Winter-to-winter recurrence of atmospheric circulation anomalies in the central North Pacific

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[1] Possible causes of the winter-to-winter recurrence (WWR) of atmospheric circulation anomalies in the central North Pacific (CNP) are investigated in the present study. Results show that tropical ENSO could not lead to the atmospheric WWR in the CNP because the persistence of ENSO itself does not show any recurrence regardless of the starting month. The effect of other external forcing, e.g., sea ice, is also not significant. These results suggest that the dominant source of the atmospheric WWR may come from internal atmospheric dynamics in the North Pacific. The Arctic Oscillation, the dominant pattern of sea level pressure variations north of 20°N, seems not to be the cause of atmospheric WWR in the CNP region. The effect of the local internal atmospheric dynamics on the atmospheric WWR may be more important in the CNP region. The CNP region was in the location of the storm track in the North Pacific. It was found that seasonal variability of storm track anomalies and associated synoptic transient eddy dynamics may be one of the causes for the atmospheric WWR. During the WWR years, transient eddy forcing on the mean flow is strong during the winter but very weak in the intervening summer, which leads to a quick transition of anomalous mean atmospheric circulation around March and the maintenance of the opposite sign anomalies for two to three seasons. But this characteristic of transient eddy forcing does not exist during the non-WWR years.

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1. Introduction

[2] The huge thermal capacity of the ocean enables sea surface temperature (SST) variation to possess obvious lag and persistence characteristics. The persistence of sea surface temperature anomalies (SSTA) has a strong seasonal dependence in the midlatitude ocean. *Namias and Born* [1970, 1974] examined the winter-to-winter SSTA lag autocorrelations, showing a tendency to recur from one winter to the next without persisting through the intervening summer. SSTA persistence with this characteristic was called winter-to-winter recurrence (WWR). *Namias and Born* [1970, 1974] speculated that this seasonal dependence of SSTA persistence is closely tied to the seasonal variation of oceanic mixed layer depth (MLD). Late winter ocean temperature anomalies are sequestered beneath the shallow summer mixed layer and are reincorporated into the deepening fall mixed layer. *Alexander and Deser* [1995] termed it the "reemergence mechanism." Subsequent studies have confirmed the WWR and reemergence mechanism occurs across much of the North Pacific [*Alexander et al.*, 1999; *Deser et al.*, 2003; *Zhao and Li*, 2010].

[3] WWR does not only exist in the SSTA in the North Pacific. *Zhao and Li* [2010, 2012], from the perspective of mean climatic characteristics and interannual variability, found that the WWR in the central North Pacific is an evolutional characteristic of the whole air-sea system with the seasons, and the WWR of atmospheric circulation anomalies and its forcing play an important role in the SSTA WWR. Through lag autocorrelation analyses used in the previous studies, *Zhao and Li* [2010] indicated that, in addition to the SSTA, the atmospheric circulation anomalies also has the WWR in the central North Pacific from the lower layer to the upper layer. And it is one of the causes of the SSTA WWR because of the dominance of atmospheric forcing of the underlying ocean [*Cayan*, 1992]. If anomalous atmospheric forcing was to occur repeatedly for several consecutive winters, but not in summer, this would tend to create recurring SSTA in winter. Different from lag autocorrelation analyses in the previous studies, *Zhao and Li* [2012] investigated interannual variability of the WWR in the central North Pacific. And they indicated that atmospheric

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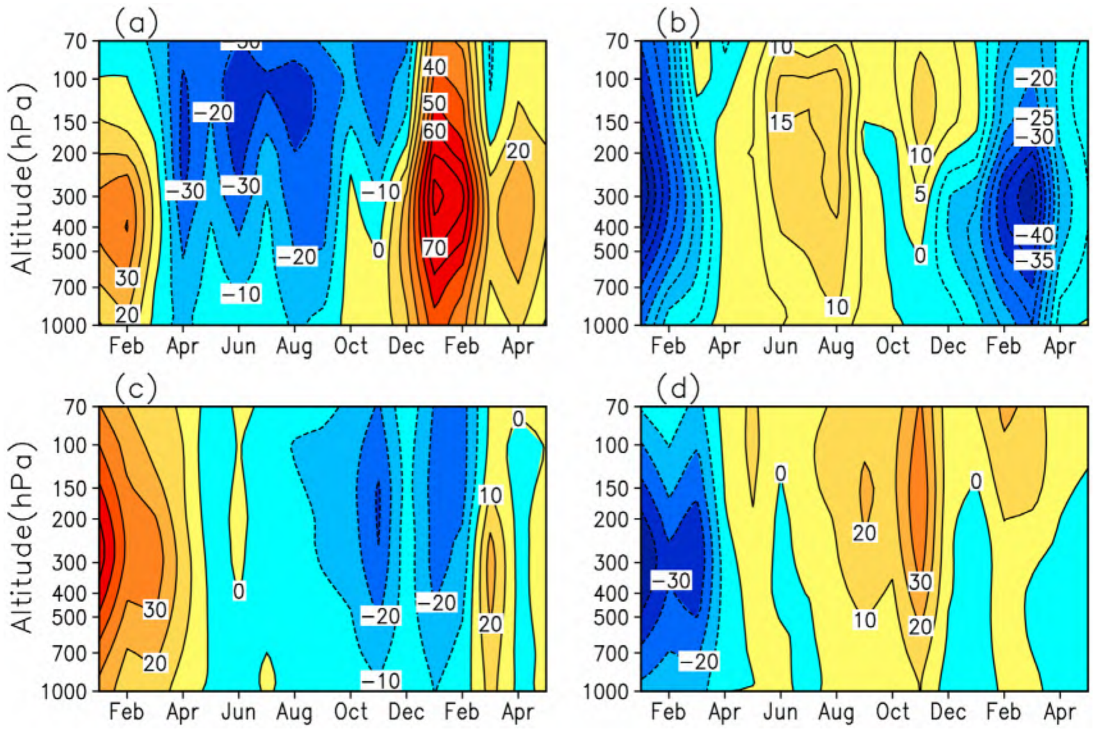

Figure 1. Time-altitude profiles of composite geopotential height anomalies from 1000 to 70 hPa in the CNP region during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

circulation anomalies exhibit the WWR phenomenon during WWR years, but do not recur in the following winter during non-WWR years. Through comparing the relative roles of atmospheric forcing and the oceanic reemergence mechanism during WWR years and non-WWR years, the important role of the WWR of atmospheric circulation anomalies in the SSTA WWR was further validated.

[4] Furthermore, the results of *Zhao and Li* [2012] showed that, although the winter-to-winter lag autocorrelations of geopotential height anomalies show a tendency to recur from one winter to the next without persisting through the intervening summer, it does not mean that the recurrent atmospheric circulation anomalies in the second winter come from those in the previous winter. Unlike the reemergence mechanism of the SSTA WWR, winter atmospheric anomalies do not persist at a certain layer of atmosphere through the intervening summer, suggesting that the recurrent atmospheric circulation anomalies in the second winter do not come from those in the previous winter. Thus, it seems that mechanisms of the WWR are markedly different between the atmosphere and ocean. While persistence is the key mechanism for SSTA WWR, it is not for the WWR of atmospheric circulation anomalies.

[5] However, the cause of the atmospheric recurrence is still an open question. In the extratropics, the atmosphere tends to drive the ocean, especially in winter [e.g., *Davis*,

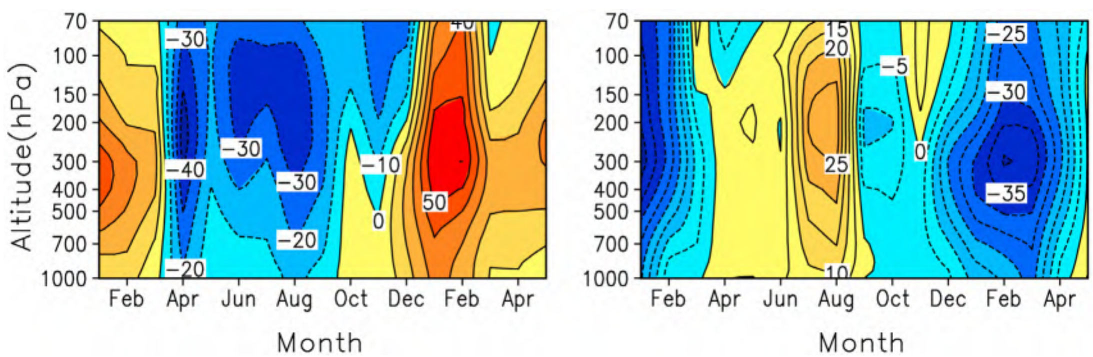

Figure 2. As in Figure 1a and 1b, but the ENSO signal is subtracted from the original data.

