Roles of interdecadal variability of the western North Pacific monsoon trough in shifting tropical cyclone formation

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Abstract

Consistent with the northward migration of the annual mean latitude of tropical cyclone (TC) lifetime maximum intensity (LMI), the basin-wide mean location of TC formation shifted northward in the western North Pacific (WNP) basin over the past four decades. Whether such a shift was related to the anthropogenic influence is important to understanding the response of TC activity to climate change. Instead of detecting the effects of individual environmental factors on this shift, here we focus on the interdecadal variability of the monsoon trough (MT), within which most TCs in the WNP basin occur, and its roles in the shift of the basin-wide mean location of TC formation using 60-year reanalysis data. Interdecadal variations of the MT exhibit two main modes: one associated with the Pacific decadal oscillation (PDO) and the other associated with the interdecadal Pacific oscillation (IPO). In addition, the north–south shift of the mean latitude of TC formation is accompanied by east–west extension of the tropical upper tropospheric trough (TUTT) and the tropical eastern Pacific cold tongue indicated by the east–west contrast of sea surface temperature (SST) anomalies. The poleward shift of the mean TC formation latitude is closely associated with the IPO mode of the MT. The westward retreat of the northwest-to-southeast-oriented MT and the accompanied westward extension of the TUTT reduced TC formation in the eastern part of the WNP basin when the cold tongue shifted westward. It is indicated that the observed poleward shift of TC formation was mainly attributed to natural variability in recent decades.

Keywords Tropical cyclone formation · Monsoon trough · Interdecadal variability

1 Introduction

Previous studies found that the basin-wide mean latitude of the lifetime maximum intensity (LMI) of tropical cyclones (TCs) in the western North Pacific (WNP) has shifted northward since the early 1980s (Kossin et al. 2014, 2016; Moon et al. 2015; Zhan and Wang 2017). This shift in LMI is consistent with a northward migration of the mean TC formation location (Daloiz and Camargo 2018; Studholme and Gulev 2018; Sharmila and Walsh 2018). Some studies suggested that the shift in TC formation is the result of

the expansion of the tropics and the Hadley circulation in a warmer climate (Lucas et al. 2014; Sharmila and Walsh 2017, 2018; Studholme and Gulev 2018). On the other hand, interdecadal fluctuations of the TC mean formation latitude and the LMI latitude shifts are statistically linked to the interdecadal Pacific oscillation (IPO), since TCs tend to form in the northwestern (southeastern) WNP in the negative (positive) phase of the IPO (Song and Klotzbach 2018). The IPO negative phase corresponded to a La Niña-like sea surface temperature (SST) anomalies, which strengthened the Walker circulation in the tropical Pacific and weakened the MT, contributing to interdecadal variations of TC frequency over the WNP (Huangfu et al. 2017; Zhao et al. 2018c). However, how the tropical SST variability affects TC formation over the WNP via modulating large-scale circulations on interdecadal scales remain elusive. Given the influence of interdecadal variations of large-scale circulations on TC frequency, intensity and tracks (Liu and Chan 2007; Matsuura et al. 2003; Bister and Emanuel 2002; Zhao et al., 2018a, b), it is necessary

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to understand the influence of the interdecadal variability on TC formation and how it contributes to the shift of the formation location.

Although variations of environmental factors such as relative humidity, relative vorticity, vertical motion and vertical wind shear have been extensively examined to investigate changes in TC formation and its future prediction (Emanuel et al. 2004; Hu et al. 2018; Wang and Murakami 2020), TC formation in the WNP basin is closely associated with the summer monsoon system. As one of the large-scale circulations related to TC formation, the monsoon trough (MT) can be identified using the 850-hPa wind or relative vorticity field during July–September (Simpson et al. 1968; Chang et al. 1970; Lau and Lau 1990; Holland 1995). The MT provides favorable conditions for TC formation, including strong low-level cyclonic relative vorticity, high mid-level relative humidity and weak vertical wind shear (Gray 1968; Ramage 1974; Chia and Ropelewski 2002; Chen et al. 2006; Zong and Wu 2015a). During the peak season of TC activity (July–September), most TC formation events in the WNP basin occur within the MT (Holland 1995; Lander 1996; Briegel and Frank 1997; Ritchie and Holland 1999; Zong and Wu 2015b). Variations of the MT associated with the Pacific SST variability have considerable impacts on TC activity over the WNP on interannual and interdecadal scales (Matsuura et al. 2003; Wang and Wu 2016). The cold SST anomalies corresponding to a La Niña-like pattern prevents the eastward extension of the MT, further leading to the decrease in TC formation frequency in the WNP basin (Zhao et al. 2018c). Therefore, variations of the MT and the related large-scale system connected with the Pacific SST on decadal to interdecadal scales is a key to understanding the recent TC activity change including the migration of TC formation locations.

