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On the Weakened Relationship between Spring Arctic Oscillation and Following Summer Tropical Cyclone Frequency over the Western North Pacific: A Comparison between 1968–1986 and 1989–2007

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ABSTRACT

This study documents a weakening of the relationship between the spring Arctic Oscillation (AO) and the following summer tropical cyclone (TC) formation frequency over the eastern part (150°–180°E) of the western North Pacific (WNP). The relationship is strong and statistically significant during 1968–1986, but becomes weak during 1989–2007. The spring AO-related SST, atmospheric dynamic, and thermodynamic conditions are compared between the two epochs to understand the possible reasons for the change in the relationship. Results indicate that the spring AO leads to an El Niño-like SST anomaly, lower-level anomalous cyclonic circulation, upper-level anomalous anticyclonic circulation, enhanced ascending motion, and a positive midlevel relative humidity anomaly in the tropical western–central Pacific during 1968–1986, whereas the AO-related anomalies in the above quantities are weak during 1989–2007. Hence, the large-scale dynamic and thermodynamic anomalies are more favorable for TC formation over the eastern WNP during 1968–1986 than during 1989–2007.

Key words: spring Arctic Oscillation, summer tropical cyclone, western North Pacific, SST

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1. Introduction

The tropical cyclone (TC) is one of the most severe natural disasters in the world. TCs can exert substantial influences on the economy, industry, agriculture, and people's daily lives (Emanuel, 2005). Thus, a better understanding of TC variability and its factors, and improved predictability of TC activity, are of great importance to society. It has been demonstrated that the TC variability over the western North Pacific (WNP) is significantly influenced by SST anomalies over the tropical central–eastern Pacific (e.g., Chan, 1985; Chen et al., 1998; Wang and Chan, 2002; Chan, 2005; Chen and Huang, 2008; Yeh et al., 2010; Cao et al., 2014a; Hsu et al., 2014) and over the Indian Ocean (Zhan et al., 2011). Studies have also indicated that atmospheric circulation systems can influence TC activity over the WNP, such as the stratospheric quasi-biennial oscillation (Chan, 1995; Ho et al., 2009), the Antarctic Oscillation (Ho et al., 2005), and the North Pacific Oscillation (Wang et al., 2007).

The Arctic Oscillation (AO) is the leading mode of climate variability over the extratropics of the Northern Hemisphere (NH) and it features an opposite fluctuation of sea level pressure (SLP) between the high and middle latitudes (Thompson and Wallace, 1998, 2000). Previous studies have demonstrated that variability of the AO can exert substantial influences on the weather and climate anomalies over the globe (e.g., Gong et al., 2001; Gong and Ho, 2002, 2003; Chen and Li, 2007; Chen et al., 2013a, 2013b). Choi and Byun (2010) found that the boreal summer AO can significantly influence the interannual variability of TC activity over the WNP in the concurrent summer. They indicated that more TCs form to the east of 150°E, recurve in the eastern part of the WNP, and move to the midlatitudes in the summer during low AO years than during high AO years. Choi et al. (2012) further showed that the frequency of summer TCs over East Asia has a positive correlation with the preceding spring AO. They demonstrated that the spring AO influences following summer TC formation over East Asia via modulation of atmospheric circulation.

As mentioned above, the AO can impact WNP TC activity in the simultaneous summer, and the preceding spring AO can modulate the frequency of TC activity over East Asia in the summer. However, it is unclear whether the spring AO can exert influences on TC activity over the WNP in the following summer. In addition, Chen et al. (2015) and Gao et al. (2014) found that the connection between the spring AO and atmospheric circulation anomalies over the North Pacific is

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not stationary. Hence, another interesting question is whether the relationship between spring AO and TC activity over the WNP in the following summer has changed in past decades. This study attempts to address the above two issues.

The rest of the paper is organized as follows. The data and analysis methods are described in section 2. In section 3, observational evidence is presented for the spring AO–summer WNP TC relationship and interdecadal change in the relationship. Section 4 compares the spring AO-associated SST, atmospheric dynamic, and thermodynamic anomalies before and after the interdecadal change. Section 5 gives a summary and a brief discussion.

2. Data and methodology

In the present study, monthly mean SST was obtained from the National Oceanic and Atmospheric Administration Extended Reconstructed SST dataset, version 3b (ERSSTv3b), which has a $2^{\circ} \times 2^{\circ}$ horizontal resolution from 1854 to the present day (Smith et al., 2008). This study employed monthly mean horizontal winds at 850 hPa and 200 hPa, vertical velocity at 500 hPa, relative humidity at 600 hPa, and SLP from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al., 1996). The NCEP–NCAR dataset has a horizontal resolution of 2.5° in latitude and longitude from 1948 to the present day. The time and position of TC genesis over the WNP were obtained from the National Climate Data Center's International Best Track Archive for Climate Stewardship (IBTrACS), version 2 (Knapp et al., 2010). This dataset merged TC datasets from different operational centers around the world. The best-track TC data included 6-hourly longitude and latitude of TC center and maximum sustained wind speed.

The period for all variables spanned from 1958 to 2008 during the June–August season. The climatological mean (1958–2008) for each calendar month was removed from the monthly mean to obtain monthly mean anomalies. The AO index was defined as the principle component corresponding to the leading EOF mode of SLP anomalies over the extratropics (20°–90°N) of the NH (Thompson and Wallace, 1998). TC genesis index was defined as the first record of TC best-track data over the WNP. Note that the TC genesis is rejected if the wind records are missing in all 6-hourly time. Previous studies have suggested that TC genesis frequency in the WNP has experienced a significant interdecadal variability (e.g., Yumoto and Matsuura, 2001; Chu, 2002; Yeh et al., 2010). In order to avoid the interference of the low frequency variability, the present analysis focused on the interannual variability of TC activity associated with the AO. Thus, the AO index, TC index, and monthly mean variables were subjected to a 7-year high-pass Lanczos filter (Duchon, 1979).

