Does the Antarctic Oscillation modulate tropical cyclone rapid intensification over the western North Pacific?

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Keywords: tropical cyclone, rapid intensification, Antarctic Oscillation

Abstract
This study investigates the modulation of tropical cyclone (TCs) rapid intensification (RI) over the western North Pacific by the Antarctic Oscillation. There is a significant inverse relationship between basinwide RI number during July–November from 1982 to 2020 and the simultaneous Antarctic Oscillation index. During positive Antarctic Oscillation years, RI occurrence is significantly suppressed over the main RI region (10\degree–20\degree N and 125\degree–150\degree E) and slightly enhanced over the South China Sea. By contrast, during negative Antarctic Oscillation years, RI is significantly enhanced over the main RI region and slightly suppressed over the South China Sea. The Antarctic Oscillation influences western North Pacific RI occurrence mainly through modulation of the large-scale dynamic environment. During positive Antarctic Oscillation years, increases in 850–200 hPa vertical wind shear and decreases in both low-level vorticity and upper-level divergence all suppress RI occurrence over the main RI region, while thermodynamic variables (e.g. TC heat potential, maximum potential intensity and low-to-middle level relative humidity) show mostly weak changes. These changes in dynamic factors can be linked to a low-level anomalous western North Pacific anticyclone triggered by the positive phase of the Antarctic Oscillation and a low-level anomalous western North Pacific cyclone generated by a negative phase of the Antarctic Oscillation.

1. Introduction
Rapid intensification (RI) is a near-universal feature of the strongest tropical cyclones (TCs) (e.g. Category 4–5 TCs on the Saffir–Simpson Hurricane Wind Scale; \geq 113 \text{kt} one-minute averaged maximum sustained winds), with RI often being defined to be an increase in TC intensity of at least 30 \text{kt} in a 24 h period (Kaplan and DeMaria 2003). RI is particularly difficult to simulate and predict and consequently continues to pose significant challenges for operational forecasting (Knaff et al 2018). Kaplan and DeMaria (2003) found that 83\% of all major (Category 3–5) hurricanes in the North Atlantic underwent RI at some point during their lifetime, and Weinkle et al (2018) found that \sim 80\% of total normalized continental US hurricane damage from 1900 to 2017 was generated by major hurricanes. Consequently, we can infer that most damage from TCs is generated by storms that underwent RI at some point during their lifetime, with Typhoon Haiyan (2013) and Hurricane Patricia (2015) providing recent examples of TCs experiencing extreme RI (Hong and Wu 2021). There is thus an urgent need to understand the features, changes and mechanisms related to RI.

It is commonly recognized that both internal dynamics and environmental conditions influence the RI process (Hendricks et al 2010). Compared with internal dynamics, environmental conditions controlled by climate variability play a dominant role in affecting RI changes on interannual to interdecadal timescales (Hong and Wu 2021). Here, we focus on the variability of RI events over the western North Pacific (WNP). On average, the WNP has more frequent RI than any other global...
TC basin (Lee et al 2016). On interannual timescales, there is a significant simultaneous correlation between WNP RI events and El Niño-Southern Oscillation (ENSO) (Wang and Zhou 2008). The average number of RI events is notably higher in El Niño years than in La Niña years, driven by an enhanced and southeastward-extended monsoon trough over the WNP (Wang and Zhou 2008). Moreover, the proportion of TCs experiencing RI to the total number of TCs have distinctly distinct features between eastern Pacific El Niño and central Pacific El Niño years (Shi et al 2020). On average, higher ratios are observed during July–October (November–December) for eastern (central) Pacific El Niño years. On interdecadal timescales, there is an inverse relationship between the number of WNP RI events and the Pacific decadal oscillation (PDO) (Wang et al 2015). RI occurs, on average, more frequently during negative PDO phases than during positive PDO phases (Wang et al 2015, Zhang et al 2020). The increase in RI during a negative PDO phase is likely driven by a northward expansion of equatorial warm water into the main RI region and an associated deepening of the depth of the 26 °C isotherm (Wang et al 2015).