In addition to the MT, the tropical upper tropospheric trough (TUTT) is another large-scale system closely associated with tropical cyclogenesis in the WNP basin (Wu et al. 2015; Wang and Wu 2016, 2018). The TUTT features a semi-permanent trough that extends east-northeast to west-southwest in the eastern Pacific centered around 200 hPa (Sadler 1976). The strong westerly vertical wind shear in the southeast flank of the TUTT generally limits the eastward extension of tropical cyclogenesis in the WNP and thus zonal shifts of the TUTT have great influence on the mean formation location of TCs (Kelley and Mock 1982; Fitzpatrick et al. 1995; Wu et al. 2015). Since the zonal position of the TUTT is related to the diversity of SST patterns in the tropical Pacific, the relatively weak (strong) SST warming in the east tropical Pacific favors

a westward (eastward) shift of the TUTT (Wang and Wu 2018). This suggests that the tropical eastern Pacific SST variability is an essential factor in the zonal shift of the basin-wide mean location of TC formation through modulating tropospheric large-scale circulations.

Unlike previous studies, the objective of this study is to understand the northward shift of the mean latitude of TC formation by focusing on the influence of large-scale circulations. In particular, the linkage of the MT and the TUTT with the relative SST changes over the tropical Pacific is examined to reveal the roles of interdecadal variability in the observed northward shift of TC formation in recent decades.

2 Data and methods

The TC best-track data in the WNP basin are from the Joint Typhoon Warning Center (JTWC). The TC formation location is defined as where the maximum wind speed of a cyclone first reaches 35 knots. The frequency of TC formation is defined as the total number of TCs that developed in each $2.5^\circ \times 2.5^\circ$ box. Our analysis mainly covers the TC peak season (July–September, JAS). The monthly mean winds are from Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) between 1958–2017 with a horizontal resolution of 1.25° latitude $\times$ 1.25° longitude. Following previous studies (Lander 1996; Briegel and Frank 1997; Ritchie and Holland 1999), the MT is defined as a convergence zone between easterlies and westerlies at 850 hPa with strong cyclonic relative vorticity. The migration of the TUTT is represented by the boundary of the zero contour of 200-hPa zonal winds over 10° – 30° N (Wu et al. 2015; Wang and Wu 2016). The interdecadal variations of the MT are derived from an empirical orthogonal function (EOF) analysis of 850-hPa relative vorticity anomaly in the WNP basin after the linear trend is removed. To reduce the influence of interannual variations, a 5-year running mean is applied.

The SST data are from National Oceanic and Atmospheric Administration Extended Reconstructed SST version 5 (ERSST v5) dataset (Huang et al. 2017). To represent the east–west SST contrast in the Pacific, we calculate the SST anomaly obtained by removing the zonal mean SST over the WNP during 1958–2017. The cold tongue in the eastern North Pacific is defined as the zero-contour of SST anomalies along 10° – 20° N. Shifts of the western boundary of the cold tongue indicate the zonal migration of the relative cold anomalous SST extension in the eastern Pacific. We also use the Pacific decadal oscillation (PDO) index (information available online at https://www.esrl.noaa.gov/psd/gcos_wgsp/Timeseries/PDO/) and the interdecadal Pacific oscillation (IPO) index (information available online at <http://www.esrl.noaa.gov/psd/data/timeseries/IPOTPI/>) to understand the air–sea interaction.