In this study, the statistical significance of the correlation coefficients is determined by using the two-tailed t-test and the effective degrees of freedom (EDOF) (Bretherton et al., 1999). The EDOF is calculated as follows:

$$M_{\mathrm{edof}} = N\frac{1-r_1 r_2}{1+r_1 r_2},$$

where N is the original sample size, and r_1 and r_2 are the lag one autocorrelations of the AO time series and summer TC formation index, respectively.

3. Observed interdecadal change in the influence of AO on TC formation

Previous studies have demonstrated that while the correlation between the total TC frequency number over the WNP and Niño3.4 index [area-averaged SST anomalies over the region (5°S–5°N, 170°E–120°W)] is not statistically significant, the correlation between the TC number over the WNP and Niño3.4 index varies significantly with the region (e.g., Wang and Chan, 2002; Chen and Huang, 2008). The frequency of TC formation is above normal in the southeastern part, but below normal in the northwestern part of the WNP during El Niño developing summers (Wang and Chan, 2002; Chen and Huang, 2008; Li, 2012). Wu et al. (2012) suggested that this is mainly associated with an eastward-extending monsoon trough, warmer SST, increased midlevel relative humidity, enhanced lower-level cyclonic vorticity, and reduced vertical zonal wind shear over the southeast quadrant of the WNP. Our analysis indicates that the connection between the spring AO and the following summer TC formation frequency over the WNP also depends strongly on the region. Thus, in order to avoid the offset of opposite signals over the WNP, the WNP is divided into two parts based on the longitude of 150°E: a western part (120°–150°E), and an eastern part (150°–180°E). The interannual anomalies of the spring AO index and the following summer TC genesis index in the western and eastern parts of the WNP are calculated using the NCEP–NCAR SLP and IBTrACS dataset, respectively.

Figure 1a displays the normalized time series of interannual anomalies of the spring AO index and the following summer TC genesis index in the eastern part of the WNP. The correlation coefficient between these two time series for the entire period (1958–2008) is not statistically significant (i.e., 0.06). However, the relationship between the two indices appears to change during 1958–2008, with more same-sign values during the 1970s and 1980s and opposite-sign values during the 1990s and 2000s (Fig. 1a). This indicates that the connection between the spring AO and the following summer TC formation frequency is not stable. We further calculate the sliding correlation between these two indices with a window of 21 years (Fig. 1b). Years shown in Fig. 1b are labeled according to the central year of the 21-year window. One can find high correlation coefficients above 0.5 during 1978–1979, which is statistically significant at the 5% confidence level according to the Student's t-test, while the positive correlation before 1973 and during 1980–1993 is not statistically

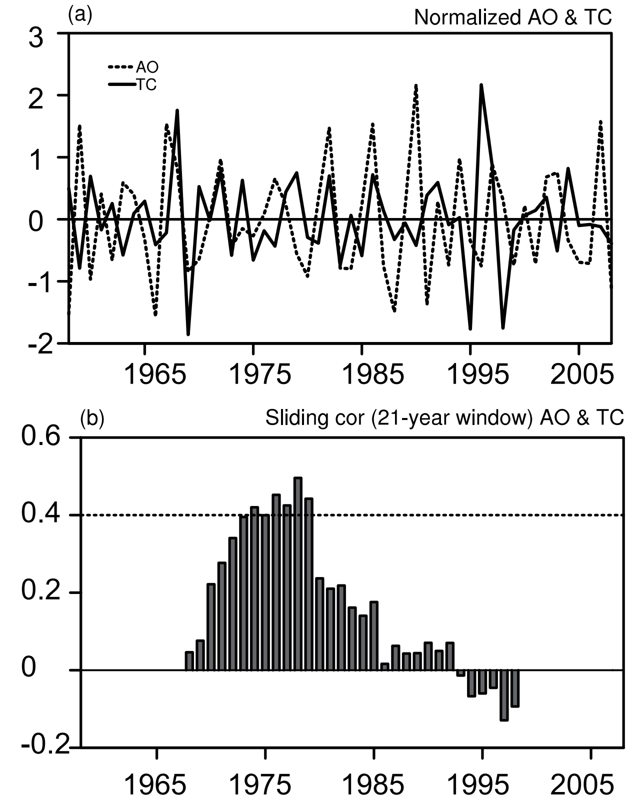

Fig. 1. (a) Normalized time series of interannual components of spring (MAM-averaged) AO index and the following summer (JJA-averaged) TC formation index in the eastern part of the WNP. (b) Sliding correlations between the interannual components of spring AO index and the following summer TC formation index displayed at the center year of a 21-year window. The horizontal black dashed line in (b) indicates the correlation coefficient is significantly different from zero at the 10% level.

significant (Fig. 1b). In addition, weak negative correlation is observed after 1993.

From Fig. 1b, the year 1978 represents the central year of the 1968–1988 window when the correlation is the highest (i.e., 0.51), and the year 1997 represents the central year of the 1987–2007 window when the correlation is the lowest (i.e., −0.12). Hence, the two periods of 1968–1986 and 1989–2007, containing 19 years, will be compared in the following analysis. Note that the two years 1987 and 1988 are not included. This is because the contrasts between the two 19-year periods are the most remarkable. Meanwhile, we also objectively detect the occurrence of the interdecadal decrease of the relationship between the spring AO and the following summer TC genesis frequency using the Bayesian change-point method developed by Chu and Zhao (2004) and Zhao and Chu (2010) (downloaded from http://www.soest.hawaii.edu/MET/Hsco/chu-index.html) (not shown). This shows that a posterior probability of a shift occurs in the mid-1980s. Thus, the two periods of 1968–1986 and 1989–2007 are basically reasonable. Hereafter, for convenience, the two selected periods are denoted as PRE (1968–1986) and POST (1989–2007) epochs in the subsequent analysis.

