SOM pattern takes place when there are four or more consecutive days that the SOM pattern has a Euclidean distance that is smaller than that for all the other SOM patterns. We define the “lag 0 day” as the day with the least Euclidean distance during the event. If the distance between two neighboring lag 0 days is less than 10 days, we discard the event with the larger Euclidean distance. Frequency-weighted zonal wind composites based upon the SOM events are then calculated separately for the positive and negative phase, and then the difference (positive minus negative phase) is determined. This calculation was performed for the full winter season (November–March), November–December (ND), and FM (the left column of Fig. 6). As can be seen, similar EP-like patterns are found for each time period. [Linearity for the composite zonal wind field is found for the entire winter, ND, and FM, as the composite 250-hPa zonal wind field based on the positive (negative) phase SOMs closely resembles the positive (negative) 250-hPa zonal wind field associated with the EP pattern (the center and right columns of Fig. 6).] This result suggests that the EP pattern does indeed arise in response to the dipole OLR anomalies shown in Fig. 4. However, it should be noted that these composite zonal wind anomalies exhibit a slight northwest–southeast tilt, whereas the EP zonal wind anomalies in Fig. 1 show a northeast–southwest tilt. Since the OLR anomalies will be shown to explain about 15% of the EP time series, this result suggests that although tropical convection plays an important role in exciting the EP pattern, other factors, such as driving of the EP pattern by high-frequency transient eddies (Fig. 3), which generates a northeast–southwest wave tilt over the northeastern Pacific, are important too.
To evaluate how much variance of the EP pattern is explained by the dipole tropical convection anomalies, we correlate the EP index with amplitude time series (referred to as projection time series) for SOM patterns 1, 6, 10, 11, 15, and 20 (the SOM patterns discussed in the above paragraph). The projection time series are obtained as

$$P_n = \frac{\sum_i \sum_j OLR(i, j, t) SOM_n(i, j) \cos \theta}{\sum_i \sum_j SOM_n^2(i, j) \cos \theta},$$

(2)

where $i$ and $j$ correspond to the grid points in the zonal and meridional direction, $t$ is time, $OLR(i, j, t)$ is the daily outgoing longwave radiation field, $SOM_n(i, j)$ is the $n$th SOM pattern, and $\theta$ is latitude. The time over which the autocorrelation of the projection time series decays to $1/e$ is defined as the $e$-folding time scale. (The corresponding $e$-folding time scales vary from 3 to 14 days, with a mean close to 8 days.) The absolute value of the correlations between the EP index and the SOM projection time series varies between 0.20 and 0.36. A second correlation is performed between the EP index and another OLR time series that is obtained by projecting the daily OLR onto the lag 7 day regressed OLR pattern shown in Fig. 4. This OLR pattern has the structure of a zonal dipole. Since the correlation is found to have a value of 0.39, which is statistically significant at
the 95% confidence level, we conclude that the zonally oriented anomaly in tropical convection explains about 15% of the variance (correlation squared) of the EP index.

To examine if these SOM patterns show similar persistence as the regressed OLR in Fig. 4, we examine the transitions between SOM patterns. That is, for each SOM pattern, for those days that a particular SOM pattern has the minimum Euclidean distance, the frequency of occurrence of all SOM patterns is calculated for both positive and negative lags. Using this approach, it is found that all SOM patterns described in the above paragraph persist from at least lag −5 through lag +5 days, with some SOM patterns being more persistent (not shown). For a few SOMs (e.g., SOMs 11 and 15) the most frequent SOM pattern either prior to lag −5 days or after lag +5 days is an adjacent pattern in the SOM grid. This indicates that the SOM patterns are associated with very weak propagation in the OLR field.

From Fig. 5, it can be seen that all the dipolelike SOMs listed above underwent an increase in their frequency from early to late winter. SOM 1 increases its frequency from 4.9% during ND to 5.1% during FM, SOM 6 from 3.6% to 7.1%, SOM 11 from 4.1% to 6.4%, with the average increase being 50%, while SOM 10 increases its frequency from 3.1% to 5.6%, SOM 15 from 2.9% to 5.3%, and SOM 20 from 3.6% to 6.6%. Since, as described above, the composite zonal wind field associated with these SOM patterns projects onto the EP pattern and accounts for about 15% of its variance, these increases in SOM frequency from ND to FM suggest that EP may have its largest variance in FM (and the WP pattern having its largest variance in ND) (AWW10) because of the increase in the frequency of days from early to late winter with dipole tropical convection anomalies. However, an examination of statistical significance of the change in the SOM pattern frequency using a Monte Carlo approach found that SOM 6 was the only pattern to undergo a change from early to late winter that exceeded the 90% confidence level. An analogous Monte Carlo test for the statistical significance of the change in the frequency for the group of positive OLR SOMs (SOMs 1, 6, and 11) and negative OLR SOMs (SOMs 10, 15, and 20) was also performed. Neither group showed a change in statistical significance at the 90% confidence level. However, the negative phase SOM group was found to be statistically significant at the 89% confidence level. (For these Monte Carlo tests, the day that each of the above SOM patterns had the minimum Euclidean distance was randomly reassigned and the change in frequency calculated. This was performed 1000 times.) This suggests that although the preference for the EP pattern in late winter is likely linked to the observed increase in the frequency of dipole convection anomalies, these early to late winter increases in the frequency of dipole convection anomalies are not greater than that to be expected for random variation.