As highlighted in Wang and Zhou (2008), RI occurrence and TC genesis share common environmental preferences, although there are some differences in the environmental factors controlling their temporal variations. As summarized in Song and Klotzbach (2019), various climate modes modulate TC genesis over the WNP, including ENSO, the Indian Ocean Dipole, the North Pacific Gyre Oscillation, the Pacific Meridional Mode, the Atlantic Meridional Mode, the North Atlantic Oscillation, the Antarctic Oscillation, the North Pacific Oscillation, the Arctic Oscillation, the Pacific-Japan teleconnection pattern, the Western Pacific teleconnection pattern and the Pacific-North American pattern. However, it is still unclear whether changes in RI over the WNP are influenced by other climate modes beyond ENSO and the PDO.

Based on the spatial pattern of the correlation between WNP RI frequency and global sea level pressure (SLP) (figure 1(c)), this study examines the possible influence of the Antarctic Oscillation (AAO) on RI occurrence over the WNP. The AAO, also referred to as the Southern Annular Mode (SAM) (Screen et al 2018), is characterized by a large-scale oscillation of SLP between polar and midlatitude regions of the Southern Hemisphere (SH) (Karoly 1990, Gong and Wang 1999, Thompson and Wallace 2000). While the AAO is the leading mode of extratropical atmospheric variability in the SH, it can influence the Northern Hemisphere (NH) via its influences on the tropics (Song et al 2009). In particular, several previous publications have reported the influence of the AAO on WNP TC activity (Ho et al 2005, Wang and Fan 2007, Choi and Wang 2020). Ho et al (2005) found that TC genesis was higher near the coast of the Philippines and east of 140° E and lower over the South China Sea (SCS) in positive AAO years relative to negative AAO years. They also showed an increase in TC passage over the East China Sea, Japan and the Philippines during positive AAO years compared with negative AAO years. Wang and Fan (2007) found a significant negative correlation between WNP typhoon frequency and the AAO index. They attributed this negative correlation to larger (smaller) vertical zonal wind shear and cooler (warmer) ocean surface water over the tropical WNP in positive (negative) AAO phases. Choi and Wang (2020) found that the genesis longitude of the last TC of the year was significantly inversely correlated with the AAO. They found a westward (eastward) migration of the mean longitude of the final TC genesis of the year in positive (negative) phases of the AAO.

Given the previously-documented AAO-WNP TC genesis relationship over the WNP, we now investigate the linkage between the AAO and WNP RI. We focus on how the AAO influences spatiotemporal changes in WNP RI through modulation of the large-scale environment. The remainder of this study is organized as follows. Section 2 discusses the TC and environmental datasets employed as well as the specific methodology used for the analysis in this manuscript. Section 3 presents spatiotemporal variations in WNP RI events driven by the AAO. Section 4 discusses how the AAO modulates changes in several environmental variables. The study concludes with a summary in section 5.

2. Data and methods

The original 6 hourly TC best track data from the Joint Typhoon Warning Center used in this study are provided by the International Best Track Archive for Climate Stewardship (IBTrACS) (v04r00; Knapp et al 2010), including TC central position and maximum sustained wind (Vmax) over the WNP (north of the equator and 100° E–180°). We focus on the recent period of 1982–2020, which is generally considered to be of higher quality given the near-global coverage of geostationary satellites starting in 1982 (Moon et al 2019). Similar to previous RI-related studies (Kaplan and DeMaria 2003, Kaplan et al 2010, Shu et al 2012, Knaff et al 2018, Shimada et al 2020), an RI event is defined as a 24 h Vmax (ΔV24) change of at least 30 kt over water—approximately the 95th percentile of ΔV24s for all TC cases. Note that RI events defined here are counted in 6 h intervals, so a TC may experience more than one RI event during its lifetime. We use the term ‘RI event’ and ‘RI occurrence’ interchangeably throughout the remainder of this manuscript. RI is investigated during July–November in this study. July–November includes the majority (84%) of
RI events that occur over the entire year (Wang and Zhou 2008, Ge et al 2018).