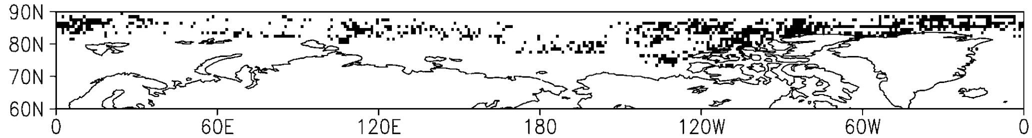

Figure 3. The spatial distribution of the WWR of the sea ice anomalies in the Northern Hemisphere for February starting months.

1976, 1978; *Wallace and Jiang*, 1987; *Zhao and Li*, 2010]. While addressing the question as to what extent the oceanic reemergence mechanism affects the overlying atmosphere through model experiments [*Bhatt et al.*, 1998; *Liu et al.*, 2007; *Cassou et al.*, 2007], coupling does not quantitatively change the structure of the patterns of atmospheric variability over midlatitude [*Bhatt et al.*, 1998]. Moreover, the monthly atmospheric response to SST is much weaker than the atmospheric internal variability in the midlatitude [*Liu et al.*, 2007], and we have to keep in mind that the dominant source of the leading mode of atmospheric variability is internal atmospheric dynamics [*Cassou et al.*, 2007]. Therefore, the cause of the atmospheric recurrence from the lower layer to the upper layer is an interesting question.

[6] The present study will focus on the following questions: if the ocean-atmosphere coupling would not be at the origin of the WWR of the atmospheric circulation, is the atmospheric persistence associated with or maintained by external forcing, for instance, the tropics? Does it come from the atmospheric response to the winter anomalies in other physical variables (e.g., snow cover, and sea ice)? *Kushnir et al.* [2002] stated that external forcing all together could explain at most 20–25% of the interannual variance of atmosphere at middle and high latitudes. Does the dominant source of the atmospheric WWR come from internal atmospheric dynamics, especially for storm tracks and associated synoptic transient eddies, which allow the possibility that the heat and momentum or vorticity fluxes may help to reinforce and maintain the anomalous mean circulation in the North Pacific [e.g., *Hoskins et al.*, 1983]? Moreover, the relationship of atmospheric WWR with the Arctic Oscillation (AO), the dominant pattern of nonseasonal sea level pressure (SLP) variations north of 20°N, also needs to be discussed.

[7] The remainder of this manuscript is organized as follows. The data sets and methods used are described in section 2, and the possible causes of the atmospheric WWR in the central North Pacific are described in section 3. Finally, a summary is provided in section 4.

2. Data and Methodology

2.1. Data

[8] The atmospheric data is from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP-NCAR) reanalysis data [*Kalnay et al.*, 1996] for the period 1950–2004 on a 2.5° × 2.5° grid. Monthly sea ice data for the period 1950–2004 are obtained from the Hadley Centre Sea Ice and SST (HADISST) [*Rayner et al.*, 2003] on a 1° × 1° grid. And 36 years (1972–2007) of snow cover data are provided by the Rutgers University Climate Lab (RUCL) [*Robinson et al.*, 1993], including snow extent for Eurasia, North America and Northern Hemisphere, which is available online at http://climate.rutgers.edu/snowcover/. The annual cycle of each variable is removed by subtracting the climatological mean monthly value at each grid point.

2.2. Methodology

[9] *Zhao and Li* [2010] indicated that atmospheric circulation anomalies display a significant WWR in the central North Pacific Ocean, closely related to the SSTA WWR near 40°N. Furthermore, the area average over the central North Pacific (CNP: 165°E–160°W, 35°N–47°N) is used by *Zhao and Li* [2012] to measure the interannual variability of the WWR in the North Pacific. We start from the original definition of the WWR (the mean climatic characteristic) to determine whether the WWR exists in the CNP region each year. Previous studies used lag correlation analysis to define the WWR [e.g., *Alexander et al.*, 1999]. The lag correlations between monthly anomalies for February and monthly anomalies for each subsequent month through February of

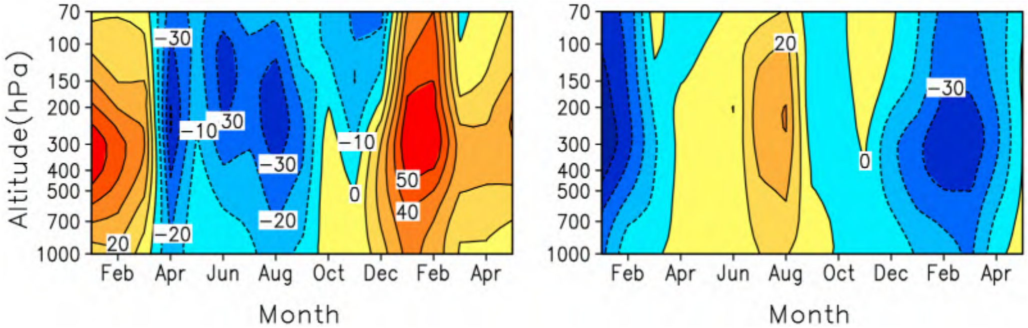

Figure 4. As in Figures 1a and 1b, but the sea ice–related anomalies are subtracted from the original data.

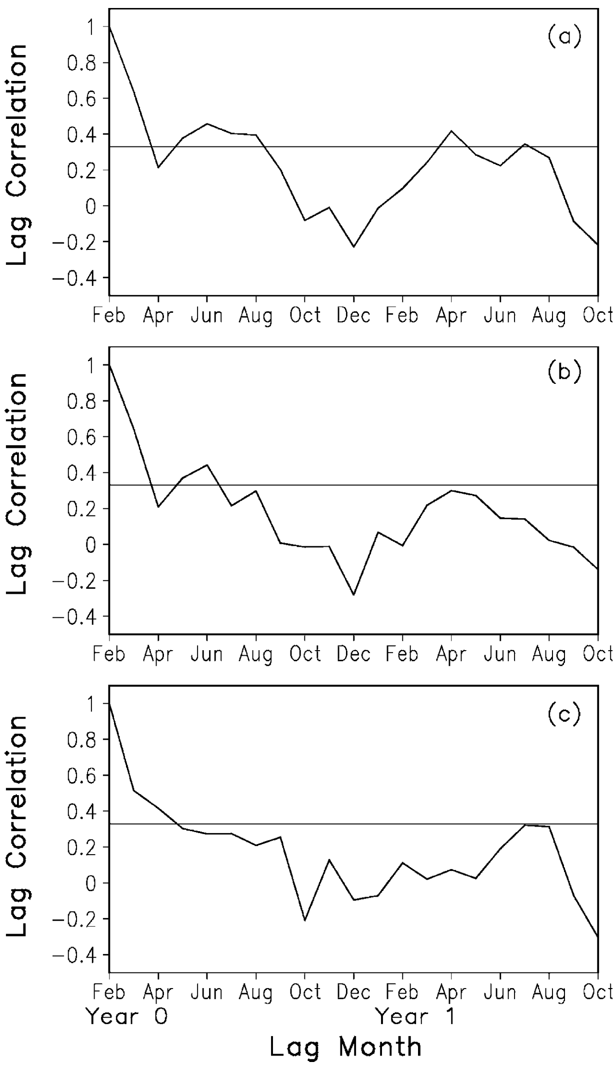

Figure 5. Lag correlations between snow cover in February and monthly snow cover from February of the current year through October of the following year for (a) the Northern Hemisphere, (b) Eurasia, and (c) North America. The thin solid line indicates the 95% confidence level.

the year after next year were calculated in the CNP region. The lag correlations have two significant characteristics: a significant decline during the following summer and an increase again during the following winter, which was called the WWR [e.g., *Alexander and Deser*, 1995]. To efficiently detect the WWR and non-WWR years, its two characteristics are quantitated for each year by using the following criteria: (1) for positive (negative) anomalies during the winter, winter anomalies are greater (less) than anomalies during the following summer; (2) the following winter anomalies are greater (less) than anomalies in the preceding summer and have the same sign as anomalies in the preceding winter. It is a positive (negative) WWR year if a year meets the two criteria and a positive (negative) non-WWR year if it does not. In this way, *Zhao and Li* [2012] identified 18 WWR years (positive cases: 1951, 1956, 1965, 1966, 1968, 1971, 1974; negative cases: 1959, 1960, 1977, 1978, 1983, 1985, 1986, 1994, 1996, 1998, 1999) and 36 non-WWR years (positive cases:1950, 1952, 1953, 1954, 1955, 1957, 1962,1963, 1967, 1969, 1972, 1976, 1982, 1989, 1990, 1991, 1993, 2000, 2002; negative cases:1958, 1961, 1964, 1970, 1973, 1975, 1979, 1980, 1981, 1984, 1987, 1988, 1992, 1995, 1997, 2001, 2003) in the CNP region during the period 1950–2003.