Interestingly, the spatial structure of the OLR anomalies in Figs. 4 and 5 resembles that for La Niña, and these anomalies are very persistent. This observation leads to the question of whether El Niño–Southern Oscillation (ENSO) influences the relationship between the tropical dipole and the frequency of occurrence of the EP pattern and the preference for the EP pattern to be active during late winter. To address this question, EP events are defined with the criteria that the EP index exceeds an amplitude of 1.0 standard deviation for four or more consecutive days, and each event must be at least 21 days apart. Also, El Niño and La Niña events are defined following the definition used by the National Oceanic and Atmospheric Administration/Climate Prediction Center (http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml). As can be seen from Table 1, during ND, positive EP events occur mostly during La Niña, and negative EP events occur during El Niño. For FM, there is again a strong preference for negative EP events to take place during El Niño, whereas positive EP events occur with the same frequency in both El Niño and La Niña. Furthermore, for neutral ENSO, both positive and negative EP events are observed to occur at intermediate frequencies both in ND and in FM. These results suggest that although EP events and the OLR fluctuate at the intraseasonal time scale (as discussed above, both the EP index and OLR projection time series have short e-folding times), the occurrence of active El Niño and La Niña events appears to modulate the frequency of EP events. Furthermore, during neutral ENSO (Table 1), we note that EP events occur more often during FM than ND. This suggests that the greater prominence of the EP pattern during late winter is unrelated to ENSO.

A calculation of the February–March mean EP index for the years 1979–2011 shows a slight upward trend, especially after 1995 (not shown). Furthermore, given

### Table 1. EP events detected in early and late winter and during El Niño, La Niña, and neutral years for 1979–2011. EP+ represents positive-phase EP events, and EP− represents negative-phase EP events. See the text for the definition of EP events.

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the above linkage between positive (negative) EP events and La Niña (El Niño), and the recent ENSO trend toward La Niña–like SST patterns (Johnson 2013), the interdecadal upward trend in the EP index can also be understood as arising in part by ENSO modulating the frequency of EP events.

4. Summary and discussion

Based on both ERA-Interim and OLR daily data for the winters from 1979 to 2011, this study shows that the newly found eastern Pacific (EP) teleconnection pattern extends from the surface to the lower stratosphere, and is associated with changes to the strength and zonal extent of the subtropical jet to the west of the date line and with meridional shifts in the latitude of the eddy-driven jet to the east of the date line.

This study focuses on the dynamical processes that excite and maintain the EP pattern. It is found that two distinct wave trains appear to be crucial for the EP formation. One wave train, at an earlier lag, propagates eastward across the North Pacific from eastern Asia toward the EP region, and this is followed by a second wave train that propagates northeastward from east of Japan to Siberia and then southeastward toward EP region. Both wave trains are associated with wave activity flux convergence and growth of the EP pattern. Furthermore, for the positive EP phase, these wave trains are preceded by enhanced tropical convection over the eastern Indian Ocean and Maritime Continent and reduced convection over the central tropical Pacific, and vice versa for the negative EP. After the EP pattern is established, it is found that driving by high-frequency (period less than 10 days) transient eddies further amplifies this pattern, likely through a positive feedback, as has been observed for other teleconnection patterns in both the North Pacific and North Atlantic.

Many previous studies have examined the mechanisms that drive extratropical teleconnections. These include 1) growth due to linear dispersion from a topographic or diabatic heating source (e.g., Hoskins and Karoly 1981; Karoly et al. 1989; Hoskins and Ambrizzi 1993; Li and Wettstein 2012), 2) barotropic growth due to zonal asymmetric variation of the background climatological flow (e.g., Simmons et al. 1983; Feldstein 2002), and 3) high-frequency transient eddy feedback (e.g., Lau 1988; Nakamura and Wallace 1993). The results of the present study suggest that mechanisms 1 and 3 play an important role in driving the EP pattern. (The role of mechanism 2 was not examined.) With regard to the first mechanism, previous modeling and observational studies have found that extratropical teleconnections are influenced by intraseasonal tropical convection (Liebmann and Hartmann 1984; Ferranti et al. 1990; Matthews et al. 2004; Yuan et al. 2011). Other more recent studies such as Mori and Watanabe (2008), Johnson and Feldstein (2010), Moore et al. (2010), and Roundy et al. (2010) show that the PNA is followed by MJO convection. Nitta (1987), Huang and Li (1988), and Zheng et al. (2013) found that convective heating over the tropical western Pacific may influence the summer and winter circulation over eastern Asia through the excitation of a stationary Rossby wave train that propagates northward from the Philippines toward eastern Asia and then to western North America. In this study, we show that a similar wave train can also excite the EP pattern.

As discussed above, the wave trains that appear to drive the EP pattern are preceded by dipole convective heating anomalies. Convective anomalies with this structure are found to be more common during February and March than in November and December. Since the EP pattern also occurs more frequently during February and March than in November and December, this finding lends support to the conclusion that the seasonal shift from the WP to EP pattern is driven in part by an increase in the frequency of the above dipole convection from early to late winter. In this study, we did not address whether one or both centers of the dipole-like convective heating anomaly are relevant for exciting the wave trains and subsequently the EP pattern. In addition, the seasonal variation of the background flow may also influence the excitation of the EP pattern. For future research, we plan to address these questions with an idealized numerical model, such as with the dynamical core of a climate model as in Yoo et al. (2012).

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