Favorable large-scale environmental factors are critical for the occurrence of RI, including high tropical cyclone heat potential (TCHP), large maximum potential intensity (MPI), a moist lower-to-middle troposphere, low vertical wind shear (VWS), an anomalous low-level cyclonic circulation and upper-level divergent flow (Shu et al 2012, Fudeyasu et al 2018, Knaff et al 2018, Liang et al 2021, Zhao et al 2022a, 2022b, Cai et al 2022, Ma and Fei 2022). High TCHP and MPI indicate an increased thermal energy supply for TC intensification. A moist lower-to-middle troposphere can release abundant latent heat associated with deep convection, favoring TC development. Low VWS reduces the vertical tilt of the TC, enhancing chances for TC intensification. An anomalous low-level cyclonic circulation provides additional positive vorticity that enhances the TC circulation, while upper-level divergence aids ascending motion that leads to more intense convection.

TCHP is a measure of ocean heat content that is warmer than 26 °C (DeMaria et al 2005) and is derived from monthly subsurface temperature profiles from the control member of the ECMWF Ocean Reanalysis System 5 (ORAS5; Zuo et al 2019) on a 1° × 1° grid. Using the formula of Emanuel (1988), MPI can be estimated by:

\[
\text{MPI} = \frac{C_k \text{ SST} - T_0}{C_d} (k^*_s - k),
\]

where \( T_0 \) is the temperature of the outflow, \( k^*_s \) is the saturation enthalpy of air at the sea surface, \( k \) is the enthalpy of the boundary layer air overlying the surface, and \( C_k \) and \( C_d \) are the surface exchange coefficients of enthalpy and momentum, respectively. We calculate MPI from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al 2003) with a resolution of 1° × 1° and the fifth generation European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis of the global climate (ERA5; Hersbach et al 2020), with a resolution of 0.25° × 0.25°. Before calculating MPI, the temperature and humidity profiles in the ERA5 are re-gridded onto a 1° × 1° grid to match the resolution of the sea surface temperature (SST) field in HadISST. All other environmental factors are calculated from ERA5. We also note that HadISST and ERA5 provide the primary forcing fields for ORAS5, leading to physical consistency between these datasets (Zuo et al 2019). July–November averages of the Niño-3.4 SST anomaly are obtained from NOAA’s Earth System Research Laboratory Physical Sciences Division.

Following Gong and Wang (1999), we use the normalized difference between the zonally averaged July–November deviation of ERA5 SLP from the climatological mean between 40° S and 65° S:

\[
\text{AAO index} = \frac{\text{SLP}_{40^\circ S} - \text{SLP}_{65^\circ S}}{\sigma_{\text{SLP}_{40^\circ S}}} - \frac{\text{SLP}_{65^\circ S} - \text{SLP}_{40^\circ S}}{\sigma_{\text{SLP}_{65^\circ S}}},
\]

where the overbar denotes the July–November mean, the square bracket indicates the climatological (1982–2020) mean for a selected longitudinal range, and \( \sigma \) stands for the standard deviation for this longitudinal range. The Gong and Wang (1999) definition calculates the AAO index by averaging across all longitudes. Moreover, as suggested in Rudeva et al (2019), we also calculate regional AAO indices for five sectors in the SH: the Atlantic (60° W–20° E), the Indian (30°–90° E), the Australian (90°–150° E), the western Pacific (150° E–150° W) and the eastern Pacific (150°–75° W).

The significance levels (p) of correlation coefficients (r), partial correlation coefficients and the differences in means between two samples are all estimated using a two-tailed Student’s t test. In evaluating statistical significant, the effective sample size proposed by Trenberth (1984) is applied to minimize the influence of autocorrelation.

3. Spatiotemporal RI changes associated with the AAO

Figure 1(a) displays the spatial distribution of the correlation coefficient between WNP RI events during July–November and simultaneous SST. Significant positive correlations are observed over a large region extending from the equatorial central Pacific to the equatorial and extratropical eastern Pacific in both hemispheres, while weak negative correlations occur over the western Pacific. This is a typical SST pattern associated with El Niño. Figure 1(b) highlights a significant relationship between the number of WNP RI events and the Niño-3.4 SST anomaly (\( r = 0.51, p < 0.01 \)), consistent with Wang and Zhou (2008).

