Association of the Poleward Shift of East Asian Subtropical Upper-Level Jet with Frequent Tropical Cyclone Activities over the Western North Pacific in Summer

XIAN CHEN

College of Meteorology and Oceanography, National University of Defense Technology, Nanjing, China

ZHONG ZHONG

College of Meteorology and Oceanography, National University of Defense Technology, and Jiangsu Collaborative Innovation Center for Climate Change, Nanjing University, Nanjing, China

WEI LU

College of Meteorology and Oceanography, National University of Defense Technology, Nanjing, China

(Manuscript received 25 April 2016, in final form 26 February 2017)

ABSTRACT

The NCEP–NCAR reanalysis dataset and the tropical cyclone (TC) best-track dataset from the Regional Specialized Meteorological Center (RSMC) Tokyo Typhoon Center were employed in the present study to investigate the possible linkage of the meridional displacement of the East Asian subtropical upper-level jet (EASJ) with the TC activity over the western North Pacific (WNP). Results indicate that summertime frequent TC activities would create the poleward shift of the EASJ through a stimulated Pacific–Japan (PJ) teleconnection pattern as well as the changed large-scale meridional temperature gradient. On the contrary, in the inactive TC years, the EASJ is often located more southward than normal with an enhanced intensity. Therefore, TC activities over the WNP are closely related to the location and intensity of the EASJ in summer at the interannual time scale.

1. Introduction

The East Asian subtropical upper-level jet (EASJ), one of the important components of the large-scale circulation system in East Asia, is located in the upper troposphere and the lower stratosphere. It is characterized by strong winds, large horizontal wind shears, and significant intraseasonal and interannual variability (Zhang et al. 2006; Schiemann et al. 2009; Lu et al. 2011). Changes in the EASJ have important influences on cold air activities, cyclone movements, and precipitation distribution (Carillo et al. 2000; Lu 2004; Sampe and Xie 2010; Liao and Zhang 2013). Previous studies have shown that changes in the EASJ are under control of external forcings and internal atmospheric dynamics. Chen and Trenberth (1988) suggested that the coupled orographic and thermal forcing is one important factor that influences the formation and variation of the EASJ. Hou (1998) argued that the intensification and poleward expansion of the Hadley cell could lead to the EASJ intensification. Since the synoptic-scale transient eddy activity in the EASJ is dynamically connected with the jet speed, increases in the synoptic-scale transient eddy activity can lead to an intensified EASJ (Ren et al. 2008; Liao and Zhang 2013). Recent studies have shown that the accumulative effect of mesoscale perturbations inside the EASJ is also one reason for the overall intensification of the EASJ (Zhong et al. 2010; Chen et al. 2014).

To date, studies about the association of tropical cyclone (TC) activity with the EASJ are limited. Based on their sensitivity experiments with a regional climate model, Zhong and Hu (2007) suggested that summertime TC activities over the western North Pacific (WNP) have significant influences on the East Asian monsoon system, whereas the EASJ is one of the important components of the monsoon system. Northward movements of TCs transport atmospheric momentum and moisture content from the low latitudes to high latitudes and affect the atmospheric circulation in these regions

Corresponding author: Zhong Zhong, zhong_zhong@yeah.net

DOI: 10.1175/JCLI-D-16-0334.1

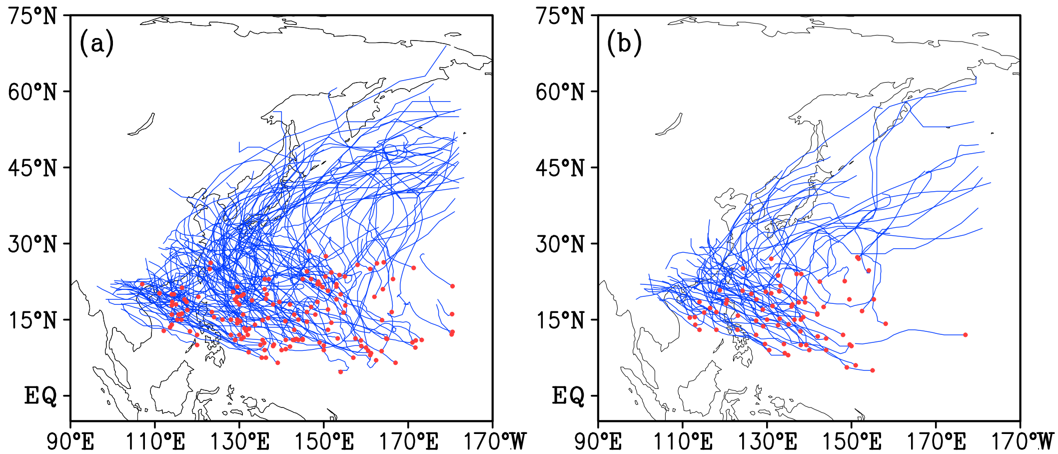

FIG. 1. TC tracks in (a) AY and (b) IY. The blue lines represent the complete TC tracks. The red dots represent the genesis locations of TCs.