[10] Figure 1 shows time-altitude profiles of composite geopotential height anomalies between 1000 and 70 hPa in the CNP region during the WWR and non-WWR years. Note that the climatological seasonal cycle of geopotential height has been removed from the monthly values prior to the calculations. For the positive cases of the WWR (Figure 1a), the seasonal evolution is characterized by two reversals in the sign of geopotential height anomalies in the CNP region. The geopotential height anomalies in the first winter are positive with a maximum in January–February; they change to negative in the following summer with a maximum in June–August; then they return to positive again in the second winter. The geopotential height field in the CNP region displays WWR from the lower layer to the upper layer during the WWR years, which shows an equivalent barotropic vertical structure and the centers of anomalies are located in the high troposphere (500–300 hPa). Therefore, unlike the reemergence mechanism of the SSTA WWR, the recurrent atmospheric circulation anomalies in the second winter do not come from those of the previous winter through anomalies in the intervening summer. It seems that mechanisms of the WWR are markedly different between the atmosphere and ocean. Seasonal evolution of the geopotential height anomalies is similar during the negative cases of the WWR (Figure 1b). But the evolution of the atmospheric circulation anomalies during the non-WWR years (Figures 1c and 1d) differs markedly from that during the WWR years. Atmospheric circulation anomalies in winter do not recur in the following winter. Based on these results, the possible causes of the atmospheric in the CNP are investigated in section 3.

3. Results

3.1. Link With the Tropical Pacific

[11] It is well known that ENSO is the leading mode of interannual variability of the climate system, and it has a significant impact on global climate variability [e.g., *Alexander et al.*, 2002]. For the WWR of SSTA in the North Pacific, *Alexander et al.* [2001] and *Zhao and Li* [2010] have pointed out that ENSO events are not essential for SSTA WWR in the North Pacific, although SSTA in the tropical Pacific associated with ENSO affect the wintertime SST in the North Pacific via changes in the extratropical atmospheric circulation. Here we need to further explore the relationship between the atmospheric WWR in the North Pacific and the ENSO variability in the tropics. A linear regression analysis is performed here. The geopotential height anomalies in the CNP region associated with the ENSO cycle, formed by regressing the geopotential height anomalies upon the Nino-3 SSTA time series, do not exhibit any WWR phenomena (not shown). Whereas the residual geopotential height anomalies, obtained by subtracting the regression value from the original data, exhibit the recurrence (Figure 2), and the spatiotemporal distribution of

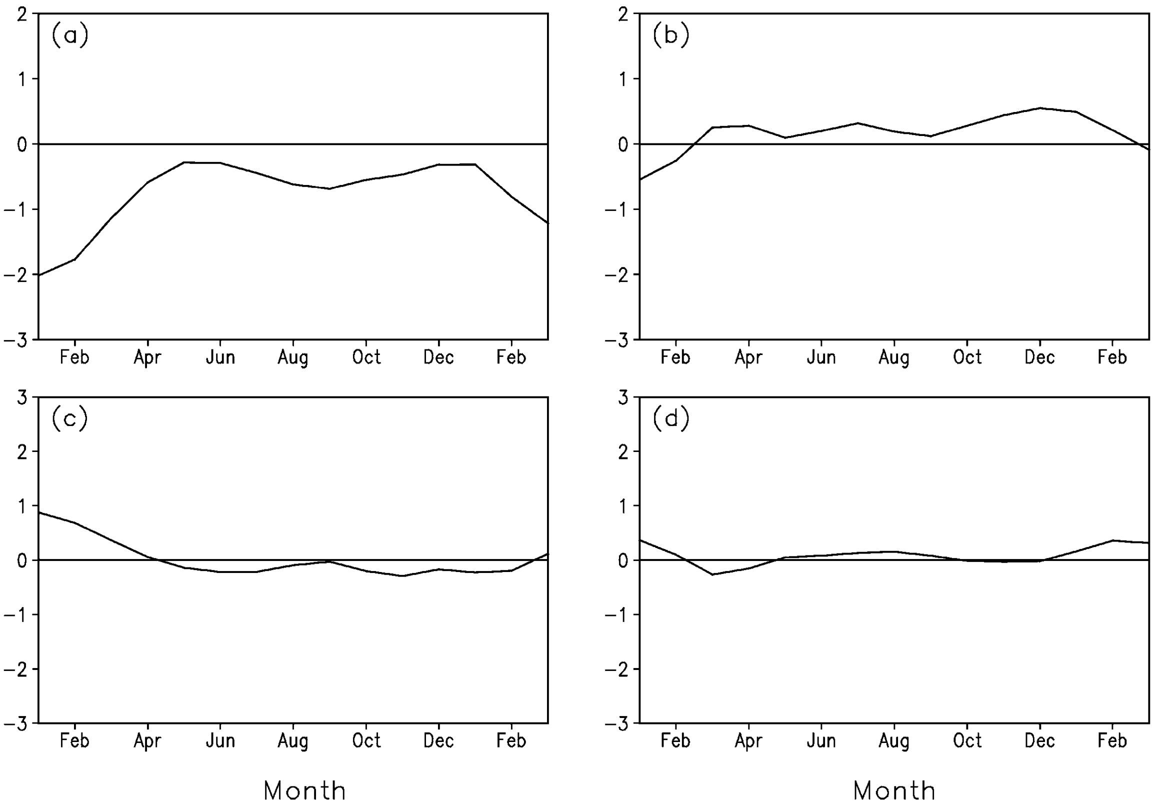

Figure 6. Evolution of the AO index during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

the atmospheric WWR is similar to that in the original data (Figure 1). This indicates that ENSO cannot lead to the atmospheric WWR in the CNP, because the WWR still exists after ENSO signal is subtracted. In fact, this result is due to the persistence characteristics of the ENSO itself. The lag correlation of the Nino-3 SSTA shows a spring persistence barrier, but without any recurrence regardless of the starting month (not shown).

3.2. Sea Ice and Snow Cover

[12] Sea ice and snow cover interact with climate. Whether the atmospheric WWR in the North Pacific is possibly caused by the winter-to-winter persistence of sea ice and snow cover anomalies? First, we need to objectively and effectively detect the WWR and its spatiotemporal distribution of sea ice anomalies. The method has been explained in detail in *Zhao and Li* [2010]. Previous studies used lag correlation analysis to define the WWR [e.g., *Alexander et al.*, 1999], and they primarily focused on analyzing dominant patterns of SSTA based on the leading empirical orthogonal function or on the regions designated subjectively. Here lag correlation analysis is made directly at each grid point, which avoids the dependence of the recurrence areas on specific spatial patterns or prior selection of regions. To efficiently detect the recurrence regions, we adopt the following criteria at each grid point: (1) lag correlation coefficient drops to an insignificant level prior to reaching a maximum and (2) the WWR is considered as a tendency for sea ice anomalies to recur from winter to the following winter without persisting through the intervening summer. WWR exists at one grid point if lag correlation meets the two criteria, and non-WWR if not.

[13] As shown in Figure 3, the sea ice anomalies exist WWR in high latitudes of the Northern Hemisphere. While the linear regression analysis, using sea ice anomalies averaged over the region (70°N–90°N, 0°W–360°W), indicates that the "ice-related" geopotential height anomalies do not appear WWR (not shown), but the 'residual field' which is linearly independent of the sea ice anomalies exhibits similar features of the WWR compared with the original data (Figure 4). It seems that sea ice anomalies in the high latitudes of the Northern Hemisphere do not influence quick transition of geopotential height anomalies in the CNP around March and the maintenance of the opposite sign height anomalies for two to three seasons.