We have also examined the connection between the spring AO index and the following summer TC genesis index in the western part of the WNP and the sliding correlation between these two indices with a window of 21 years (figure not shown). Results show that the connection between these two indices during the entire time period (i.e., −0.05) and the sliding correlation are both insignificant. This indicates that the spring AO has no obvious influence on the TC activity over the western part of the WNP in the following summer.

4. Mechanisms for interdecadal change between the AO and TC formation

4.1. *Interdecadal change of spring AO-related anomalies in the following summer*

As stated in the previous section, periods exist when the relationship between the spring AO and the following summer TC formation over the eastern part of the WNP is significant, and there are others when this relationship is insignificant. In this section, we investigate why the connection between the spring AO and the following summer TC genesis in the eastern part of the WNP has changed. For this purpose, we compare the spring AO-related anomalies in SST, atmospheric dynamic, and thermodynamic variables between the PRE and POST epochs. Gong et al. (2011) showed that the spring AO can exert a significant influence on the following East Asian summer monsoon (EASM). They indicated that the tropical central–eastern Pacific SST anomalies play an important role in relaying the influence of spring AO on the following EASM. As demonstrated by Gong et al. (2011), during positive spring AO years, a significant cyclonic circulation anomaly can be observed in the tropical western Pacific in the simultaneous spring. Westerly anomalies to the south of the cyclonic circulation anomaly reduce climatological easterly winds and result in positive SST anomalies over the tropical central Pacific via reduction of upward latent heat flux. The spring AO-induced cyclonic circulation anomaly over the WNP sustains from spring into the following summer via coupling with the tropical central Pacific SST and subsequently influences the EASM via modulating the subtropical WNP high. It is speculated that Pacific SST anomalies may also be important in relaying the influence of spring AO on the following summer TC activity in the eastern part over the WNP.

Figure 2 shows the following summer SST anomalies regressed upon the normalized spring AO index in the PRE and POST epochs. During the PRE epoch, positive SST anomalies are seen in the eastern and northeastern Pacific with a southwestward extension to the equatorial central Pacific (Fig. 2a). Negative SST anomalies are present in the subtropical WNP between 10°N and 30°N. This SST anomaly pattern bears a close resemblance to that in the El Niño developing phase. This result is consistent with previous studies (Wang and Chan, 2002; Li, 2012) that showed more frequent TCs in the southeastern quadrant of the WNP during El Niño developing summers. In contrast, during the POST epoch,

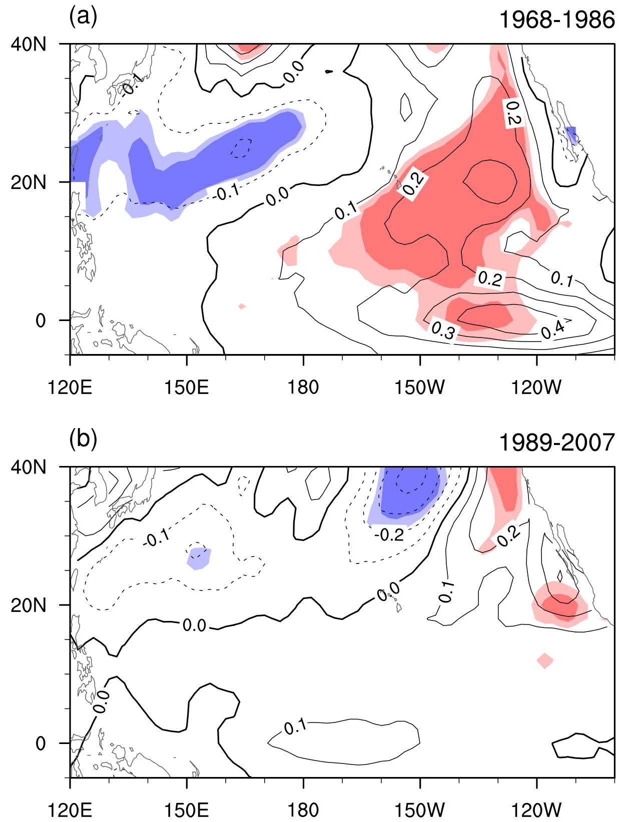

Fig. 2. Anomalies of the following summer (JJA) SST obtained as regression upon the normalized spring AO index in (a) 1968–1986 and (b) 1989–2007. The heavy and light shading indicates that the anomalies are significantly different from zero at the 5% and 10% levels, respectively. Contour interval: 0.1°C.

the spring AO-related SST anomalies in the following summer are fairly weak over the tropical Pacific, except for small patches over the subtropical Pacific between 20°N and 30°N (Fig. 2b). Hence, the interdecadal change in the influence of spring AO on the following summer SST around the late 1980s may be attributed to summer SST anomalies in the Pacific Ocean.

Further, we examine atmospheric circulation in response to the spring AO. Figure 3 displays anomalies of horizontal winds at 850 hPa and 200 hPa in the summer obtained by regression upon the normalized spring AO index for the two epochs. Notable differences appear in the spring AO-related horizontal wind anomalies at both the lower and upper levels between the two periods. During the PRE epoch, a remarkable lower-level cyclonic circulation anomaly is present in the center of the WNP, accompanied by significant westerly anomalies in the tropical Pacific spanning from 140°E to 180°E and significant easterly anomalies in the subtropical Pacific between 25°N and 35°N (Fig. 3a). Chen et al. (2014a) demonstrated that the interaction between synoptic-scale eddies and low frequency flow plays an important role in the formation of the spring AO-related westerly wind anomalies over the WNP. They indicated that spring AO variability influences the strength of the westerly jet stream over the East Asia–North Pacific midlatitudes. It was shown by Lau (1988) that the westerly jet stream change is accompanied by variation of synoptic-scale eddies (also called the storm track). Storm track anomalies, in turn, feed back to the low frequency flow. This explains the formation of the cyclonic circulation anomaly in the WNP and related westerly wind anomalies over the tropical western–central Pacific.