Fig. 1 The mean 850-hPa winds (streamlines) and relative vorticity (shading; unit: $\times 10^{-5} \text{ s}^{-1}$) with the frequency of TC formation (contour) over the WNP between July–September (JAS) 1958–2017

3 Two natural modes of the MT and its influences on the location of TC formation

The northwest-to-southeast-oriented MT is roughly between 10° – 20° N over the WNP in boreal summer, coinciding

with the high frequency of TC formation (Fig. 1). The mean latitude of TC formation is 18.4° N in the TC peak season (JAS). In agreement with previous studies (Holland 1995; Lander 1996; Briegel and Frank 1997; Ritchie and Holland 1999; Zong and Wu 2015b), TC formation events in the WNP basin primarily occur within the MT. To obtain interdecadal modes of the MT, we perform EOF analysis to 850-hPa detrended relative vorticity over the domain 90° – 180° E, 0° – 40° N in 1958–2017. The leading EOF mode (EOF1) of relative vorticity shows an east–west dipole along 120° – 150° E south of 10° N (Fig. 2a). The principal component of EOF1 (PC1) displays an oscillation of roughly 30–40 years with the negative phase between the 1970s and the late-1990s, which is closely associated with the PDO index ($r = -0.89$; Fig. 2b). Hereafter EOF1 is referred to as the PDO mode of the MT. In general, the positive (negative) phase of the PDO mode corresponds to the cooling (warming) phase of the PDO. The PDO may lead to a north–south dipole of 500-hPa geopotential height anomaly, which modulates the intensity and extension of the WNP subtropical high (Liu and Chan 2007), further affecting zonal winds and TC formation over the WNP.

Fig. 2 **a** EOF1 of 850-hPa detrended relative vorticity (shading; unit: $\times 10^{-5} \text{ s}^{-1}$) with climatological mean 850-hPa winds (streamlines) and **b** PC1 with the PDO index and the mean TC formation number. **c**, **d**

Same as **a**, **b**, but for EOF2. In **d**, PC2 is with the IPO index. EOF1 and EOF2 explain 23.6% and 14.2% of the total variance respectively

Fig. 3 The 850-hPa regressed winds onto PC1 (streamlines) and the composite of the frequency of TC formation (shading) in **a** the positive and **c** negative phases of EOF1. **b, d** Same as **a, c**, but for EOF2. The difference of 850-hPa regressed winds (vector) and the frequency of TC formation (shading) in the opposite phase of **e** EOF1 and **f**

EOF2 (the positive phase minus the negative phase). In **a–d**, streamlines show climatological mean fields plus wind anomalies in the positive and negative phase. The blue lines indicate the mean location of the MT

To explore changes of the MT location and the associated TC formation locations in the positive and negative phases of the PDO mode, we regress 850-hPa winds onto PC1 and make a TC formation composite by averaging the frequency of TC formation in the two phases (the positive phase: 1958–1970 and 2001–2017, and the negative phase: 1971–2000), respectively (Fig. 3a, c). We find that there is little zonal shift of the MT in the two phases, since the positive anomalies of relative vorticity located to south of 10° N are far removed from the mean position of the MT. Thus, the difference of the mean location of TC formation in the opposite phases is only 0.1° in latitude (18.4° N in the positive phase and 18.5° N in the negative phase) and 0.2° in

longitude (136.8° E in the positive phase and 137.0° E in the negative phase). That is to say, the PDO mode has little influence on the south-north migration of the TC formation locations.

Although the MT activity linked with the PDO has little influence on the shift of the mean TC formation latitude, the number of TCs that form over the WNP experienced an interdecadal decrease during the end of the twentieth century (Yumoto and Matsuura 2001; Chan 2005; Huangfu et al. 2017; Shan and Yu 2020). Not only the TC formation number, but the frequency of intense typhoon occurrence also exhibits decadal to multidecadal variations related to the PDO, since the favorable dynamic conditions (e.g., weak

vertical wind shear) lead to TC further intensification as it moves northwestward during the developing period (Chan 2008). Time series of the basin-wide mean TC formation number over the WNP depict a combination of a decreasing linear trend and multidecadal variations (Fig. 2b). This suggests that the PDO might be one of the factors affecting TC frequency.

