Upper-Tropospheric Forcing on Late July Monsoon Transition in East Asia and the Western North Pacific

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ABSTRACT

By investigating the large-scale circulation in the upper troposphere, it is demonstrated that the rapid late July summer monsoon transition in the East Asia and western North Pacific (EA-WNP) is associated with a weakened westerly at the exit of the East Asian jet stream (EAJS). Even in a normally stable atmosphere under the influence of the North Pacific (NP) high in late July, convection rapidly develops over the mid-oceanic region of the western NP (15°–25°N, 150°–170°E). Prior to the rapid transition, the EAJS weakens and shifts northward, which induces a series of changes in downstream regions; the northeastern stretch of the Asian high weakens, upper-tropospheric divergence in the region southwest of the mid-NP trough increases, and convection is enhanced. At the monsoon transition, upper-level high potential vorticity intrudes southward and westward, convection expand from the mid NP westward to cover the entire subtropical western NP, the lower-tropospheric monsoon trough deepens, surface southwesterly flow strengthens, and the western stretch of the NP high shifts northward; 10° latitude to the south of Japan. This series of changes indicates that the EA-WNP late July monsoon transition is initiated from changes in the upper-tropospheric circulation via the weakening of the EAJS south of 45°N. The weakening of the EAJS south of 45°N is related to a reduced gradient of the geopotential height on the northern flank of the Asian high, which is related to the massive inland heating and weakening of the South Asian monsoon circulation. The exact timing of the monsoon onset might be tied to the hypothesized “Silk Road pattern” and/or a strong weakening of the South Asian monsoon circulation.

1. Introduction

The Asian–Pacific summer monsoon experiences step-wise and rapid changes (Matsumoto 1992; Ueda et al. 1995; Hsu et al. 1999; Wu and Wang 2001; LinHo and Wang 2002; Li and Wang 2005). The differences in land–sea distribution (Indian subcontinent, Bay of Bengal, Indochina Peninsula, and South China Sea) lead to large discrepancies of the thermal regime and are responsible for the differences of monsoons in those different regions. In East Asia (EA) and the western North Pacific (WNP, EA-WNP), the summer monsoon proceeds with various stages of surges and breaks associated with the evolution of the North Pacific (NP) high pressure system (NP high), in particular, the western stretch of the high (Lu 2001; Hattori et al. 2005; Kim et al. 2009). Evolutions of the Asian–Pacific summer monsoon can generally be classified into three phases (Chou et al. 2011). The first transition is the onset of active convections and heavy rains in Indochina in the middle of May, which corresponds to the inception of the mei-yu season in southern China and Taiwan (Chen et al. 2004). The second transition is the onset of Indian summer monsoon in early to middle of June, when the solar heating of land in northern India and the Tibetan Plateau strengthens (Bhat et al. 1996). The third transition occurs in the EA-WNP region in late July when the previously stable subtropical western NP (SWNP) becomes convective and cloudy and the baiu season in Japan ends (Ueda et al. 1995).

The NP high evolves following the seasonal procession of solar heating of the Asian continent. The location of the western stretch of the high has a dominating control of the EA-WNP summer monsoon. Climatologically, the western stretch of the NP high migrates northward from the Philippines in May through Taiwan in mid July and rapidly shifts northward by ~10° latitude to the south of Japan in late July. Corresponding to the northward shift of the high is the deepening of the monsoon trough over the SWNP. The change in large-scale circulation is

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associated with the monsoon onset in the SWNP and the termination of the early summer mei-yu–baiu rainy season (Ueda et al. 1995; Suzuki and Hoskins 2009; Wu et al. 2009). The monsoon precipitation is redistributed in the EA-WNP due to the rapid monsoon phase transition (Chen et al. 2004; Li and Wang 2005; Ding 2007; Chou et al. 2009).

There are a number of studies on the cause of the abrupt late July monsoon transition. Some addressed the forcing from the tropics and subtropics, yet some demonstrated the forcing from midlatitudes. As for the tropical–subtropical forcing, the changes in heat and surface moisture fluxes in the boundary layer were investigated by relating them to tropical cyclones (Holland 1995; Ueda and Yasunari 1996), the propagation of an intraseasonal oscillation (ISO; Hsu and Weng 2001; Wu and Wang 2001), and the air–sea interaction due to high sea surface temperature (SST; Ueda et al. 1995; Wu 2002a).

Ueda et al. (2009) recently evaluated the impact of SST on the monsoon transition using a general circulation model simulations. The GCM simulations showed a weak impact of SST on the SWNP onset. They suggested that gradual tropospheric moistening triggers a threshold transition. The strong tropospheric moistening preceding the late July monsoon transition appeared in the CloudSat Cloud-Proﬁling Radar (CPR) observations (Wu et al. 2011).