A typical SLP pattern associated with El Niño is found in the spatial distribution of correlation coefficients between the number of RI events and simultaneous SLP, consisting of negative correlations over the central and eastern Pacific and positive correlations over the western Pacific and the Indian Ocean (figure 1(c)). We also find significant positive correlations over the SH polar regions, similar to the SLP pattern associated with a negative AAO. There is an inverse relationship between the number of RI events and the AAO index (figure 1(d)), with a small but significant correlation coefficient of −0.37 (\( p = 0.02 \)), highlighting lower (higher) RI frequency during positive (negative) AAO phases. To confirm this, we next
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Figure 1. (a), (c), (e) Spatial distributions of (a), (c) correlations and (e) partial correlations between July–November WNP RI events and (a) July–November SST and (c), (e) July–November SLP during the period from 1982 to 2020. Hatched areas denote correlations significant at the 0.05 level. In (e), the effect of the Niño-3.4 SST anomaly is linearly removed via partial correlations. (b), (d), (f) Annual variations of (b), (d) WNP RI events during July–November and (f) RI event residual after regressing on the Niño-3.4 SST anomaly from 1982 to 2020, as well as July–November averages of (b) the Niño-3.4 SST anomaly and (d), (f) the AAO index. Correlation coefficients and their significance levels are shown in each panel.

control for the linear effect of ENSO using partial correlation analysis. The spatial distribution of the partial correlation between RI frequency and SLP is more like the negative AAO pattern, with significant positive (negative) values over the polar (midlatitude) regions in the SH (figure 1(e)). After linearly removing the ENSO influence through regression on the Niño-3.4 SST anomaly, the correlation between the number of RI events and the AAO index remains almost unchanged ($r = -0.34$, $p = 0.04$) (figure 1(f)). This implies that the AAO-RI relationship is independent of ENSO, given the very weak relationship between the AAO and ENSO ($r = -0.16$, $p = 0.33$).

Figure 2(a) illustrates the lag correlation coefficients of WNP RI events during July–November with the AAO index averaged during different seasons. Significant correlations are only found during the timespan from July–September to October–December, which covers the full July–November period investigated here. By contrast, there is no significant relationship between WNP RI events and the AAO state during the preceding winter and spring. This implies that the AAO-RI relationship is primarily simultaneous.

Figure 2(b) shows the correlations between WNP RI frequency and five regional AAO indices between 1982 and 2020. Weak and insignificant AAO-RI correlations are observed in the Atlantic, Indian and Australian sectors. Significant correlations are only found in the Pacific sector, with the highest correlation ($r = -0.40$) occurring in the western Pacific sector. This implies that among different SH sectors,
the AAO change in the western Pacific has the greatest impact on modulating WNP RI occurrence.

In order to compare the spatial differences in RI occurrence between different AAO phases, the seven highest AAO years (hereafter positive AAO years) and the seven lowest AAO years (hereafter negative AAO years) from 1982 to 2020 are selected based on the July–November-averaged value of the AAO index, after excluding ENSO years (Table 1). El Niño (La Niña) years are defined as July–November averages of the Niño-3.4 SST anomaly greater (less) than +0.5 °C (−0.5 °C), as has been done in previous studies (Ho et al 2005, Choi and Byun 2010, Choi and Moon 2012). There are 17 ENSO years in the full 39 year dataset, while the remaining 22 years are used for identifying positive and negative AAO years. Consequently, since we select 14 years from the remaining 22 years, approximately 2/3 of ENSO neutral years are chosen for the analysis. Given that the spatial distributions of anomalies of RI occurrence and environmental factors during positive AAO years are almost

![Figure 2](image_url)

**Figure 2.** (a) Lead and lag correlation coefficients of July–November WNP RI events with the AAO index averaged during different seasons. The horizontal dashed line refers to the 0.05 significance level based on a Student’s *t*-test. (b) Simultaneous correlation coefficients of July–November WNP RI events with regional AAO indices calculated over different portions of the SH. The horizontal dashed line refers to the 0.05 significance level based on a Student’s *t*-test. Significant correlations are hatched.