(Jansen and Ferrari 2009; Ha et al. 2013). In the present study, the association of TC activity over the WNP with the summertime EASJ and possible mechanisms are investigated.

The data and methods used in this study are described in section 2. Results and conclusions are presented in sections 3 and 4, respectively.

2. Data and methods

The TC best-track dataset from the Regional Specialized Meteorological Center (RSMC) Tokyo Typhoon Center for the summer (June–August) of 1971–2010 is used in this study (JMA 2012). This dataset includes TC name, TC position in latitude and longitude, central pressure, and maximum sustained winds at 6-h intervals. Only those TCs reaching tropical storm intensity (maximum sustained wind speed $\geq 17.2\,\mathrm{m\,s^{-1}}$) are selected for this study. The three-dimensional atmospheric winds and temperature are extracted from the NCEP–NCAR reanalysis at a 6-h interval at 2.5° × 2.5° resolution (Kalnay et al. 1996).

In 1971–2010, the 10 years with the most TCs in summer [active years (AY)] over the WNP are 1971, 1972, 1978, 1981, 1989, 1992, 1994, 1997, 2002, and 2004, and the 10 years with the least TCs in summer [inactive years (IY)] are 1975, 1977, 1979, 1980, 1983, 1995, 1998, 2007, 2008, and 2010. Changes in atmospheric wind composite and meridional air temperature difference (MATD) composite in the AY and IY are investigated in this study to reveal the impact of TC activity on the EASJ.

In addition, the apparent heat source Q_1 is calculated for AY and IY according to the formula proposed by Yanai et al. (1973):

$$Q_1 = c_p\left[\frac{\partial T}{\partial t} + \mathbf{V}\cdot\nabla T + \left(\frac{p}{p_0}\right)^{\kappa}\omega\frac{\partial\theta}{\partial p}\right], \qquad (1)$$

where T is the air temperature, θ the potential temperature, $\mathbf{V}$ the horizontal wind vector, ω the vertical velocity, p the pressure, $p_0 = 1000\,\mathrm{hPa}$, $\kappa = R/c_p$, and R and c_p are the gas constant and the specific heat, respectively.

3. Results

Figure 1 displays the genesis locations and tracks of TCs for the AY and IY, respectively. Statistical results indicate that there were 153 TCs in the 10 active years. These TCs mainly formed over the tropical ocean within 110°–170°E; only 74 TCs formed in the 10 inactive years, which is less than half of the TCs formed in the 10 active years. Most of these TCs in the IY formed over the tropical ocean to the west of 150°E. After the formation of TCs, they tended to move northwestward and then turned northward, affecting mid- and high latitudes. In the 10 active years, 38 TCs made landfall in the continent of Asia, and 65 TCs reached north of 35°N. In contrast, only 20 TCs made landfall and 24 TCs reached north of 35°N in the 10 inactive years.

The summertime average zonal winds at 200 hPa for AY, IY, and the difference between AY and IY was shown in Fig. 2. In both AY and IY, the jet stream with winds greater than $20\,\mathrm{m\,s^{-1}}$ is located at the midlatitudes of Asia and the North Pacific between 35° and 50°N. Furthermore, two jet cores are most commonly found to the north of the Plateau (85°–95°E) and over the Kuroshio Extension to the east of Japan (140°–165°E), respectively (Zhang et al. 2006; Liao and Zhang 2013). However, the jet core locations are different between the AY and IY. Since the ridge line of the South Asian