[14] The snow cover (including Northern Hemisphere, Eurasia and North America) do not display significant WWR, as shown in Figure 5. And because the data lengths of snow cover are shorter, the effects of the snow cover anomalies on the WWR of the atmosphere do not be discussed here.

3.3. AO

[15] Effects of external forcing on the atmospheric WWR are not significant. Does the dominant source of the atmospheric WWR come from internal atmospheric dynamics? The AO is the dominant pattern of nonseasonal SLP

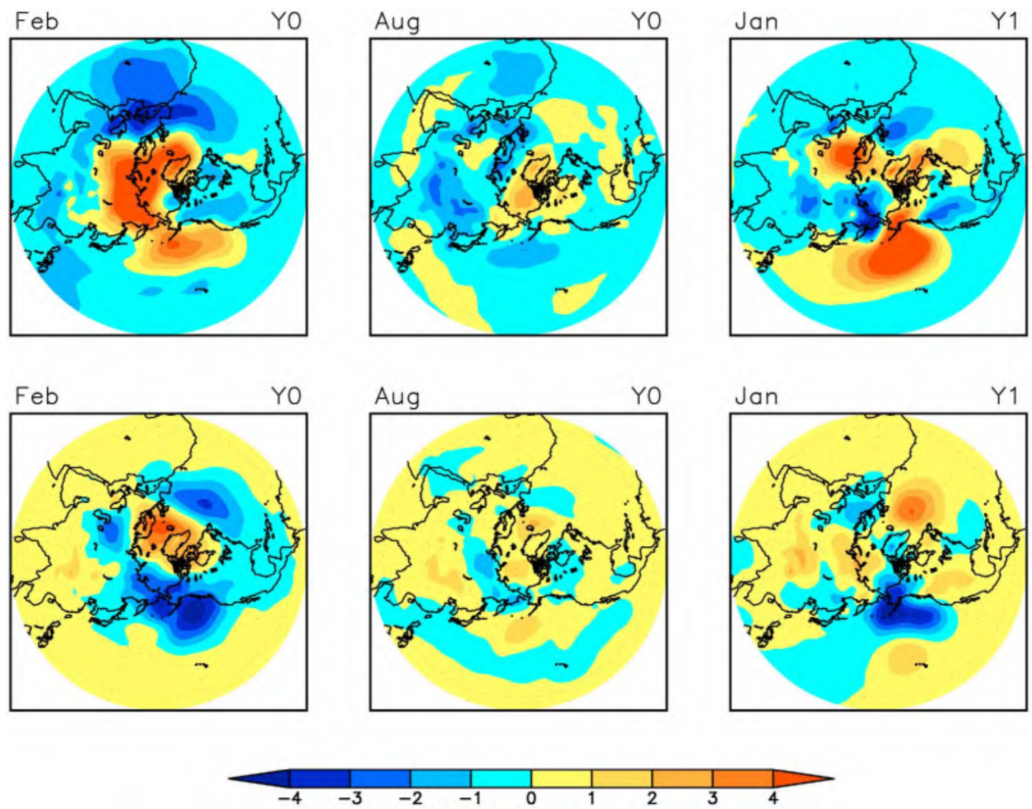

Figure 7. Evolution of the composite SLP anomalies in the Northern Hemisphere during (top) the positive WWR years and (bottom) the negative WWR years. Y0 indicates the current year, and Y1 indicates the next year.

variations north of 20°N, and it is characterized by SLP anomalies of one sign in the Arctic and anomalies of opposite sign across the subtropical and midlatitudes [e.g., *Thompson and Wallace*, 1998; *Li and Wang*, 2003]. Does the WWR of atmospheric circulation anomalies in the CNP represent seasonal variability of the AO?

[16] Figure 6 shows the evolution of the AO index during the WWR and non-WWR years. The monthly AO index is obtained from http://web.lasg.ac.cn/staff/ljp/data-NAM-SAM-NAO/NAM-AO.htm and is defined as the difference in the normalized monthly zonal mean SLP between 35°N and 65°N [*Li and Wang*, 2003]. The AO is at the negative phase during the positive WWR years (Figure 6a). The AO is at the positive phase during the negative WWR years, but it is much weaker than that during the positive WWR years (Figure 6b). The seasonal evolution of the AO index does not have reversals in the sign during spring and fall. It seems that AO does not display significant WWR characteristic during the WWR years of the CNP region. During the non-WWR years, AO index is much weaker than that during the WWR years throughout the year, especially there is a very small seasonal variability of AO index (Figures 6c and 6d). These results suggest that the WWR of atmospheric circulation anomalies in the CNP does not represent seasonal variability of the AO.

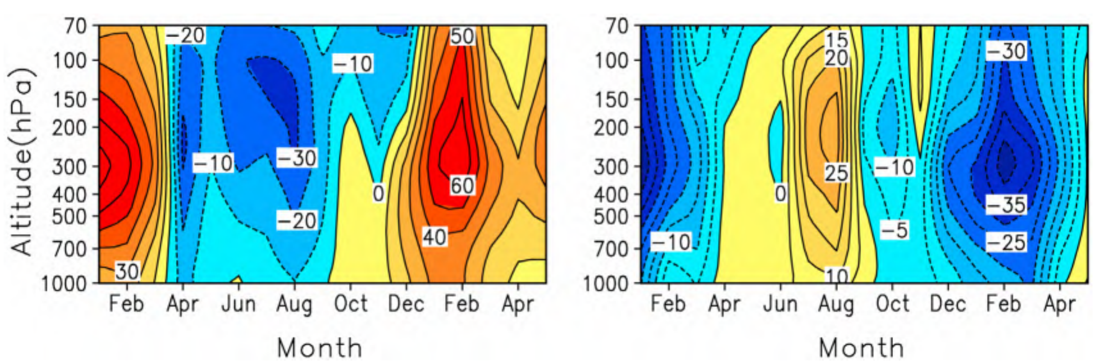

Figure 8. As in Figures 1a and 1b, but the AO-related anomalies are subtracted from the original data.