The El Niño-like SST pattern in Fig. 2a and westerly wind distribution in Fig. 3a are closely coupled. The westerly anomalies associated with the spring AO over the tropical western–central Pacific can influence the tropical central–eastern Pacific SST anomalies through triggering an eastward propagating and downwelling equatorial Kelvin wave (Wang et al., 1999; Chen et al., 2014a, 2014b). Meanwhile, westerly anomalies in the equatorial central Pacific could be strengthened by SST anomalies over the eastern Pacific during El Niño developing summers (Li, 2012). On the contrary, during the POST epoch, the cyclonic circulation anomaly over the tropical North Pacific is much weaker and shifts southward compared to that during the PRE epoch (Fig. 3b). It can thus be seen that extremely weak westerly anomalies are present in the tropical western–central Pacific, indicating that the spring AO does not exert a large influence on the following SST anomalies in the tropical eastern Pacific in the POST epoch (Fig. 3b and Fig. 2b).

On the other hand, at the upper level, a significant anticyclonic circulation anomaly is observed in the subtropical Pacific along 20°N, which is accompanied by significant easterly anomalies in the tropical Pacific along 140°–180°E and significantly westerly anomalies in the subtropical Pacific along 20°–30°N (Fig. 3c). The distributions of horizontal winds at the lower and upper levels during the PRE epoch are consistent with a previous observational study that showed both lower-level cyclonic vorticity and upper-level anticyclonic vorticity are favorable for TC formation (McBride and Zehr, 1981). In contrast, during the POST epoch the corresponding anticyclonic circulation anomaly at the upper level is insignificant over the central Pacific and shifts eastward compared to that in the PRE epoch (Fig. 3d).

Gray (1968) identified three dynamic and three thermodynamic necessary conditions for tropical cyclogenesis: low-level positive relative vorticity, weak vertical wind shear, non-zero Coriolis force, warm SST, high midlevel relative humidity, and a conditional unstable atmosphere. The dynamic and thermodynamic anomalies associated with TC formation are compared during the PRE and POST periods. Figure 4 shows the anomalies of summer SLP, vertical velocity at 500 hPa, and relative humidity at 600 hPa obtained by regression upon the normalized spring AO index in the PRE and POST periods, respectively. There are significant differences in these three variables between the two epochs. During the PRE epoch, significant negative SLP and 500-hPa vertical velocity anomalies are observed in the eastern part of the WNP, implying lower pressure at the surface and enhanced upward motion at the middle level (Figs. 4a and c). On the contrary, during the POST epoch, the SLP and vertical motion anomalies over the WNP are much weaker and insignificant com-

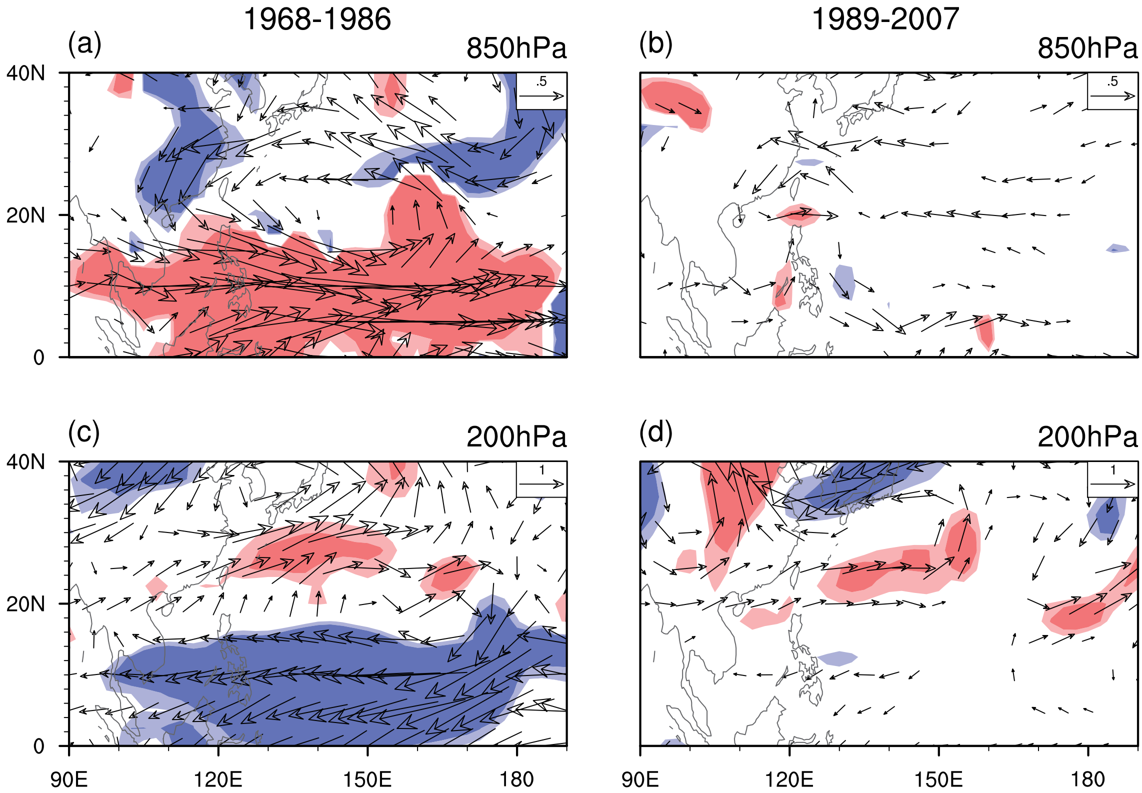

Fig. 3. Anomalies of the following summer (a, b) horizontal winds at 850 hPa and (c, d) at 200 hPa obtained as regression upon the normalized spring AO index in (a, c) 1968–1986 and (b, d) 1989–2007. The heavy and light shading indicates that the anomalies in either direction are significantly different from zero at the 5% and 10% levels, respectively. The wind vector scale is shown in the top right of each panel; units: m s^{-1}.