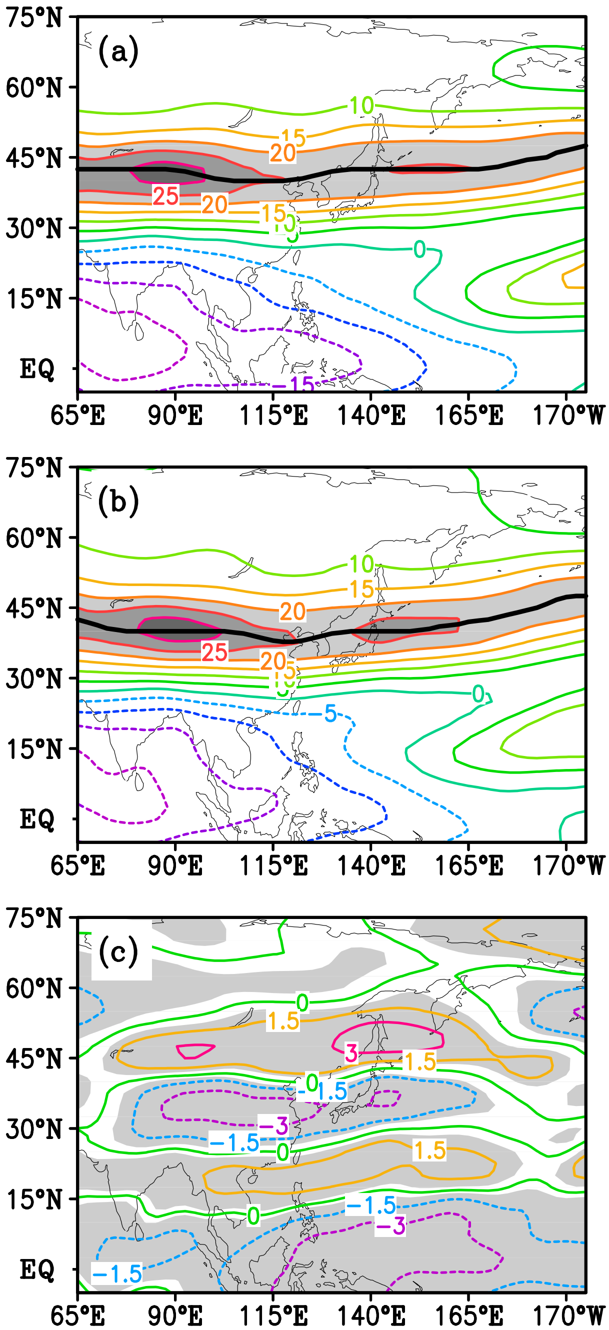

FIG. 2. Summertime average zonal winds (units: m s^{-1}) at 200 hPa for (a) AY, (b) IY, and (c) the difference between AY and IY, where the thick solid lines in (a) and (b) indicate the jet axis, zonal winds exceeding 20 m s^{-1} in (a) and (b) are shaded, and the difference at the 99% significant level is shaded in (c).

high is located at around 30°N in the summer (Zarrin et al. 2010; Ren et al. 2015), zonal westerlies prevail over regions to the north of 30°N while easterlies control the Asian–western Pacific region to the south of 30°N (Figs. 2a,b). Differences in the wind fields between AY and IY (Fig. 2c) are characterized by quasi-zonal westerlies to the north of the EASJ and quasi-zonal easterlies to the south of the EASJ that extends from the Tibetan Plateau to the Kuroshio Extension to the east of Japan. Such a pattern of difference in wind fields between AY and IY indicates that the axis of the EASJ shifts northward in AY. In addition, the quasi-zonal easterlies shown in the wind field difference between AY and IY over the Maritime Continent and the equatorial western Pacific suggest that the easterlies in the upper level above this region are stronger in AY than in IY. The quasi-zonal westerlies in the wind field difference between AY and IY cover the region from south of the EASJ to about 15°N, with the largest difference occurring in northern Luzon Island of the Philippines and the southern Bashi Channel. Thereby, the quasi-zonal easterlies and westerlies differences caused by the difference of the large-scale wind system between AY and IY alternate from the equator to high latitudes, exhibiting a meridional wave train–like flow difference between AY and IY.

In Fig. 2, it also shows that the maximum zonal wind speed at the jet core to the north of the Tibetan Plateau is around 32 m s^{-1} in both AY and IY, and the jet axis is located at about 42.5°N in AY and 40°N in IY. The maximum zonal wind speed along the jet core over the ocean to the east of Japan is 25 m s^{-1} in AY and 26 m s^{-1} in IY, and the jet core locates farther north in AY than in IY (Figs. 2a,b). The largest westerly (easterly) wind speed difference greater than 3 m s^{-1} between AY and IY is found to the north (south) of the jet core above the Tibetan Plateau and over the ocean area to the east of Japan. Areas of zonal wind speed difference larger than 1.5 m s^{-1} between AY and IY are mainly located within 75°–165°E (Fig. 2c). To further investigate the difference in EASJ between AY and IY, the zonal winds are averaged along 75°–165°E. The height–latitude cross sections of the zonally averaged zonal winds for AY, IY, and the difference between AY and IY are shown in Fig. 3.