**Table 1.** List of positive (the seven highest) AAO years and negative (the seven lowest) AAO years during July–November from 1982 to 2020, as well as the number of corresponding WNP RI events.

| Year | Positive AAO RI events | Year | Negative AAO RI events |
|------|------------------------|------|------------------------|
| 1989 | 66                     | 1992 | 63                     |
| 1990 | 45                     | 1996 | 59                     |
| 1993 | 44                     | 2003 | 58                     |
| 2001 | 63                     | 2005 | 63                     |
| 2008 | 47                     | 2013 | 70                     |
| 2012 | 43                     | 2014 | 57                     |
| 2017 | 32                     | 2019 | 66                     |
| Mean | 48.4                   | Mean | 60.4                   |
Figure 3. (a) Climatological (1982–2020) July–November RI occurrence. (b) Difference in July–November RI occurrence between positive and negative AAO years. Black crosses denote differences significant at the 0.05 level. The black dashed rectangle in (a) and (b) denote the main RI region (10°–20° N, 125°–150° E). Positive and negative AAO years are listed in table 1.

a mirror image of those during negative AAO years (figures not shown), we only display differences in these variables between positive and negative AAO years in the analysis that follows. The average number of RI events are 48.4 and 60.4 in positive and negative AAO years, respectively, with a difference that is statistically significant at the 0.05 level.

The climatological (1982–2020) distribution of WNP RI occurrence during July–November is displayed in figure 3(a) by counting 6 h RI events over individual 5° × 5° grids. Higher RI occurrences are observed in a region east of the Philippines (10°–20° N and 125°–150° E; the main RI region), corresponding to the most TC-active region over the WNP. Generally speaking, RI occurrence is suppressed east of 125° E and enhanced west of 125° E during positive AAO years compared with negative AAO years (figure 3(b)). Given that the region with fewer RI occurrences is much larger than the region with more RI occurrences, the number of basinwide RI events is, on average, reduced during positive AAO years.

In addition, during positive AAO years, significant decreases in RI events occur over the main RI region. Although the largest overall increases in RI occurrence are found over the SCS, they are not statistically significant at even a relaxed 0.10 level. This east-west dipolar pattern is different from changes in RI occurrence induced by ENSO. As reported in Wang and Zhou (2008), mean RI occurrence tends to migrate equatorward and eastward during El Niño years and tends to migrate poleward and westward during La Niña years. Consequently, ENSO-related spatial changes in RI occurrence exhibit a southeast-northwest dipolar pattern.

4. Environmental changes associated with the AAO

Environmental conditions play a critical role in controlling RI (Wang and Zhou 2008, Hong and Wu 2021). Figure 4 displays differences in environmental variables between positive and negative AAO years. These variables have been found to influence RI activity not only on synoptic timescales (Shu et al 2012, Knaff et al 2018) but also on interannual to interdecadal timescales (Wang and Zhou 2008, Wang et al 2015). Similar spatial patterns are observed in differences of thermodynamic variables (TCHP, MPI
Differences in environmental variables between positive and negative AAO years, including (a) TCHP, (b) MPI, (c) 700–500 hPa relative humidity, (d) 850–200 hPa vertical wind shear, (e) 850 hPa relative vorticity and (f) 200 hPa divergence. Hatched areas denote differences significant at the 0.05 level. The green dashed rectangle highlights the main RI region. Positive and negative AAO years are listed in Table 1.

figure 4. Differences in environmental variables between positive and negative AAO years, including (a) TCHP, (b) MPI, (c) 700–500 hPa relative humidity, (d) 850–200 hPa vertical wind shear, (e) 850 hPa relative vorticity and (f) 200 hPa divergence. Hatched areas denote differences significant at the 0.05 level. The green dashed rectangle highlights the main RI region. Positive and negative AAO years are listed in table 1.