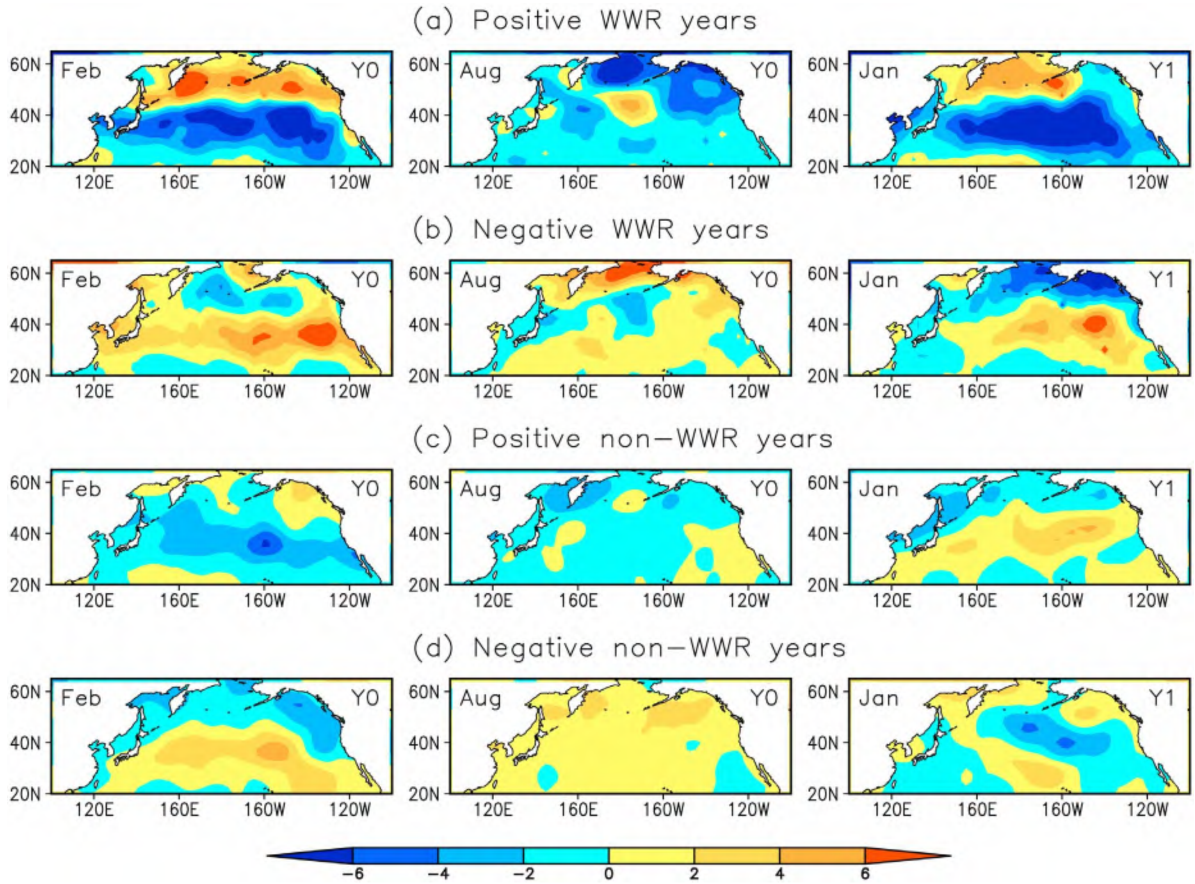

Figure 9. Evolution of root-mean-square (RMS) anomalies of daily geopotential height at 300 hPa in the North Pacific with a 2.5–6 day band-pass filter during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

[17] Figure 7 shows the evolution of the composite SLP anomalies in the Northern Hemisphere during the positive WWR years (Figure 7, top) and the negative WWR years (Figure 7, bottom). In the positive WWR years, the pattern of SLP anomalies during the first winter is similar with AO at the negative phase. There are high-pressure anomalies throughout the polar region, low-pressure anomalies across the subtropical and midlatitudes. But the SLP anomalies are very weak in summer. Although the SLP anomalies strengthen in the second winter, the pattern is different from that in the first winter. In the negative WWR years, the pattern of SLP anomalies is not similar with the AO at the positive phase, because there are high-pressure anomalies in the polar region. Thus, the pattern of SLP anomalies north of 20°N does not displays significant WWR characteristic during the WWR years of the CNP region. The atmospheric WWR in the CNP region is a local phenomenon.

[18] Subtracting the AO signal, the geopotential height anomalies still exhibit the recurrence, and the spatiotemporal distribution of the WWR is similar to that in the original data (Figure 8). This result suggests that the effect of the local internal atmospheric dynamics on the atmospheric WWR may be more important in the CNP region.

3.4. Variability in Storm Track Anomalies and Their Effects on the Mean Flow

[19] We next focus on the role of midlatitude synoptic transient eddies in the evolution of the mean atmospheric anomalies in the North Pacific. By mapping the root-mean-square (rms) statistics based on time-filtered data which retain periods within the 2.5–6 day band, it was reported that the most active disturbances tend to travel eastward through continuous phase propagation within two elongated zones spanning the North Pacific and North Atlantic, which was called storm tracks [e.g., *Blackmon et al.*, 1977; *Hoskins and Hodges*, 2002]. The systematic changes in the storm tracks and associated synoptic transient eddy allow the possibility that the heat and momentum or vorticity fluxes may help to reinforce and maintain the anomalous mean circulation [e.g., *Hoskins et al.*, 1983]. The CNP region is in the location of storm track in the North Pacific, so it is interesting to investigate changes in storm tracks during the WWR years and their link to the WWR of the atmospheric circulation anomalies.

[20] The eddy statistics used for this part come from NCEP daily analyses filtered to retain fluctuations between 2.5 and 6 days using the band-pass filter. Note that the climatological seasonal cycle has been removed from the daily values prior

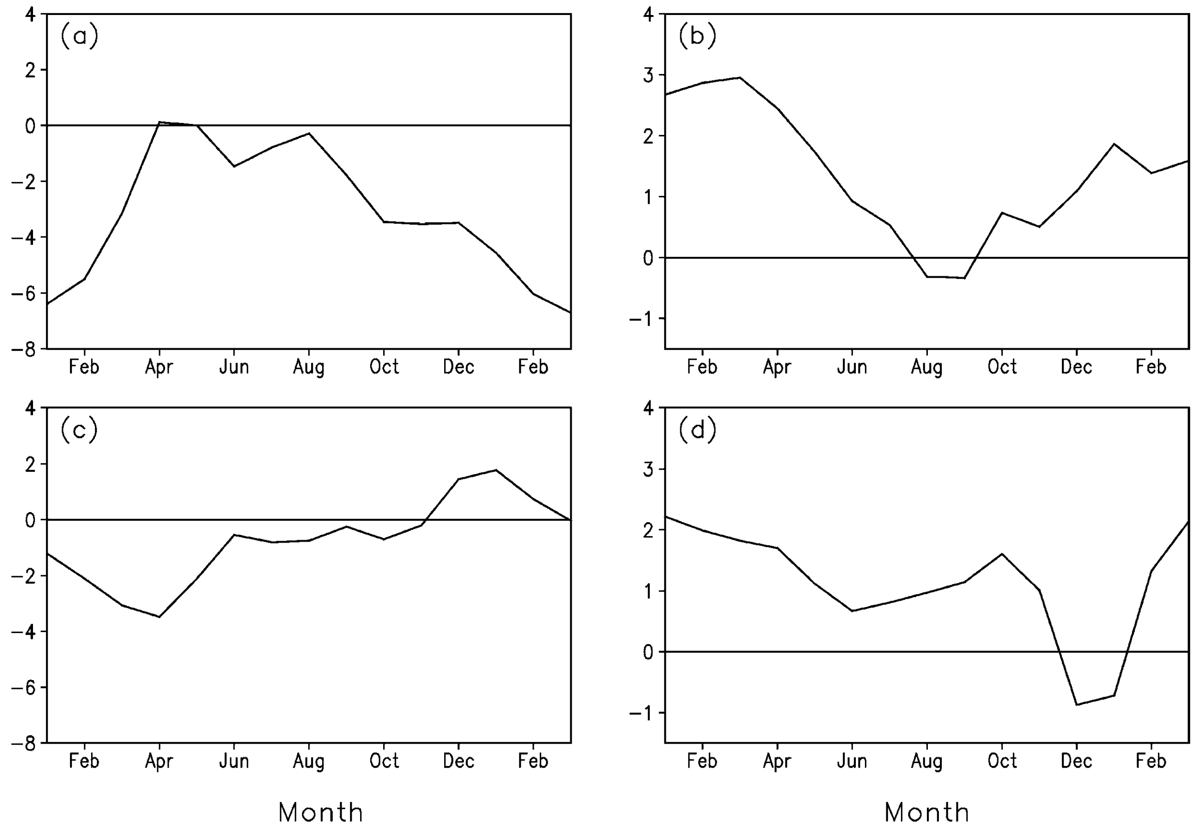

Figure 10. Evolution of RMS anomalies of daily geopotential height at 300 hPa in the CNP region with a 2.5–6 day band-pass filter during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

to the calculations. Figure 9 shows the evolution of RMS anomalies of daily geopotential height at 300 hPa with a 2.5–6 day band-pass filter in the North Pacific during the WWR and non-WWR years, which represents the anomalies of intensity of storm tracks and associated synoptic transient eddy. During the WWR years (Figures 9a and 9b), the storm track anomalies in two winters present a dipole structure in the ocean basin, which is different from the summer structure. In winter, the dipole-like structure of the storm track anomalies is similar to the second storm track mode in the North Pacific described by *Lau and Nath* [1991]. They pointed out that this mode depicts northward or southward migration of the storm tracks from their time mean positions, and this mode exhibits the strongest relationship with the monthly averaged circulation. During the non-WWR years (Figures 9c and 9d), the storm tracks anomalies shows a reversal in sign from the first to second winter, and the anomalies are weaker than those during the WWR years.

[21] Figure 10 shows the evolution of the RMS anomalies of daily geopotential height at 300 hPa in the CNP region. For the positive (negative) WWR years, there are reduced (enhanced) eddy activity anomalies across winter and enhanced (reduced) activity anomalies during summer (Figures 10a and 10b), coinciding with the seasonal evolution of the atmospheric circulation anomalies (Figures 1a and 1b). Diminished (enhanced) storminess is consistent with large-scale slackening (strengthening) of the westerlies. But the evolution during non-WWR years differs markedly from that during the WWR years. During the non-WWR years, the variability of storm track anomalies shows a reversal in sign from the first to second winter (Figures 10c and 10d).