Figure 3 shows that the jet axis is located at around 200 hPa in both AY and IY (Figs. 3a,b). To the north of 30°N, the westerlies gradually increase upward from the ground to 200 hPa and then gradually decrease above 200 hPa. The maximum zonal wind speeds of 26.5 and 27.5 m s^{-1} are located at 42.5° and 40°N in AY and IY, respectively (denoted by × in Figs. 3a,b). The boundary of the abnormal westerlies and easterlies of wind differences between AY and IY is located near 41°N, with the abnormal westerlies (easterlies) located to the north (south) of the boundary (Fig. 3c). To the south of 30°N, weak westerlies prevail in the lower troposphere and are gradually replaced by easterlies in upper levels in both

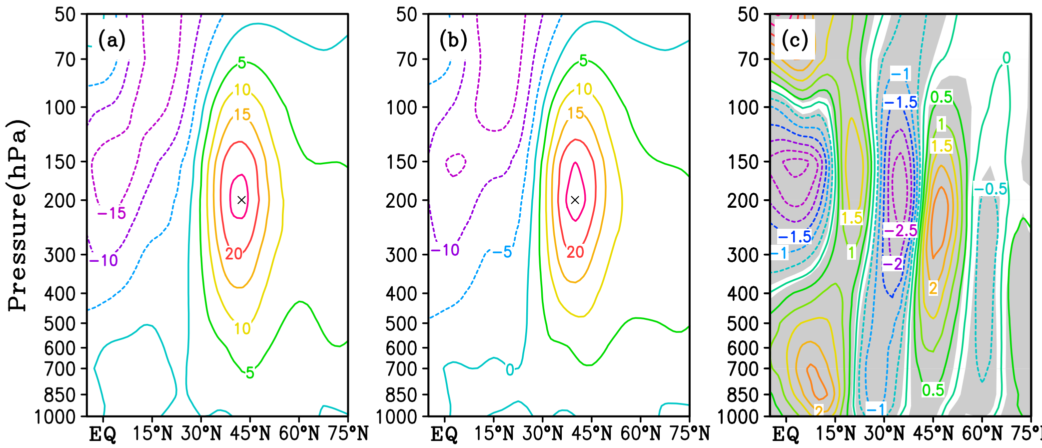

FIG. 3. Height–latitude cross sections of zonal wind averaged along 75°–165°E (units: m s^{-1}): (a) AY, (b) IY, and (c) difference between AY and IY. The × indicates the location of maximum zonal wind speed. In (c), the difference at the 99% level is shaded.

AY and IY, reaching the maximum at around 150 hPa. In addition, differences in wind fields between AY and IY display a quasi-barotropic structure to the north of 20°N and a baroclinic structure to the south of 20°N. Moreover, the easterlies and westerlies shown in the wind field difference (Figs. 2c and 3c) alternatively occur from the equator to high latitudes, and the top height of the wind field difference decreases as the latitude increases to the north of 40°N, which suggests that the effect of TC activity on zonal wind could reach upward to the stratosphere in lower and midlatitudes and gradually disappear in the troposphere in high latitudes. Note that the features shown in Fig. 3 are common for zonal mean winds along the jet axis for AY, IY, and their differences, respectively (figure not shown), indicating that the displacement of the EASJ as a whole is related to TC activities.

The thermal wind equation can be expressed as a linear relationship between the vertical shear of the zonal wind and the meridional air temperature gradient:

$$\frac{\partial u}{\partial p} = \frac{R}{fp}\left(\frac{\partial T}{\partial y}\right), \tag{2}$$

where u is the zonal wind and f is the Coriolis parameter. For the large-scale circulation systems, such as EASJ, the thermal wind adjustment is mainly toward the temperature field (Chen 1964). If the atmospheric temperature is low (high) in the north and high (low) in the south, westerly winds will increase (decrease) with height. Here we present MATD, defined by the difference of temperature between a grid point and its adjacent south point in the reanalysis data, as the meridional temperature gradient; if the MATD is negative (positive) below the EASJ, then the zonal winds will increase (decrease) with height.

Figure 4 presents the height–latitude cross section of the MATD difference between AY and IY averaged along 75°–165°E. It shows clearly that the distribution of MATD is obviously different in AY and in IY over the entire troposphere and the lower stratosphere. The larger MATD difference centers can reach up to 100 hPa in the tropics and subtropics; then, it lowers with latitude increasing as zonal wind difference in Fig. 3c. In addition, upward from the ground to 250 hPa, the MATD difference between AY and IY, which is associated with the EASJ intensity change, is negative (positive) to the north (south) of 41°N. Correspondingly, the abnormal