Apart from the tropical and subtropical forcing, the inﬂuence of midlatitude circulation variability on the monsoon transition was also investigated, which includes upper-tropospheric disturbances in the mid North Paciﬁc Ocean (Sato et al. 2005; Lu et al. 2007; Wu et al. 2009); variability of the upper-tropospheric westerlies (Yang et al. 2004; Zhang et al. 2006; Lin and Lu 2008); the atmospheric thermal contrast between the midlatitude Asian continent and its eastern adjacent ocean (Wu 2002b; Cheng et al. 2008); the equivalent barotropic Rossby wave train that develops downstream across Eurasia (Enomoto et al. 2003; Suzuki and Hoskins 2009); the stratosphere–troposphere interaction and baroclinic energy conversion over Mongolia (Inoue and Takahashi 2009); and the Okhotsk high (Nakamura and Fukamachi 2004; Sato and Takahashi 2007) and the Bonin high (Enomoto et al. 2003; Ha and Lee 2007). Murakami and Matsumoto (1994) suggested that the baiu rainy season has a weak connection to the tropical convective system, for example, the intertropical convergence zone.

Wu et al. (2009) suggested that the monsoon transition was forced by the upper-tropospheric disturbances in the midoceanic trough. In this study we further provide information of the atmospheric conditions prior to and during the abrupt EA-WNP monsoon transition by detailed analysis of the corresponding changes in the large-scale circulation. Identification of the abrupt monsoon transition is given in section 2. In section 3, we investigated the convection evolution over the mid North Paciﬁc Ocean and its relation to the large-scale circulation in the upper troposphere. In section 4, we investigated changes of the monsoon circulation in South and Southeast Asia and its relation to the Asian high variation during the late July EA-WNP monsoon transition. Conclusions are given in section 5.

2. Data and identiﬁcation of abrupt monsoon transition

Data used in this study include (i) the outgoing longwave radiation (OLR) at the top of the atmosphere from the National Oceanic and Atmospheric Administration (Liebmann and Smith 1996). It has a temporal resolution of 1 day and a spatial resolution of 2.5° latitude–longitude; (ii) geopotential height, wind, and atmospheric temperature from the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996). It has a temporal resolution of 6 h and a spatial resolution of 2.5° latitude–longitude.

We investigated the late July monsoon transition not only in the SWNP but also in the broader EA-WNP. The abrupt monsoon transition was identiﬁed in two steps using OLR data. 1) We ﬁrst applied an empirical orthogonal function (EOF) analysis to the OLR data from 1974 to 2010 (a total of 36 years with no data in 1978) to evaluate the abrupt transition in the EA-WNP domain (0°–40°N, 120°E–180°E). In computing the EOFs, we subtracted the seasonal mean and applied a 20–80-day bandpass ﬁlter to the OLR data to remove interannual variations and synoptic signals. Results of the EOF analysis show that the leading EOF (EOF1) attains a peak magnitude in the SWNP and that the associated principle component (PC1) has a large change in late July and early August in 16 of the 36 years. 2) To identify the date of the abrupt monsoon transition (onset date) in those 16 years with a large change in PC1, we compared 20-day mean values of the OLR in the SWNP, deﬁned in this study for the region (15°–25°N, 130°–150°E), before and after a given date in the period when the PC1 has a large change. When the 20-day mean OLR values before and after the given date have the largest decrease among all dates in the period when the PC1 has a large change, this is then identiﬁed as the onset date. The use of the 20-day mean OLR in the identiﬁcation of the onset date is to make certain that variations of OLR before and after the onset date are related to the monsoon transition and not just short-term weather variations.

The results using data from the 16 summers are shown in Fig. 1, which include the OLR EOF1 (Fig. 1a) and...
PC1 (Fig. 1b, solid lines) and the area-mean OLR in the SWNP (Fig. 1b, dashed lines). It is clear that the rapid decrease of OLR in the SWNP is highly correlated with the OLR PC1. The correlation coefficient between the two is $0.7$ with a 99% confidence level. It has been well recognized that monsoon onset in the SWNP coincides with the termination of the baiu rains. It can be seen in Fig. 1 that a decrease of OLR in the SWNP corresponds to an increase of OLR in the region near Japan. These changes are two essential components of the monsoon transition in the EA-WNP. Suzuki and Hoskins (2009) defined the end day of the baiu season by the $850$-hPa equivalent potential temperature ($\theta_e$) when it exceeded $330$ K at $130^\circ$–$140^\circ$E.

3. Convection evolution and the changing circulation in the upper troposphere

Figure 3 shows pentad-mean evolution of circulation and geopotential heights at $850$ hPa (left column) and convection (right column) during and prior to the rapid monsoon transition. Three pentads prior to the rapid transition (Fig. 3a), the subtropical midoceanic region $15^\circ$–$25^\circ$N, $150^\circ$–$170^\circ$E (indicated by the boxes) is clear under the influence of the NP high. One pentad later, the midoceanic region becomes cloudy (low OLR) even though the atmosphere is still stable under the influence of the NP high pressure (Wu et al. 2009). One pentad prior to the onset (Fig. 3c), the OLR continues to decrease in this region, while the ridge of the NP high stays nearly unchanged. This is in the late period of the second phase of the Asian–Pacific summer monsoon, which corresponds to the withdrawal of the East Asian summer monsoon (Chen et al. 2004; LinHo and Wang 2002; Ueda et al. 2009), and the SWNP and adjacent areas are under the control of the subtropical high pressure. Deep convection is suppressed regardless of high SST; high SST does not incite convections, rather it is a result of a stable atmosphere with strong solar heating. At the