and 700–500 hPa relative humidity) between positive and negative AAO years (figures 4(a)–(c)). During positive AAO years, there are decreases (increases) in TCHP, MPI and 700–500 hPa relative humidity east (west) of 140° E. However, there are no significant differences in TCHP and MPI over almost all of the WNP, including the main RI region (figures 4(a) and (b)). Given that the oceanic state is considered in these two variables, the AAO appears to play a minor role in modifying oceanic conditions in the WNP. Sasaki and Motoi (2022) found an accelerated increase in TCHP over the main RI region over the past six decades (1955–2020), which they attributed to a combination of the linear trend and a phase change of the PDO from positive to negative. The significant TCHP changes in Sasaki and Motoi (2022) were observed mainly on decadal timescales or longer, while the insignificant TCHP differences in our study were caused by AAO phase changes on interannual timescales. By contrast, significant increases in TCHP and MPI are mainly concentrated over the SCS (figures 4(a) and (b)). 700–500 hPa relative humidity show significant decreases in a small region from 10° to 20° N and near 150° E (figure 4(c)). Significant moisture reductions are found over the eastern part of the main RI region. By comparison, there are no significant moisture changes over the SCS.

Compared with thermodynamic variables, there are larger regions with significant differences in dynamic variables between positive and negative AAO years (figures 4(d)–(f)). During positive AAO years, significant increases in 850–200 hPa VWS occur over the subtropical WNP, while significant VWS decreases occur over the equatorial western WNP (figure 4(d)). Significant differences in 850 hPa relative vorticity are mainly concentrated west of 140° E (figure 4(e)). There are significant reductions in relative vorticity over both the western part of the main RI region and the SCS. In addition, 200 hPa divergence shows strong differences east of 125° E, while only changing weakly over the SCS (figure 4(f)).

In summary, over the main RI region, an increase in VWS and decreases in low-to-mid-level humidity, low-level vorticity and upper-level divergence all act to reduce RI occurrence during positive AAO years compared with negative AAO years, while changes in TCHP and MPI are small (figure 4). By comparison, over the SCS, the RI-favoring effect of greater TCHP and MPI is likely to partly balance the RI-suppressing effect of lower 850 hPa vorticity, leading to insignificant RI occurrence changes (figures 4(a), (b) and (e)).
Although thermodynamic factors were found to play an important role in modulating WNP RI activity, as shown in RI-related numerical experiments (e.g. Oey and Huang 2021) and trend analysis (e.g. Song et al 2021), they appear to only play a small contribution in the interannual AAO-RI relationship.

We show that dynamic variables likely play a more important role in influencing RI occurrence over the main RI region than thermodynamic variables do. Figure 5 shows how the AAO modulates these dynamic factors. Climatologically, the main RI region corresponds to the southwestern edge of the western Pacific subtropical high at lower levels (figure 5(a)) and the northeastern part of the South Asian high at upper levels (figure 5(b)). Associated with these circulation patterns, are westerlies and easterlies over the main RI region at 850 hPa and 200 hPa, respectively. During positive AAO years, there are significantly lower heights over the polar region and higher heights over the midlatitude region in the SH at both 850 hPa and 200 hPa (figures 5(c) and (d)). The AAO exhibits a quasi-barotropic structure throughout the troposphere, extending from the surface to 200 hPa (Hu et al 2016). Consistent with Ho et al (2005) and Zhu (2009), an anomalous counterclockwise circulation at 850 hPa is shown east of Australia, centered at 45° S, 180° (figure 5(c)). North of this anomalous anticyclone, significant anomalous southeasterlies are observed, extending from the subtropics to the tropics between 160° E and 180°. These anomalous southeasterlies further strengthen the cross-equatorial flow over the western equatorial Pacific, which then turns towards the north due to the Coriolis effect. As a consequence, an anomalous anticyclone is formed over the WNP, associated with significantly higher 850 hPa heights. By contrast, at higher levels (e.g. 200 hPa), the AAO influence is mainly limited to the SH, since there are almost no significant changes in heights and wind vectors in the NH (figure 5(d)). We find over the WNP that the AAO primarily modulates the lower atmosphere, given the significant height changes at 850 hPa.