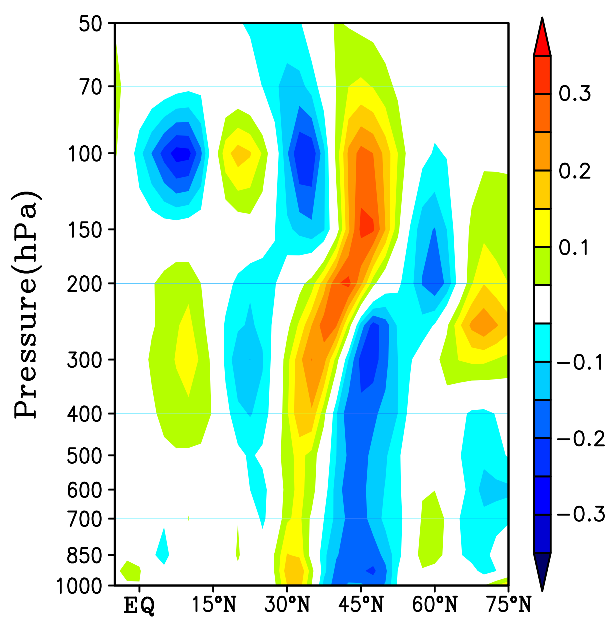

FIG. 4. The height–latitude cross section of difference in the MATD between AY and IY averaged along 75°–165°E (unit: K).

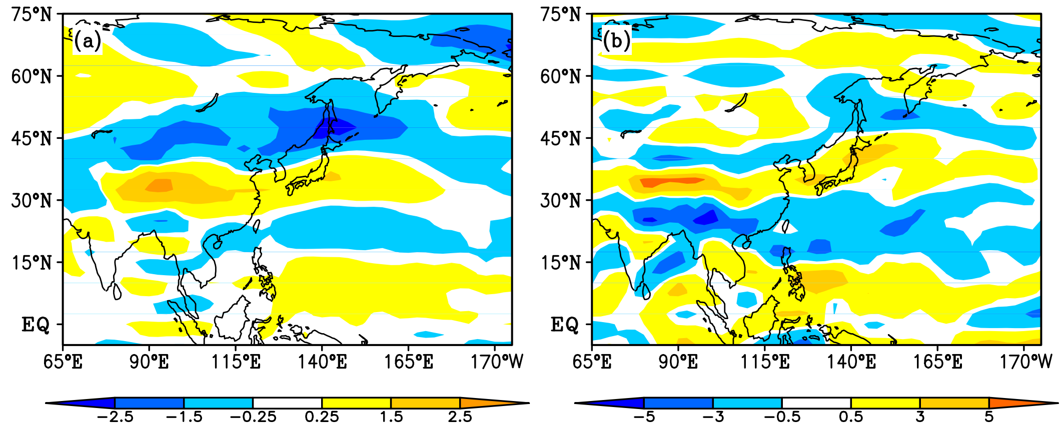

FIG. 5. Distributions of (a) MATD difference (unit: 10^{-1} K) and (b) MQD difference (unit: 10^{-5} K s^{-1}) averaged from the surface to 200 hPa between AY and IY.

westerly flow (easterly flow) of zonal wind difference between AY and IY increases with height (Fig. 3c). Above 250 hPa, the larger MATD difference center gradually tilts northward, leading to the out-of-phase MATD difference at 150 and at 250 hPa. As a result, the abnormal westerly flow shown in the zonal wind difference between AY and IY gradually disappears in upper levels above the EASJ axis. This means that the changes in large-scale MATD induced by frequent TC activities over the WNP may contribute to the overall poleward shift of the EASJ in AY. Note that the out-of-phase feature of MATD difference betweenthe upper troposphere and the lower stratosphere appears nearly in all latitudes where the impact of TCs could reach, suggesting that the TC activity could make the change of wind system in both the troposphere and stratosphere. In addition, positive and negative MATD difference occurs alternatively from the equator to high latitudes over the entire troposphere and the lower stratosphere, which implies that the TC activities could exhibit their influences through stationary Rossby waves (Kawamura and Ogasawara 2006).

To reveal the heating effect related to TC activities, as with the definition of MATD, the meridional Q_1 difference (MQD) was calculated with reanalysis data. Figure 5 displays the difference distribution of MATD and MQD between AY and IY averaged from the surface to 200 hPa. It shows clearly that the meridionally positive and negative MATD differences occur alternatively over East Asia and the Pacific, corresponding to the meridional wave train–like zonal wind difference between AY and IY (Figs. 2c and 3c). In addition, it also shows that the largest MATD differences are found to occur over the Tibetan Plateau and the adjacent ocean to the east of Japan (Fig. 5a), resulting in the largest zonal wind differences to the south and north of the EASJ axis (Fig. 2c). Since the air temperature is mainly restricted by the heating, from Fig. 5b, one can see that the MQD difference also shows a wave train–like meridional variation as MATD difference does, and the phase of MQD difference matches that of the MATD difference. Therefore, to a certain extent, the MQD difference between AY and IY is responsible for the MATD difference, which leads to the variation of the EASJ, and the frequent TC activities in AY would contribute to the poleward shift of EASJ axis.