[22] Changes in the storm tracks and associated synoptic eddy activity in the North Pacific help to reinforce and maintain the anomalous circulation in the upper troposphere [e.g., *Hoskins et al.*, 1983]. It is clear that the change in storm tracks plays a significant role in shaping the anomalous mean pattern. Moreover, the model simulations of *Held et al.* [1989], *Lau and Nath* [1990], and *Kushnir and Lau* [1992] indicated that the net effect of the transient eddies is to provide a strong positive feedback in the Pacific. In *Hoskins et al.* [1983], the scalar momentum equations are

$$D\bar{u} = f\bar{v}_{am} + \nabla \cdot E \quad (1)$$

$$D\bar{v} = -f\bar{u}_{am} - \left(\overline{u'v'}\right)_x \quad (2)$$

where the bar signifies a time average and the prime a deviation for the average, $\bar{v}_{am}$ and $\bar{u}_{am}$ is the residual circulation, and $E = \left(\overline{v'^2 - u'^2}, -\overline{u'v'}\right)$, $D = \bar{u}\frac{\partial}{\partial x} + \bar{v}\frac{\partial}{\partial y}$. The

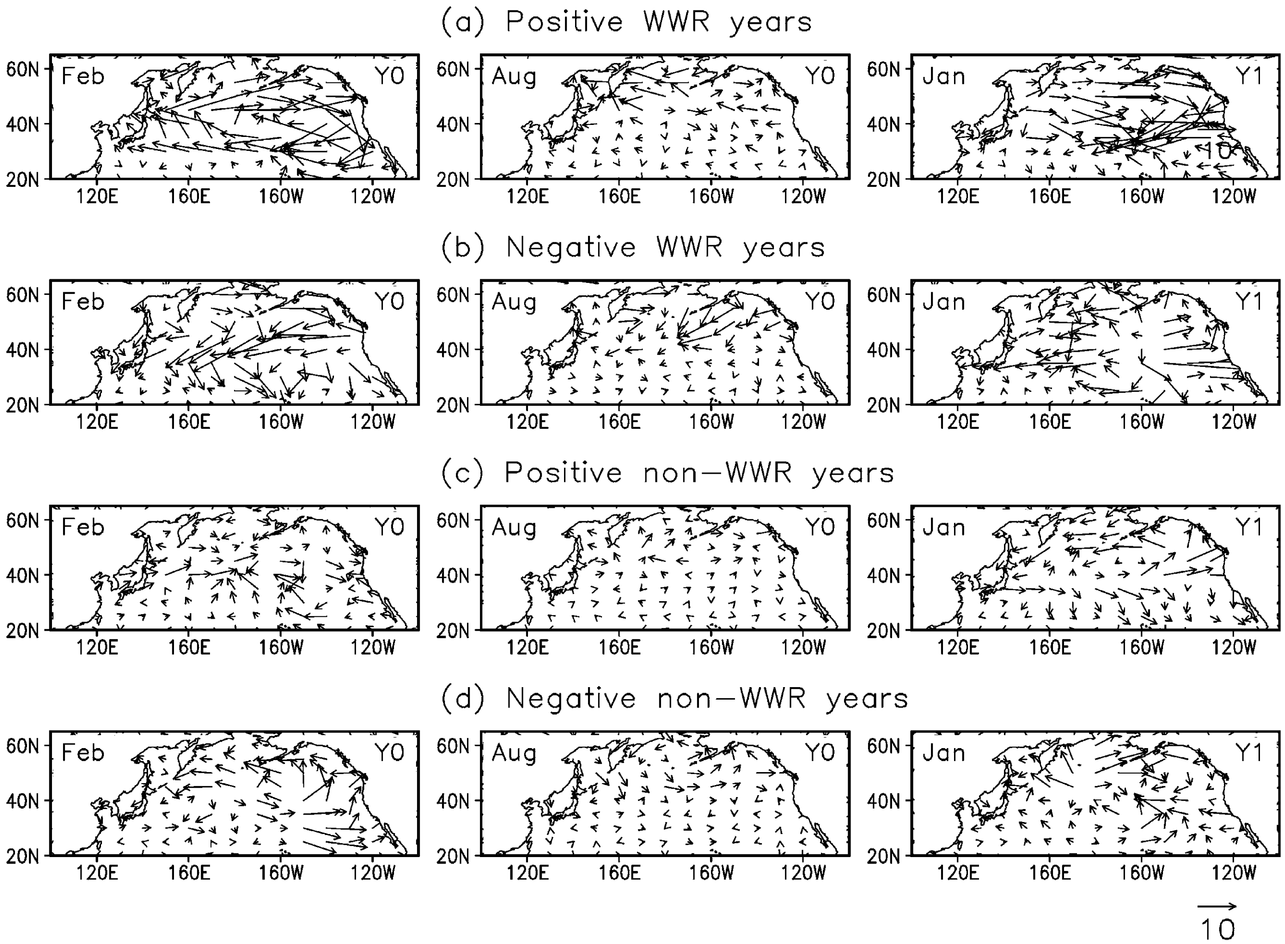

Figure 11. Evolution of $E = \left(\overline{v'^2 - u'^2}, -\overline{u'v'}\right)$ at 300 hPa in the North Pacific during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

eddy forcing term $\left(\overline{u'v'}\right)_x$ in (2) is generally very small so that the mechanical forcing by the eddies is accurately represented by $\nabla \cdot E$ in the x momentum equation given in (1). Where E is divergent there is a forcing of mean horizontal circulation consistent with a tendency to increase westerly mean flow. Where E is convergent, the mean flow circulation forcing is consistent with a tendency to decrease westerly mean flow.

[23] To pursue the mechanism for that the seasonal variability in storm tracks anomalies causes the WWR of atmospheric circulation anomalies, we compute the Eliassen-Palm (E-P) flux $E = \left(\overline{v'^2 - u'^2}, -\overline{u'v'}\right)$ following the formulation of *Hoskins et al.* [1983]. Figure 11 shows the evolution of the E in the North Pacific during the WWR and non-WWR years. During the positive (negative) WWR years, the *E is* convergent (divergent) in the North Pacific in two winters but very weak in the intervening summer (Figures 11a and 11b). Where E is divergent there is a forcing of mean horizontal circulation consistent with a tendency to increase westerly mean flow. Where E is convergent, the mean flow circulation forcing is consistent with a tendency to decrease westerly mean flow [*Hoskins et al.*, 1983]. But this characteristic is not found during the non-WWR years (Figures 11c and 11d).

[24] A more accurate assessment of impact on the mean flow at 300 hPa in the CNP region is given in Figure 12, which shows the evolution of the $\nabla \cdot E$ in the CNP during the WWR and non-WWR years. The seasonal evolution of the $\nabla \cdot E$ coincides with that of the anomalous mean atmospheric circulation (Figure 1). Thus, transient eddy forcing could cause the WWR of atmospheric circulation anomalies through the barotropic effects of eddy momentum fluxes. Seasonal variability in storm track anomalies and associated synoptic eddy may be one of the causes for the WWR of atmospheric circulation, which lead to the quick transition of height anomalies around March and the maintenance of the opposite sign height anomalies for two to three seasons. More important, seasonal change in storm track and $\nabla \cdot E$ is earlier than that of the oceanic temperature anomalies (bottom panels of Figure 3 in *Zhao and Li* [2012]), which is strong evidence that the atmosphere forces the ocean in the North Pacific. The $\nabla \cdot E$ is coherent in the vertical but strongest at 300 hPa near the tropopause. Since the intensity of momentum fluxes in the troposphere increase with height, the $\nabla \cdot E$