pared to those in the PRE epoch (Figs. 4b and d). In addition, a notable difference is seen in the spring AO-associated relative humidity at 600 hPa between the two epochs. During the PRE epoch, significant positive relative humidity anomalies are observed to the east of 150°E (Fig. 4e), while relative humidity anomalies are very weak during the POST epoch (Fig. 4f). These results indicate that the large-scale dynamic and thermodynamic conditions in association with the spring AO variability during the PRE epoch provide a more favorable environment for TC formation in the eastern part of the WNP compared with those during the POST epoch. This result is consistent with previous observational and modeling analyses (e.g., Gray, 1968; Cao et al., 2014b, 2014c).

4.2. *Interdecadal changes of spring AO-related anomalies in the simultaneous spring*

In the previous section, we demonstrated that the interdecadal change in the connection between spring AO and following summer TC activity is attributed to the spring AO-related SST anomalies in the tropical central–eastern Pacific, which play an important role in relaying the influence of spring AO on the following summer TC formation. During the PRE epoch, significantly positive SST anomalies can be seen in the tropical central–eastern Pacific in the following summer. These spring AO-related SST anomalies result in atmospheric circulation anomalies that are favorable for the TC formation during the same periods. In contrast, during the POST epoch, spring AO-related SST anomalies are weak over the tropical Pacific during the following summer. Hence, in this epoch, the influence of spring AO cannot be connected well to the following summer TC activity.

An important question to be addressed is why spring AO-related SST anomalies in the tropical central–eastern Pacific in the following summer are more significant and stronger during the PRE epoch than during the POST epoch. To address this issue, we first examine spring AO-related atmospheric circulation anomalies in the simultaneous spring. Figure 5 displays anomalies of SLP, winds at 850 hPa, and geopotential height at 500 hPa and 200 hPa in the spring, obtained as regression upon the spring AO index. Atmospheric circulation anomalies in association with the spring AO over the North Pacific in both the PRE and POST epochs display a barotropic vertical structure (Fig. 5), consistent with the previous findings regarding wintertime AO (Thompson and Wallace, 1998, 2000). Notable differences in the atmospheric circulation anomalies can be observed over the North Pacific between the PRE and POST epochs. During the PRE epoch, atmospheric circulation anomalies over the North Pacific display a dipole structure, with significantly positive SLP, geopotential height and anticyclonic circulation anomalies over the midlatitudes of the North Pacific, and signifi-