To further explain the contribution of TC-associated heating to regional circulation, the Q_1 within 800 km around the center of each TC along its track shown in Fig. 1 is accumulated. The seasonal mean MQD difference based on the Q_1 composite averaged from the surface to 200 hPa between AY and IY is shown in Fig. 6. It can be seen that the influence scope of TCs' latent heating is limited in the areas around their tracks; the difference pattern is similar to that between AY and IY (Fig. 5b) to the east of the Indochina Peninsula, basically with about half of the magnitude, suggesting that the TC latent heat release may be a direct driving mechanism that makes changes in the air temperature gradient and leads to the meridional displacement of the EASJ in its east portion, even though the feedback effects of TCs on the general circulation were not considered.

Besides the direct impact of TCs on the east portion of EASJ as well as the feedback effect on general circulation, how does the EASJ exhibit its meridional displacement as a whole in AY compared with IY? To answer this question, as shown in Fig. 7, the distribution of correlation coefficient between summertime TC number over WNP and seasonal mean wind speed at 850 hPa in the Asian monsoon area during 1971–2010 was drawn. It clearly shows that the positive correlation between the summertime TC number and the Asian

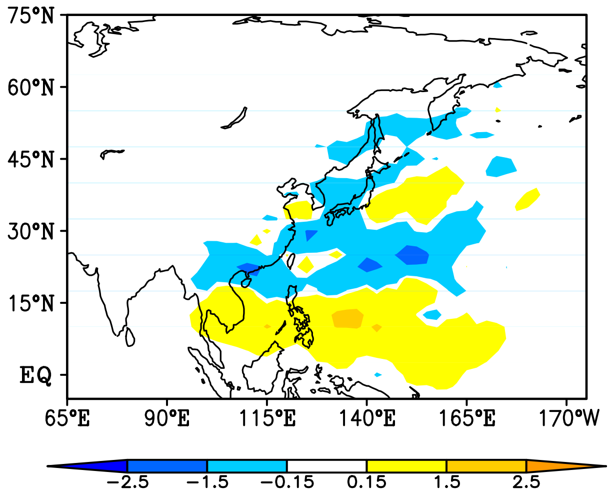

FIG. 6. Distributions of seasonal mean MQD difference (unit: 10^{-5} K s^{-1}) based on the apparent heat source Q_1 composite along TC tracks averaged from the surface to 200 hPa between AY and IY.

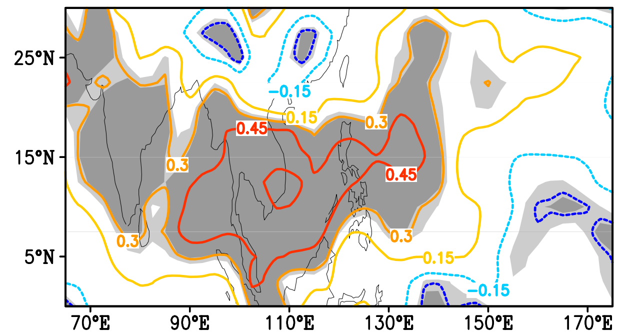

FIG. 7. Correlation coefficients between summertime WNP TC number and seasonal mean wind speed at 850 hPa in the Asian monsoon area during the period 1971–2010. Light (dark) shading indicates areas where the correlation coefficient is statistically significant at the 90% (95%) confidence level by the two-tailed Student's t test.

summer monsoon intensity is significant, which implies that the stronger the summer monsoon, the greater the TC number, since the intensive WNP monsoon trough with stronger positive relative vorticity would be favorable for TC formation (Ha et al. 2013). Meanwhile, the intensive monsoon flow over the Indian subcontinent and Bay of Bengal will strengthen the convection activities there (Mohanty et al. 2002; Akter and Tsuboki 2014) associated with larger diabatic heating (Guan and Yamagata 2003), which will lead to an anomalous meridional circulation and make the circulation system over the Tibetan Plateau move northward. Therefore, the integral poleward shift of EASJ in summer could be attributed to the enhanced Asian summer monsoon to some extent, and the TCs over WNP do their direct contribution in the east portion of the jet through meridional air temperature gradient change caused by the latent heat release.