During positive AAO years, the WNP is mainly controlled by an anomalous anticyclone at 850 hPa and an anomalous cyclone at 200 hPa (figures 5(c) and (d)). At 850 hPa, the main RI region is located over the southern part of the anomalous anticyclone that is characterized by anomalous low-level negative vorticity and easterly winds (figure 5(c)). At 200 hPa, the main RI region is located over the southwestern quadrant of the anomalous cyclone, where there is anomalous upper-level convergence and westerly winds (figure 5(d)). This circulation implies decreases in 850 hPa vorticity and 200 hPa divergence over the
main RI region during positive AAO years. Additionally, because the anomalous winds are of the same direction as the climatological winds at both 850 hPa and 200 hPa, VWS is enhanced over the WNP RI region during positive AAO years.

5. Summary

The modulation of WNP TC RI by the AAO is investigated in this study. There is a significant inverse relationship between the number of WNP RI events during July–November and the simultaneous AAO index during the period from 1982 to 2020, regardless of whether the impact of ENSO is removed. WNP RI events are, on average, lower (greater) in positive (negative) phases of AAO. During positive AAO years, there are significant RI occurrence decreases over the main RI region, while increases in RI occurrence over the SCS are not significant. This east-west dipolar pattern is somewhat different from the southeast-northwest dipolar pattern of ENSO-induced RI changes.

The influence of the AAO on RI occurrence can be explained by differences in large-scale environmental variables. In general, dynamic variables (e.g. 850–200 hPa VWS, 850 hPa relative vorticity and 200 hPa divergence) likely play a more important role in influencing RI occurrence over most of the WNP than do thermodynamic variables (e.g. TCHP, MPI and 700–500 hPa relative humidity). During positive AAO years, an increase in VWS and decreases in low-to-mid-level relative humidity, low-level vorticity and upper-level divergence all suppress RI occurrence over the main RI region, while changes in TCHP and MPI are weak over the main RI region. Positive phases of the AAO are associated with an anomalous anticyclone at 850 hPa and an anomalous cyclone at 200 hPa over the WNP. The main RI region is located over the southern part of the anomalous 850 hPa anticyclone and over the southwestern quadrant of the anomalous 200 hPa cyclone. Associated with these features are anomalous low-level negative vorticity and upper-level convergence over the main RI region, as well as anomalous low-level easterlies and upper-level westerlies that enhance VWS. Additionally, the RI-favoring effect of increased TCHP and MPI is likely partly cancelled out by the RI-suppressing effect of lower 850 hPa vorticity, leading to only weak increases in RI occurrence over the SCS.

Our results are mainly obtained through composite analysis, which has its potential drawbacks (Boschat et al 2016). Our composite analysis does not exclude the potential contribution of a third factor, and consequently these results should be verified by numerical sensitivity tests. We intend to conduct numerical simulations to verify these results in future work.

Data availability statement

All data used in this study are freely available online. Western North Pacific TC best track data provided in IBTrACS are available at: https://doi.org/10.25921/82ty-9e16. The Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST) is available at: www.metoffice.gov.uk/hadobs/hadisst/data/download.html. The fifth generation European Centre for Medium-Range Weather Forecasts (ECMWF) atmospheric reanalysis of the global climate (ERA5) is retrieved from: https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels-monthly-means?tab=form. The European Centre for Medium-Range Weather Forecasts (ECMWF) Ocean Reanalysis System 5 (ORASS) data are provided at: https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-orass?tab=form. The monthly timeseries of the Niño-3.4 SST anomaly is obtained from: https://psl.noaa.gov/data/climateindices/

The data that support the findings of this study are available upon reasonable request from the authors.

Acknowledgments

This work was jointly funded by the National Key Research and Development Program of China (2018YFC1507305), the National Natural Science Foundation of China (61827901, 42175007, 41905001 and 42192552) and the China Postdoctoral Science Foundation (2020M680789). Klotzbach would like to acknowledge financial support from the G Unger Vetlesen Foundation.

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