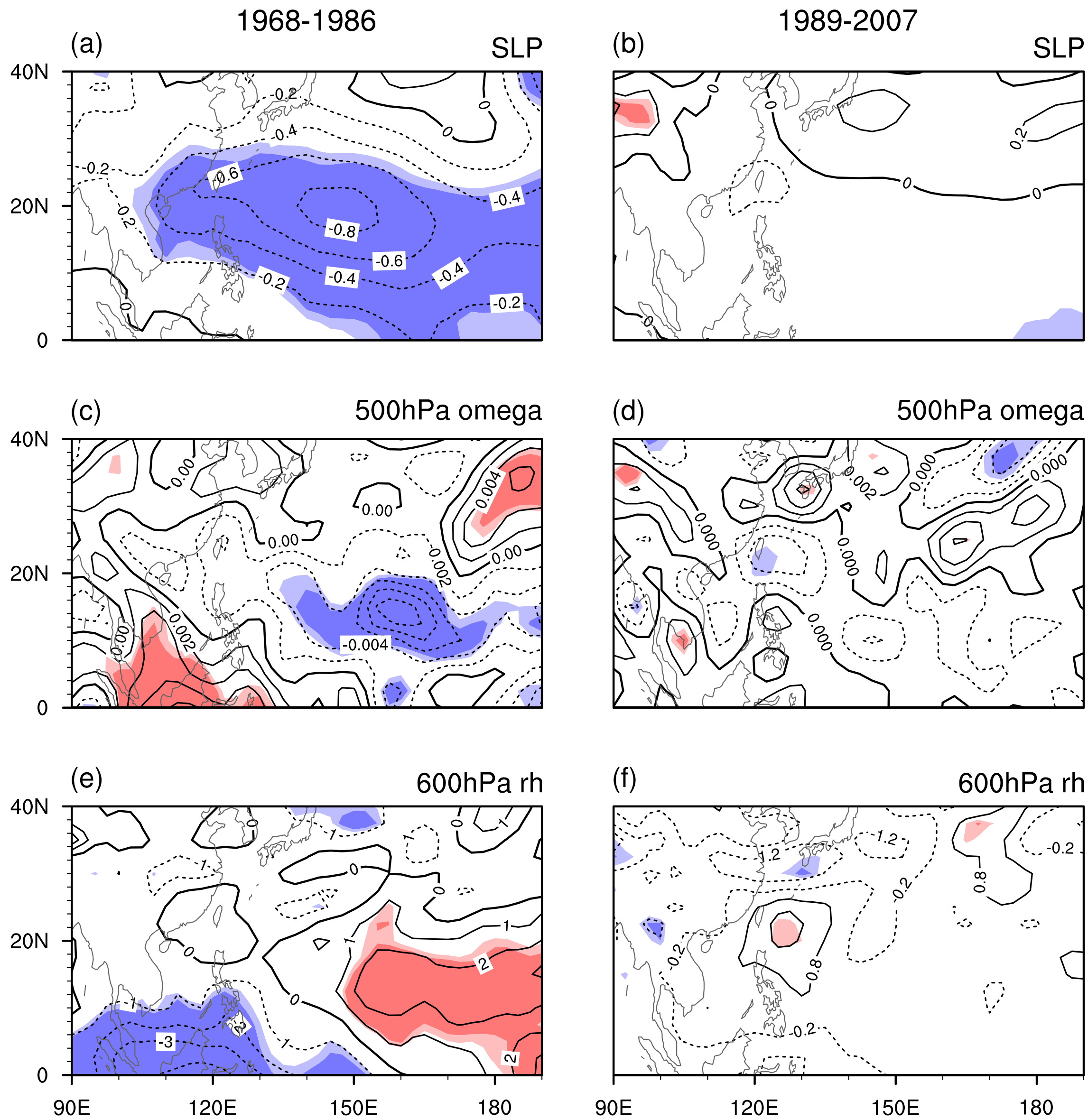

Fig. 4. As in Fig. 2 but for anomalies of the following summer (a, b) SLP, (c, d) vertical velocity at 500 hPa, and (e, f) relative humidity at 600 hPa obtained as regressions upon the normalized spring AO index in (a, c, e) 1968–1986 and (b, d, f) 1989–2007. Contour intervals are 0.2 hPa in (a, b), 0.002 Pa s^{-1} in (c, d), and 1 % in (e, f).

cantly negative SLP, geopotential height and cyclonic circulation anomalies over the subtropical western–central North Pacific (Figs. 5a, c, e, and g). It should be mentioned that significant westerly wind anomalies can be observed over the tropical WNP to the south of the spring AO-related cyclonic circulation anomalies (Fig. 5c). Previous studies have demonstrated that spring AO-related westerly wind anomalies play a critical role in relaying the influence of spring AO on the tropical central–eastern Pacific SST anomalies several months later through stimulating an eastward propagating Kelvin wave (e.g., Barnett et al., 1989; Huang et al., 2001; Nakamura et al., 2006, 2007; Chen et al., 2014a).

During the POST epoch, atmospheric circulation anomalies over the North Pacific are much weaker and located more northeastward compared with those in the PRE epoch (Fig. 5). It should be emphasized that westerly wind anomalies over the tropical WNP are extremely weak. As a result, during the POST epoch, the influence of spring AO on the following summer SST over the tropical central–eastern Pacific is weak.

Previous studies have demonstrated that the interaction between low-frequency mean flow and synoptic-scale eddies (also called the storm track) plays an important role in forming the spring AO-related atmospheric circulation anomalies over the subtropical North Pacific (Gong et al., 2011; Chen et al., 2014a, 2014b). They showed that in the positive spring AO years, significant easterly wind anomalies can be found over the midlatitudes of the North Pacific. Westward acceleration (i.e., easterly anomalies) of the low-frequency mean flow is accompanied by weakened synoptic-scale eddy