4. Conclusions and discussion

The NCEP–NCAR reanalysis and the RSMC TC best-track dataset have been exploited in this study to investigate the association of TC activity over the WNP with the summertime meridional displacement of EASJ. It indicates that, accompanying the frequent TC activities in AY over WNP, the EASJ, from the northern flank of the Tibetan Plateau to the Kuroshio Extension east to Japan (75°–165°E), exhibits an integral poleward shift in summer. It is also found that the latent heat release by TCs over the WNP contributes directly to the northward displacement of the east portion of EASJ through altering the meridional air temperature gradient in troposphere and lower stratosphere. Whereas for the northward shift of the west portion of EASJ in AY, it could be attributed to the intensive South Asian summer monsoon. On the contrary, the EASJ is located abnormally southward in IY and has stronger than normal intensity.

It was pointed out that the summertime EASJ meridional displacement is closely related to the evolution of the Pacific–Japan (PJ) teleconnection pattern at the interannual time scale (Lu 2004; Lu and Lin 2009; Sun et al. 2010; Zhong et al. 2015), and corresponding to enhanced atmospheric convection over the Philippine Sea, the EASJ tends to exhibit a slight poleward displacement in July and August (Lu 2004). What is more, it was also found that the PJ teleconnection could be stimulated by stationary Rossby waves induced by typhoons over the WNP (Kawamura and Ogasawara 2006) and vice versa (Choi et al. 2010; Kubota et al. 2016). Therefore, the PJ teleconnection pattern could be considered as a bridge linking the EASJ and TC activities over WNP. On the one hand, TCs over the WNP can make their influence on EASJ through stimulating PJ connection pattern, but on the other hand TCs could directly produce the meridional displacement of EASJ through the diabatic heating effect on the meridional temperature gradient in the troposphere and stratosphere, at least for the east portion of EASJ.

Acknowledgments. We thank anonymous reviewers for their valuable comments and suggestions, which were very important in improving the quality of this work. The authors contributed equally to this study and share the first authorship. This work is sponsored by the National Natural Science Foundation of China (41430426 and

41130963) and the R&D Special Fund for Public Welfare Industry (Meteorology) (GYHY201306025).

REFERENCES

Akter, N., and K. Tsuboki, 2014: Role of synoptic-scale forcing in cyclogenesis over the Bay of Bengal. *Climate Dyn.*, **43**, 2651–2662, doi:10.1007/s00382-014-2077-9.

Carillo, A., P. M. Ruti, and A. Navarra, 2000: Storm tracks and zonal mean flow variability: A comparison between observed and simulated data. *Climate Dyn.*, **16**, 219–228, doi:10.1007/s003820050015.

Chen, C. S., 1964: On the formation of the thermal wind equilibrium state in a simple baroclinic atmosphere. *Sci. China*, **13**, 279–289.

——, and K. E. Trenberth, 1988: Forced planetary waves in the Northern Hemisphere winter: Wave-coupled orographic and thermal forcing. *J. Atmos. Sci.*, **45**, 682–704, doi:10.1175/1520-0469(1988)045<0682:FPWITN>2.0.CO;2.

Chen, X., Z. Zhong, W. Lu, and X. Tang, 2014: Distribution characteristics and the possible mechanism of mesoscale disturbance in the subtropical upper-level jet stream over East Asia (in Chinese). *Chin. J. Geophys.*, **57**, 2455–2464.

Choi, K.-S., C.-C. Wu, and E.-J. Cha, 2010: Change of tropical cyclone activity by Pacific-Japan teleconnection pattern in the western North Pacific. *J. Geophys. Res.*, **115**, D19114, doi:10.1029/2010JD013866.

Guan, Z., and T. Yamagata, 2003: The unusual summer of 1994 in East Asia: IOD teleconnection. *Geophys. Res. Lett.*, **30**, 1544–1547, doi:10.1029/2002GL016831.

Ha, Y., Z. Zhong, Y. Hu, and X. Yang, 2013: Influences of ENSO on western North Pacific tropical cyclone kinetic energy and its meridional transport. *J. Climate*, **26**, 322–332, doi:10.1175/JCLI-D-11-00543.1.

Hou, A. Y., 1998: Hadley circulation as a modulator of the extratropical climate. *J. Atmos. Sci.*, **55**, 2437–2457, doi:10.1175/1520-0469(1998)055<2437:HCAAMO>2.0.CO;2.

Jansen, M., and R. Ferrari, 2009: Impact of the latitudinal distribution of tropical cyclones on the ocean heat transport. *Geophys. Res. Lett.*, **36**, L06604, doi:10.1029/2008GL036796.