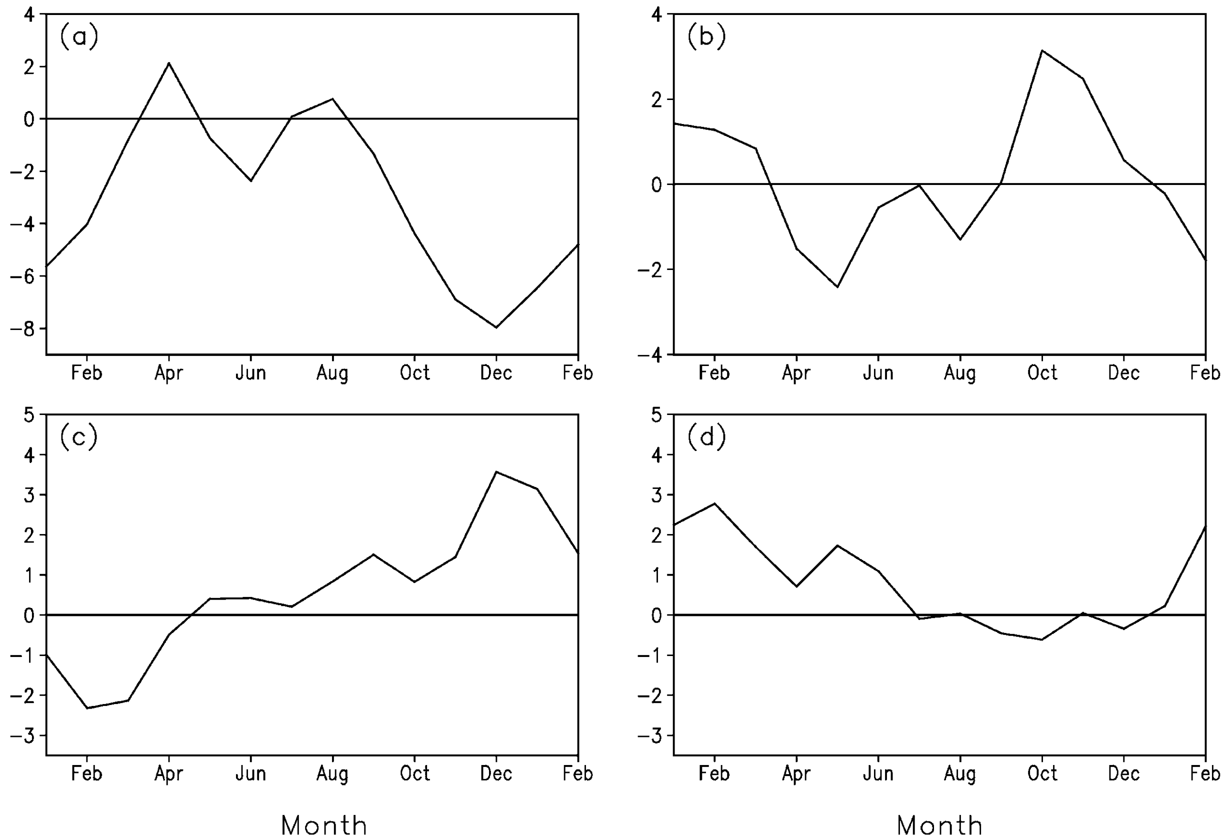

Figure 12. Evolution of the $\nabla \cdot E$ at 300 hPa in the CNP region during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

in the upper troposphere are stronger than those at the lower troposphere (Figure 13).

4. Summary and Discussion

[25] Previous studies on the WWR in the North Pacific only focused on physical processes contributing to the SSTA WWR. *Zhao and Li* [2010, 2012] indicated that the WWR also exists in the atmospheric circulation anomalies and it is one of the causes of the SSTA WWR in the CNP, in addition to the oceanic reemergence mechanism [*Alexander and Deser*, 1995]. The present study investigated the causes of atmospheric WWR in the CNP. The results indicate that tropical ENSO is not essential for the atmospheric WWR to occur in the North Pacific, because the persistence characteristic of the ENSO itself does not show any recurrence regardless of the starting month. Effect of other external forcing, e.g., sea ice, is also not significant. These results suggest that the dominant source of the atmospheric WWR may come from internal atmospheric dynamics in the North Pacific.

[26] The AO is the dominant pattern of nonseasonal SLP variations north of 20°N. The AO index is at the negative phase during the positive WWR years, which is strong in winter and weak in summer; the AO index is at the positive phase during the negative WWR years, but it is much weaker than that during the positive WWR years. In addition, the spatial pattern of the SLP anomalies during the WWR years is not very similar with the AO, especially for the negative WWR years. Subtracting the AO signal, the recurrence still exhibits in the CNP. It seems that the WWR of atmospheric circulation anomalies in the CNP does not represent seasonal variability of large-scale pattern of the atmospheric circulation anomalies in the Northern Hemisphere. Effect of the local internal atmospheric dynamics on the atmospheric WWR may be more important in the CNP region.

[27] In the North Pacific, the systematic changes in the storm tracks and associated synoptic transient eddy allow the possibility that the heat and momentum or vorticity fluxes may help to reinforce and maintain the anomalous mean circulation [e.g., *Hoskins et al.*, 1983]. The CNP region is in the location of the storm track in the North Pacific. For the positive (negative) WWR years, there are reduced (enhanced) activity anomalies of storm tracks and associated synoptic transient eddy across winter and enhanced (reduced) activity anomalies during summer, coinciding with the seasonal evolution of the atmospheric circulation anomalies.

[28] Seasonal variability in storm track anomalies and associated synoptic eddy dynamics may be one of the causes for the WWR of atmospheric circulation in the CNP, which leads to quick transition of height anomalies around March and the maintenance of the opposite sign atmospheric circulation anomalies for two to three seasons. First, the barotropic forcing of momentum transports on the mean flow also

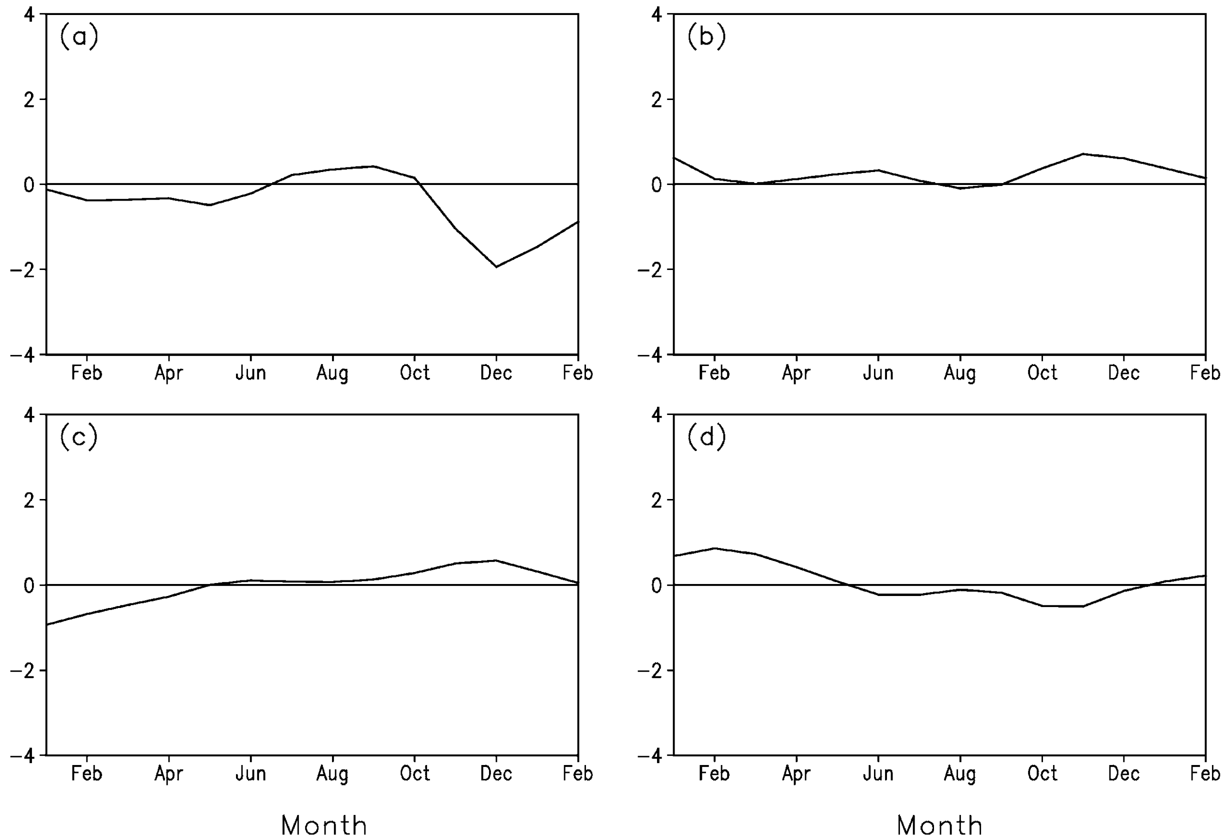

Figure 13. As in Figure 12, but for 850 hPa.