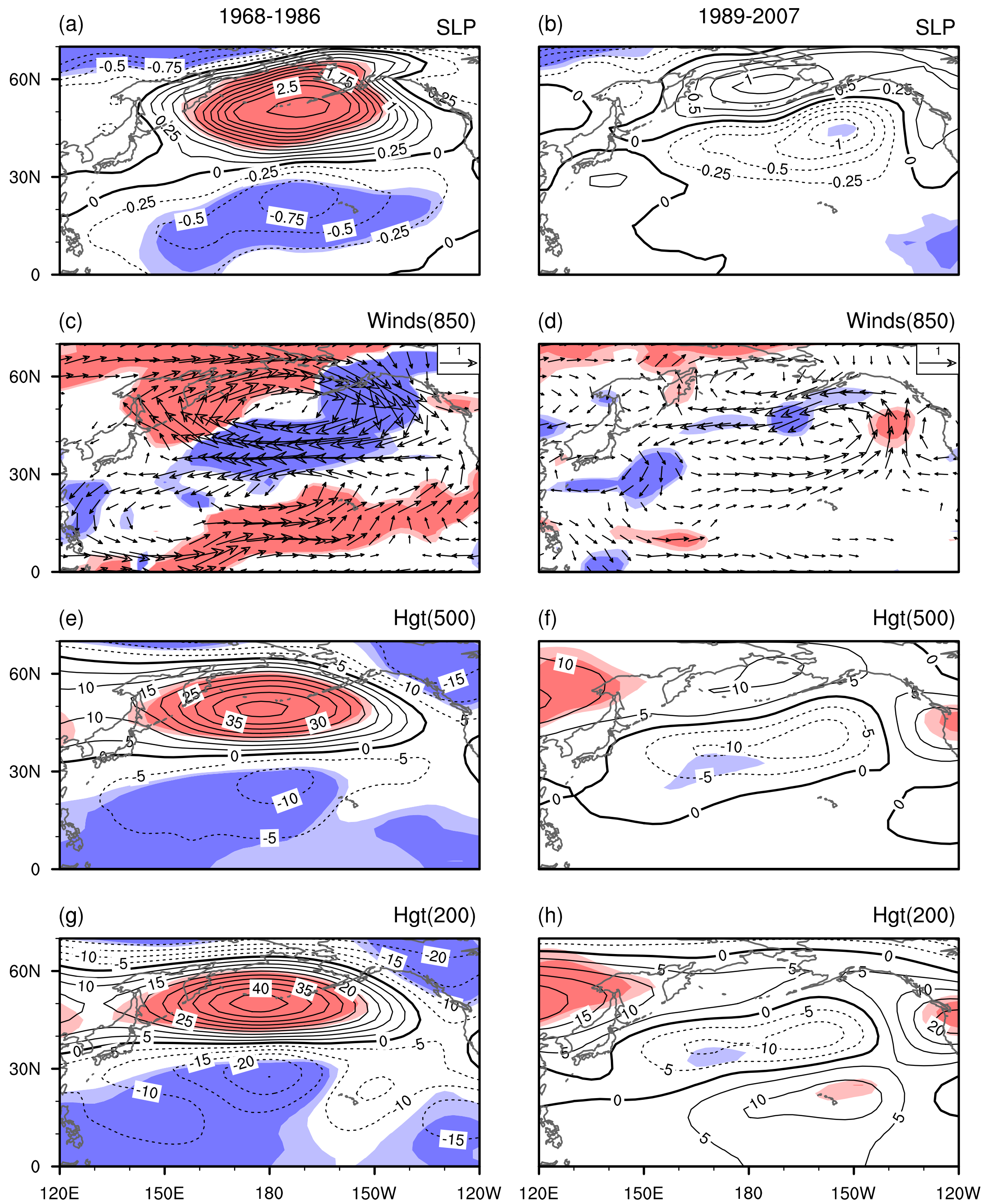

Fig. 5. Anomalies of spring (MAM-averaged) (a, b) SLP, (c, d) 850 hPa winds, and geopotential height at (e, f) 500 hPa and (g, h) 200 hPa regressed upon the normalized spring AO index in (a, c, e, g) 1968–1986 and (b, d, f, h) 1989–2007. The dark and light shading in (a, b) and (e–h) indicates that the anomalies are significantly different from zero at the 5% and 10% level, respectively. The heavy and light shading in (c, d) indicates that the anomalies in either direction are significantly different from zero at the 5% and 10% levels, respectively. The wind vector scale is shown in the top right of each panel; units: m s^{-1}. Contour intervals are 0.25 hPa in (a, b) and 5 m in (e–h).

activity, as well as positive geopotential height tendency immediately to its north and negative geopotential height tendency to its south, and vice versa (Lau, 1988). The above-mentioned processes can explain the formation of spring AO-related cyclonic circulation anomalies over the subtropical WNP (Figs. 5c and d). From Figs. 5c and d, easterly wind anomalies at 850 hPa over the midlatitudes of the North Pacific are stronger and located more southward in the PRE epoch than in the POST epoch. The anomalous zonal winds in the upper troposphere (200 hPa) in the PRE and POST

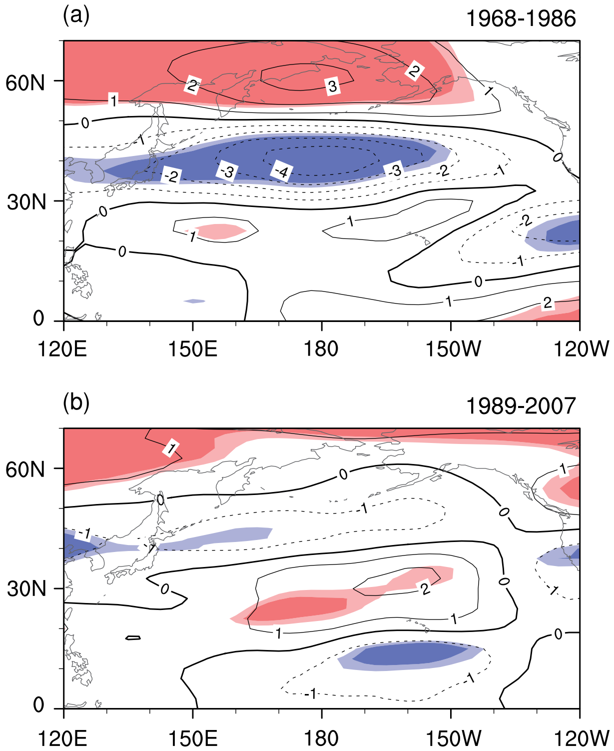

Fig. 6. Anomalies of spring (MAM-averaged) zonal winds at 200 hPa regressed upon the normalized spring AO index in (a) 1968–1986 and (b) 1989–2007. The heavy and light shading in (a, b) indicates that the anomalies are significantly different from zero at the 5% and 10% levels, respectively. Contour intervals are 1 m s^{-1} in (a, b).

epochs are further examined in Fig. 6. As expected, the easterly wind anomalies at 200 hPa over the midlatitudes of the North Pacific are also much weaker in the POST epoch than in the PRE epoch, which is consistent with the results obtained in the lower troposphere. Associated with the wind anomalies shown in Fig. 6, storm track anomalies also display the notable differences over the North Pacific during the PRE and POST epochs (Fig. 7). During the PRE epoch, in accompaniment with easterly wind anomalies over the midlatitudes of the North Pacific (Fig. 6a), significantly negative storm track anomalies are observed here (Fig. 7a). Based on the study of Lau (1988), negative storm track anomalies over the midlatitudes of the North Pacific are accompanied by negative geopotential height tendency to its south and positive geopotential height tendency to its north. Hence, negative storm track anomalies over the North Pacific can contribute directly to the formation of negative geopotential height anomalies and cyclonic circulation anomalies over the subtropical North Pacific and to the maintenance of positive geopotential height anomalies over the midlatitudes of the North Pacific. By contrast, during the POST epoch, spring AO-related negative storm track anomalies over the midlatitudes of the North Pacific are much weaker and located more northward compared to those in the PRE epoch (Fig. 7). As a result, negative geopotential height anomalies and cyclonic circulation anomalies over the subtropical North Pacific are much weaker and located more northward in the POST epoch than those in the PRE epoch.