JMA, 2012: RSMC Tokyo—Typhoon Center, best track data. JMA Regional Specialized Meteorological Center, accessed 1 July 2013. [Available online at http://www.jma.go.jp/jma/jma-eng/jma-center/rsmc-hp-pub-eg/RSMC_HP.htm.]

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–472, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kawamura, R., and T. Ogasawara, 2006: On the role of typhoons in generating PJ teleconnection patterns over the western North Pacific in late summer. *SOLA*, **2**, 37–40.

Kubota, H., Y. Kosaka, and S.-P. Xie, 2016: A 117-year long index of the Pacific-Japan pattern with application to interdecadal variability. *Int. J. Climatol.*, **36**, 1575–1589, doi:10.1002/joc.4441.

Liao, Z., and Y. Zhang, 2013: Concurrent variation between the East Asian subtropical jet and polar front jet during persistent snowstorm period in 2008 winter over southern China. *J. Geophys. Res. Atmos.*, **118**, 6360–6373, doi:10.1002/jgrd.50558.

Lu, R., 2004: Associations among the components of the East Asian summer monsoon system in the meridional direction. *J. Meteor. Soc. Japan*, **82**, 155–165, doi:10.2151/jmsj.82.155.

——, and Z. Lin, 2009: Role of subtropical precipitation anomalies in maintaining the summertime meridional teleconnection over the western Pacific and East Asia. *J. Climate*, **22**, 2058–2072, doi:10.1175/2008JCLI2444.1.

——, H. Ye, and J. G. Jhun, 2011: Weakening of interannual variability in the summer East Asian upper-tropospheric westerly jet since the mid-1990s. *Adv. Atmos. Sci.*, **28**, 1246–1258, doi:10.1007/s00376-011-0222-5.

Mohanty, U., R. Bhatla, P. Raju, O. Madan, and A. Sarkar, 2002: Meteorological fields variability over the Indian seas in pre and summer months during extreme monsoon seasons. *J. Earth Syst. Sci.*, **111**, 365–378, doi:10.1007/BF02701981.

Ren, X., Y. Zhang, and Y. Xiang, 2008: Connections between wintertime jet stream variability, oceanic surface heating and transient eddy activity in the North Pacific. *J. Geophys. Res.*, **113**, D21119, doi:10.1029/2007JD009464.

——, D. Yang, and X. Q. Yang, 2015: Characteristics and mechanism of the subseasonal eastward extension of the South Asian high. *J. Climate*, **28**, 6799–6822, doi:10.1175/JCLI-D-14-00682.1.

Sampe, T., and S. P. Xie, 2010: Large-scale dynamics of the meiyu-baiu rainband: Environmental forcing by the westerly jet. *J. Climate*, **23**, 113–133, doi:10.1175/2009JCLI3128.1.

Schiemann, R., D. Lüthix, and C. Schär, 2009: Seasonality and interannual variability of the westerly jet in the Tibetan Plateau region. *J. Climate*, **22**, 2940–2957, doi:10.1175/2008JCLI2625.1.

Sun, X., R. J. Greatbatch, W. Park, and M. Latif, 2010: Two major modes of variability of the East Asian summer monsoon. *Quart. J. Roy. Meteor. Soc.*, **136**, 829–841, doi:10.1002/qj.635.

Yanai, M., S. Esbensen, and J.-H. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J. Atmos. Sci.*, **30**, 611–627, doi:10.1175/1520-0469(1973)030<0611:DOBPOT>2.0.CO;2.

Zarrin, A., H. Ghaemi, M. Azadi, and M. Farajzadeh, 2010: The spatial pattern of summertime subtropical anticyclones over Asia and Africa: A climatological review. *Int. J. Climatol.*, **30**, 159–173, doi:10.1002/joc.1879.

Zhang, Y., X. Kuang, W. Guo, and T. Zhou, 2006: Seasonal evolution of the upper-tropospheric westerly jet core over East Asia. *Geophys. Res. Lett.*, **33**, L11708, doi:10.1029/2006GL026377.

Zhong, Z., and Y. Hu, 2007: Impact of tropical cyclones on the regional climate: An East Asian summer monsoon case. *Atmos. Sci. Lett.*, **8**, 93–99, doi:10.1002/asl.158.

——, H. Yuan, J. Li, and H. Fan, 2010: Characteristics of the mesoscale perturbations and momentum transportation associated with an intensification process of upper level jet (in Chinese). *J. Meteor. Sci.*, **30**, 639–645.

——, X. Tang, W. Lu, and Z. Chen, 2015: The relationship of summertime upper level jet over East Asia and the Pacific-Japan teleconnection. *J. Meteor. Sci.*, **35**, 672–683.