The second EOF mode (EOF2) has a significant impact on the zonal shift of the MT. EOF2 exhibits a belt of strong negative anomalies at 10° N east of 125° E, indicating a northwest-to-southeast migration of the MT (Fig. 2c). The principal component of EOF2 (PC2) fluctuates on interdecadal scales and is generally inversely related to the IPO index ($r = -0.58$; Fig. 2d). For convenience, EOF2 is referred to as the IPO mode of the MT, with the positive (negative) phase of IPO mode corresponding to the cold (warm) phase of the IPO. The cold SST anomalies are associated with an anomalous descending motion in the tropical Pacific and an anomalous cyclone is triggered in the eastern Pacific in the upper levels, which enhances the Walker circulation (Zhang and Delworth 2007; Hu et al. 2018; Zhao et al. 2018c). In the lower levels, such anomalies in the vertical motion and the corresponding convective heating associated with the ascending branch of the Walker circulation over the western North Pacific induce easterly anomalies (Gill 1980), leading to the westward shift of the MT (Matsuura et al. 2003; Fig. 3b). Hence, the east boundary of the MT significantly retreats to the west of 140° E in the positive phase of the IPO mode (including: 1968–1979, 1986–1988, 1997–2000 and 2008–2015).

As a result, the meridional migration of the MT associated with the IPO can primarily shift TC formation locations. The mean latitudes of TC formation are 19.0° N in the positive phase and 18.0° N in the negative phase, indicating a statistically significant difference of 1.0° at the 95% confidence level. For the mean TC formation longitude, the difference is 1.2° with significance at the 90% confidence level (136.1° E in the positive phase and 137.3° E in the negative phase). The northwestern shift of the TC formation locations is closely linked to the IPO negative phase, which is in agreement with Song and Klotzbach (2018).

4 Long-term shift of the TC formation location

Consistent with previous studies (Studholme and Gulev 2018; Sharmila and Walsh 2018; Daloz and Camargo 2018), the mean latitude of TC formation locations has shifted northward with a significant linear trend of 0.36° latitude per decade during 1980–2013. This poleward shift becomes weaker during 1980–2017 ($0.27^{\circ} \pm 0.17^{\circ}$ decade $^{-1}$; Fig. 4a). During the period of 1958–2017, however, the trend is insignificant

Fig. 4 **a** Time series of the latitude of TC formation (black curves) with PC2 (blue curves), and **b** the residual of the multivariate regression of the TC formation latitude onto PC1 and PC2. Black dashed lines indicate the long-term linear trend of the latitude of TC formation, and red dashed lines for the shorter period since 1980

($0.10^{\circ} \pm 0.09^{\circ}$ decade $^{-1}$; Fig. 4a), suggesting that the trend over the recent four decades might result from internal variability.

The striking meridional shift of the MT in the IPO mode primarily contributes to the meridional migration of the TC formation location. We find that variations of the mean TC formation latitude are highly correlated to the IPO mode with a significant correlation of $r = 0.76$ (Fig. 4a). To remove the impact of the IPO mode, we exclude the PC2-related variation by a regression from the TC formation latitude. The northward migration of TC formation locations is insignificant in the residual during 1980–2017 ($0.11^{\circ} \pm 0.17^{\circ}$ decade $^{-1}$). To examine the contribution of internal variability on the TC formation location, we further remove the two interdecadal variations by regression with respect to PC1 and PC2. Following this extra step, the poleward shift vanishes in the period since 1980 ($0.08^{\circ} \pm 0.17^{\circ}$ decade $^{-1}$; Fig. 4b) and there is no significant long-term trend in the past 60 years ($0.09^{\circ} \pm 0.09^{\circ}$ decade $^{-1}$; Fig. 4b). This suggests that the long-term poleward migration of TC formation locations is insignificant, while the trend over the past 40 years is mainly attributed to natural variability.