Based on the above analyses, the factors responsible for the interdecadal change of the connection between spring AO and the following summer TC formation may be rooted in the interdecadal change of the spatial pattern of the spring AO over the North Pacific, as shown in Fig. 5. Change of the spatial pattern of the spring AO over the North Pacific would change the spring AO-related anomalies in the subtropical North Pacific via the interaction between synoptic-scale eddies and low-frequency mean flow. In particular, the interaction between synoptic-scale eddies and low-frequency mean flow is much stronger and located more southward in the PRE epoch than in the POST epoch. Therefore, significant westerly wind anomalies tend to be induced over the tropical WNP in the PRE epoch and exert a significant influence on the SST anomalies over the tropical central–eastern Pacific. SST anomalies over the tropical central–eastern Pacific in the following summer in association with spring AO further exert impacts on the activity of TCs over the WNP. In contrast, the westerly wind anomalies over the WNP are insignificant and weak in the POST epoch. Undoubtedly, the spring AO cannot prolong its influences on the following summer TC activity over the WNP.

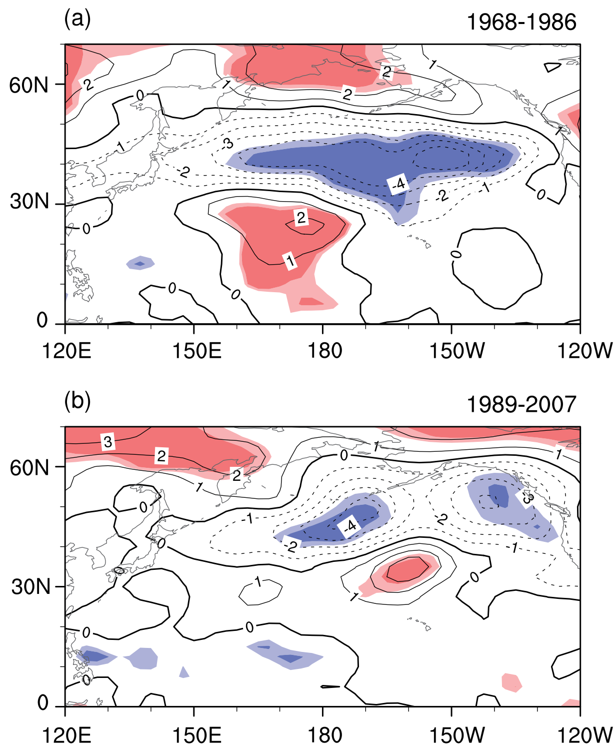

Fig. 7. Anomalies of spring (MAM-averaged) storm track at 200 hPa regressed upon the normalized spring AO index in (a) 1968–1986 and (b) 1989–2007. The heavy and light shading in (a, b) indicates that the anomalies are significantly different from zero at the 5% and 10% levels, respectively. Contour intervals are 1 m in (a, b).

5. Summary

Previous studies regarding the effects of the AO on TC formation focused on the connection between spring AO and following summer TC formation frequency over East Asia, or between AO and TC formation frequency over the WNP in the concurrent summer. In this study, we have presented observational evidence to show that a significant connection exists between the spring AO and the following summer TC frequency in the eastern part of the WNP (150°–180°E) during the 1968–1986 (PRE) epoch, and their connection becomes weak during the 1989–2007 (POST) epoch. In contrast, the connection between the spring AO and the following summer TC frequency in the western part of the WNP (120°–150°E) remains insignificant in the whole analysis period.

We compared the spring AO-related SST, atmospheric dynamic, and thermodynamic conditions between the PRE and POST epochs to understand the interdecadal change in the relationship between the spring AO and the following summer TC frequency over the eastern part of the WNP. Results show that SST anomalies in the tropical central–eastern Pacific play a key role in relaying the influence of the spring AO on the following summer TC formation. During the PRE epoch, an El Niño-like SST pattern over the tropical Pacific can be seen in the following summer in association with the spring AO variability. Meanwhile, a remarkable cyclonic circulation anomaly at the lower level is present over the WNP as a Matsuno–Gill type atmospheric response to the tropical central–eastern Pacific SST anomalies (Matsuno, 1966; Gill, 1980). In addition, an upper-level anticyclonic circulation anomaly, lower surface pressure, enhanced ascending motion, and an obviously positive relative humidity anomaly are found to be strong and significant over the eastern part of the WNP. In contrast, the spring AO-related SST, wind, vertical motion, and relative humidity anomalies are rather weak during the POST epoch. These notable differences indicate that the large-scale dynamic and thermodynamic conditions in relation to the spring AO are more favorable for subsequent summer TC formation over the eastern part of the WNP during the PRE epoch compared to those during the POST epoch.

The differences of spring AO-related cyclonic circulation anomalies over the subtropical WNP between the two epochs are related to the changes in the spatial pattern of the spring AO over the North Pacific. The high pressure center over the midlatitudes of the North Pacific and associated easterly wind anomalies to its southern flank are much stronger and located more southward during the PRE epoch than those during the POST epoch. The westerly jet stream over the East Asia–North Pacific region is significantly affected by the spring AO during the PRE epoch, while spring AO influences on the westerly jet stream over the North Pacific is weak during the POST epoch. As a result, the interaction between synoptic-scale eddies and low-frequency mean flow may be much stronger during the PRE epoch than during the POST epoch. This can explain the stronger cyclonic circulation anomalies over the WNP and stronger westerly wind anomalies to its south over the tropical WNP during the PRE epoch than during the POST epoch.

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