displays the notable WWR characteristic during the WWR years, but it does not recur during the non-WWR years. During the positive (negative) WWR years, the E at 300 hPa is convergent (divergent) during two winters but very weak in the intervening summer. Where E is convergent (divergent) there is a forcing of mean horizontal circulation consistent with a tendency to decrease (increase) westerly mean flow [*Hoskins et al.*, 1983]. Thus, transient eddy forcing could induce the WWR of atmospheric circulation anomalies through the barotropic processes. In addition to the barotropic transient eddy forcing, the effect of the baroclinic forcing of eddy heat transports on the mean flow is also important. The heat flux by transient eddies is a maximum at lower tropospheric levels [e.g., *Hoskins et al.*, 1983], so that a quite complete picture of mean flow forcing may be obtained by plotting low-level poleward heat flux. As shown in Figure 14, the evolution of the $\overline{v'T'}$ anomalies at 850 hPa in the North Pacific also displays notable WWR characteristic during the WWR years, but this characteristic does not exist during the non-WWR years. Thus, both the momentum and heat fluxes by transient eddies act to force the WWR of atmospheric circulation anomalies in the CNP region.

[29] Unlike the reemergence mechanism of the SSTA WWR, atmospheric anomalies in summer do not appear to be a link between those in the preceding and following winters, because the geopotential height field displays WWR from the lower layer to the upper layer (Figures 1a and 1b), which shows an equivalent barotropic vertical structure and the centers of anomalies are located in the high troposphere (500–300 hPa). Winter atmospheric anomalies do not persist at a certain layer of atmosphere through the intervening summer. It seems that the recurrent atmospheric circulation anomalies in the second winter do not come from those of the previous winter through anomalies in the intervening summer. Thus, mechanisms of the WWR are markedly different between the atmosphere and ocean. While persistence is the key mechanism for SSTA WWR, it is not true for the WWR of atmospheric circulation anomalies. In addition, the mechanism determining the seasonal evolution the storm tracks anomalies and associated transient eddy forcing still remains an open question. It has often been stated that the atmosphere has a very short memory. The limited memory of the atmosphere suggests that the observed WWR in the storm track anomalies may be a response to forcing by some other part of the climate system. However, it is not the midlatitude ocean because the atmosphere forces the ocean primarily [*Zhao and Li*, 2010], and it is not the tropical ocean (Figure 2) and sea ice in the high latitudes of the Northern Hemisphere (Figure 4). Thus, it may be a manifestation of natural climate variability, which needs further work to be done.

[30] As is well known, the extratropical interaction between ocean and atmosphere is not one way. For the atmosphere, what extent the oceanic reemergence mechanism affects the overlying atmosphere including the

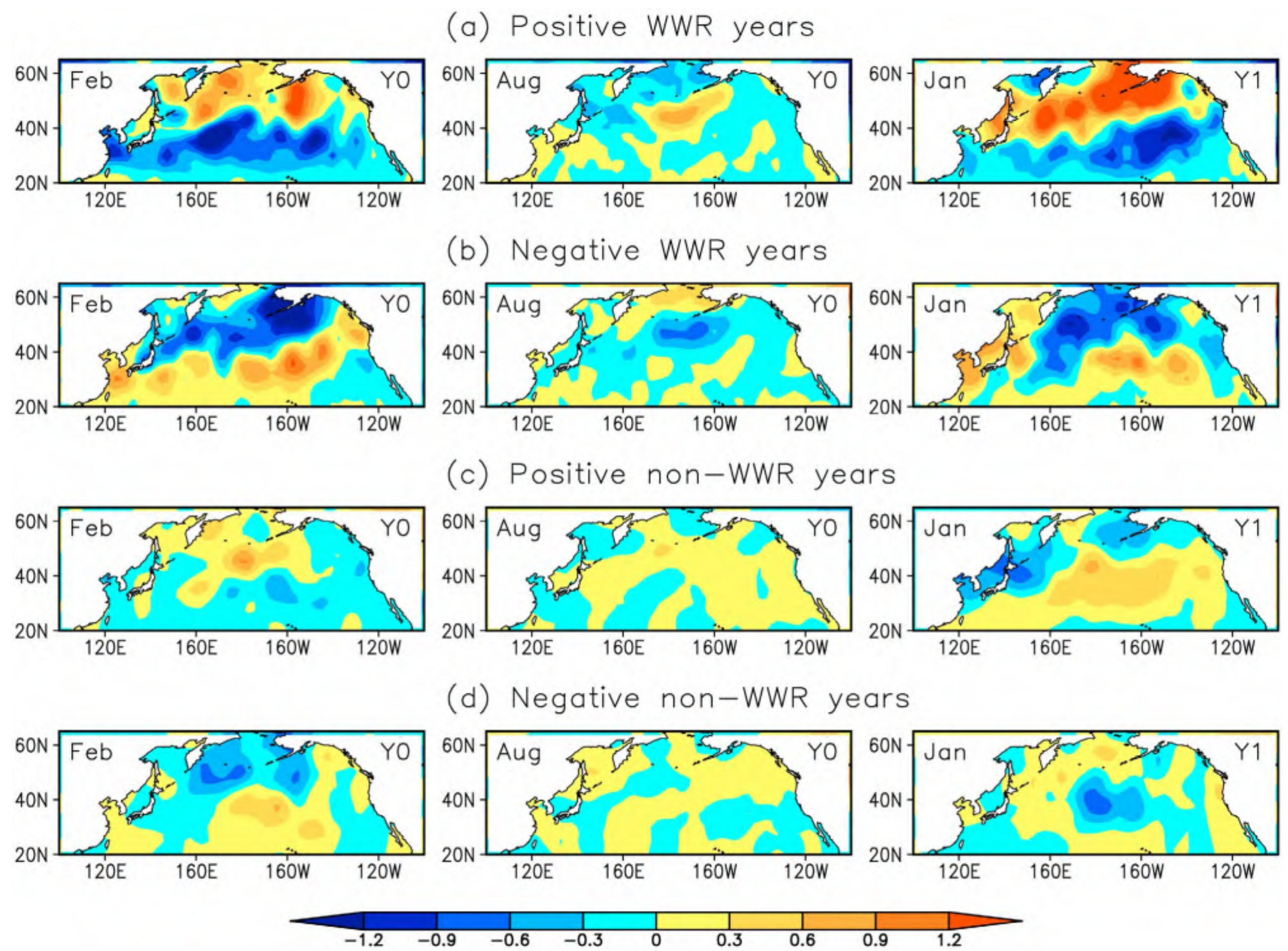

Figure 14. Evolution of the $\overline{v'T'}$ anomalies at 850 hPa in the North Pacific during (a) the positive WWR years, (b) the negative WWR years, (c) the positive non-WWR years, and (d) the negative non-WWR years.

atmospheric WWR? Isolating and quantifying the role of oceanic reemergence upon the atmospheric circulation is difficult in observations because of the dominance of atmospheric forcing of the underlying ocean. Results of model experiments may give us some insight. *Bhatt et al.* [1998], *Cassou et al.* [2007], and *Liu et al.* [2007] investigated the atmospheric responses to the recurrence of the SSTA in the North Atlantic and North Pacific, respectively. It seems that interaction of the WWR between the ocean and the atmosphere exists. It does not mean that the SSTA WWR would be the origin of the WWR in the atmospheric circulation. There is strong evidence that at midlatitude the atmosphere forces the ocean, especially in winter. As claimed by *Cassou et al.* [2007], while addressing the question as to what extent the oceanic reemergence mechanism affects the overlying atmosphere, we have to keep in mind that the dominant source of the NAO variability is internal atmospheric dynamics. *Bhatt et al.* [1998] pointed out that the primary mode in the North Atlantic is reproduced in both the control and coupled integrations; thus, coupling does not quantitatively change the structure of the patterns of atmospheric variability over the midlatitude North Atlantic, and the feedback of the ocean on the atmosphere is subtle. *Liu et al.* [2007] also indicated that the monthly atmospheric response to SST is much weaker than the atmospheric internal variability in the midlatitude.

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