5 Impacts of tropical SST on TC formation poleward migrations

Since the shift of tropical cyclogenesis is closely associated with the relative SST anomaly distribution in the tropical Pacific (Wang and Wu 2018), how the IPO affects TC

formation via modulating large-scale circulations over the Pacific is essential to understanding the internally driven shift of TC formation locations. Thus, we further examine the tropical SST variability by focusing on the east–west contrast of SST anomalies over the Pacific in the two opposite phases of the IPO mode. In addition, the tropical SST is closely linked with the activity of the northeast-to-southwest-oriented TUTT in the upper levels, which has a considerable influence on the eastern boundary of TC formation as a result of unfavorable environments (e.g., large vertical wind shear; Wang and Wu 2018). We also explore the features of the 200-hPa wind field during the same period.

To investigate corresponding changes of the tropical SST and the upper-level atmospheric circulation in the positive

and negative phases of the IPO mode, we regress the mean SST and 200-hPa winds onto PC2 respectively (Fig. 5). The zonal mean SST is removed to better reflect the SST contrast in the Pacific. The boundary of the cold tongue in the eastern North Pacific reaches to 175° E, indicating the extension of cold SST anomalies in the positive phase of the IPO mode. The southern part of the TUTT extends near 10° N and 150° E in the positive phase of the IPO mode, while it retreats northeastward in the opposite phase. In the positive IPO mode, the extended cold SST anomalies suppress convective heating in the central and eastern Pacific, which is unfavorable for TC genesis. The cyclonic anomalous winds are triggered in the eastern Pacific, enhancing westerly anomalies along 10°–20° N and resulting in the southwestward shift of

Fig. 5 **a** The 200-hPa regressed winds (streamlines; anomalies plus climatological mean fields) and **b** SST anomalies onto PC2 (shading; unit: °C) in the positive phase of the IPO mode. **c**, **d** Same as **a**, **b**, but for the negative phase of the IPO mode. The difference of **(e)** 200-hPa

regressed winds (vector) and **(f)** SST anomalies in the opposite phase of the IPO mode (the positive phase minus the negative phase). In **a**, **c**, blue lines indicate the mean location of the TUTT. In **b**, **d**, black contours indicate SST anomalies equal to zero over the North Pacific

the TUTT. In the lower levels, the enhanced Walker circulation induces easterly anomalies to the west of the southern portion of the MT (Gill 1980; Wang and Wu 2018), shifting the MT northwestward (Fig. 3b). As a result, the westward shifts of the TUTT and the corresponding westward movement of the MT increase vertical wind shear that stunts TC formation in the eastern part of the WNP basin.

6 Conclusions

Since most tropical cyclogenesis events occur within the MT over the WNP, the two interdecadal modes of the MT and their roles in the shift of the basin-wide mean location of TC formation are examined over the past 60 years, instead of examining individual environmental factors. While there is no significant long-term trend in the mean latitude of TC formation, the poleward shift of the mean TC formation latitude is closely associated with the IPO mode of the MT. The northwest-to-southeast-oriented MT migration is accompanied by the east–west shifts of the TUTT in the upper troposphere and the tropical eastern Pacific cold tongue (Fig. 6). The westward retreat of the MT and the accompanied westward extension of the TUTT reduced the TC formation in the eastern part of the WNP basin when the cold tongue shifted westward. Therefore, the observed poleward shift

of TC formation locations was mainly attributed to natural variability in recent decades.

Uncertainties may exist in the TC records, the reanalysis dataset, and the SST data used in this study, especially before the satellite era (Landsea 2005; Emanuel 2010). However, our analysis demonstrates the consistent migrations of the MT, the TUTT, and the tropical eastern Pacific cold tongue in the corresponding phase of the IPO mode. This suggests that the uncertainties in the datasets have little influence on the conclusions in this study.

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Fig. 6 Schematic diagram of the impacts of large-scale systems on TC formation in the WNP. The westward extension of the cold tongue in the eastern Pacific induces a decrease in convective heating and triggers cyclonic anomalous winds in the upper troposphere, enhancing westerly anomalies resulting in the southwestward shift of the TUTT. In the lower levels, the extending cold SST anomalies enhance easterly anomalous winds to the west of the bottom part of the TUTT, resulting in the MT retreating northwestward. As a result, the strong vertical wind shear weakens vertical motions over the central and eastern Pacific that is unfavorable for tropical cyclogenesis, resulting in the poleward migration of TC formation in the WNP

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