more than $30$ W m$^{-2}$ in the SWNP, and the location of the $330$-K $\theta_e$ rapidly moves northward through $39^\circ$–$40^\circ$N. This is an indication that the cool air from the west is replaced by warm and moist air transported from the tropics to Japan right after the monsoon onset. Thus, identification of the late July monsoon transition in the EA-WNP by either the OLR in the SWNP or the $330$-K $\theta_e$ at $850$ hPa is consistent.
monsoon transition (Fig. 3d), cloudy regions expand westward to cover the entire subtropical western NP, the monsoon trough rapidly deepens, and the western stretch of the NP high ridge shifts northward to the south of Japan by \( \sim 10^\circ \) latitude.

Figure 4 further shows changes of the 850-hPa circulation and OLR. One pentad prior to the monsoon transition (upper panels), the geopotential height increases significantly in the midlatitude EA-WNP and decreases slightly in the SWNP. Corresponding to these changes is an anomalous anticyclonic circulation in the middle latitudes and an anomalous cyclonic circulation in the SWNP. At the monsoon transition (lower panels), the geopotential height continues to increase, but only slightly, in the midlatitudes and decreases over a broad region covering the entire subtropical and tropical regions from South Asia to the EA-WNP. There is a large-scale anomalous cyclonic circulation with the center located near the SWNP. The anomalous cyclonic circulation indicates strengthening of the westerly wind in the Indian Ocean and the South China Sea after the monsoon transition. Prior to the monsoon transition (upper-right panel), the OLR decreases in the SWNP and the adjacent oceanic regions, but increases in the
baiu/Changma region (East China Sea, South Korea, and southern Japan). It continues to decrease in the SWNP and South China Sea and increases in the baiu/Changma region during the monsoon transition (lower-right panel). It is noticed that convection in the Maritime Continent and the adjacent oceans is strong and the OLR is already low prior to the monsoon transition. Hence, the OLR change in these regions is small during the monsoon transition. The results shown in Fig. 4 are similar to that of Ueda et al. (1995) who compared the abrupt change of convection between the periods 15–24 July and 25 July–3 August in 1980–89 over the western NP.

Wu et al. (2009) demonstrated that, prior to the late July monsoon onset, convection in the subtropical mid-oceanic region moves westward to the SWNP. To provide further information on this west-moving system, we show in Fig. 5 the lagged correlation coefficient between the OLR in the SWNP and individual 2.5° grid boxes. The lagged correlation coefficient was derived by regression of the OLR of each grid box against the SWNP OLR in seven pentads centered at the onset. Thus, there is a pair of 112 OLR (seven pentads × 16 yr) involved in the calculation of the correlation coefficient at each 2.5° grid box. Since by definition the SWNP at the onset has the largest cloud amount, that is, lowest OLR, the maximum correlation of >0.3 shown in the top panel of Fig. 5 indicates that the location of a large amount of clouds is located to the east and northeast of the SWNP three pentads prior to the onset. The cloudy area continuously expands and moves westward. It expands to cover the South China Sea and Philippine Sea one pentad after the onset (bottom panel) when the monsoon trough rapidly deepens. The deepening of the monsoon trough can be seen in the left panel of Fig. 3d.

Ueda et al. (1995) demonstrated that the abrupt northward shift of convection in the western NP coincided with the demise of the baiu season. Enomoto et al. (2003) suggested that the demise of baiu might be caused by the development of the Bonin high. They hypothesized that development of the Bonin high was a result of the propagation of stationary Rossby waves along the midlatitude upper-level Asian jet (the Silk Road pattern), and the location of the Bonin high was affected by the intensity of the East Asian jet stream (EAJS). Sato et al. (2005) investigated the southward intrusion of the midlatitude upper cold low (UCL) and the formation of the Marcus convergence zone (MCZ). They found that southward intrusion of high potential vorticity (PV) associated with the UCL contributed to the midoceanic convection in the tropical and subtropical western NP. It is noticed that the MCZ is a subtropical convergence zone extending northeastward from the warm pool east of the Philippines, passing through Marcus Island (~24°N, 153°E) and farther to the northeast of the island. This convergence zone coincides with the low OLR region in the subtropical midocean shown in the box of Fig. 3. Lin and Lu (2008) demonstrated that the northward jump of the EAJS is related to the withdrawal of baiu. Conclusions

Fig. 4. As in Fig. 3 but for the differences between consecutive pentads. Dots indicate regions where the differences are confident at the 90% level.
from these studies suggest that the monsoon transition might well be connected to the EAJS, the Bonin high, and the UCL.

To further support the suggestion that the northward shift of the NP high prior to the monsoon transition is related to changes in the upper tropospheric circulation, we show in Fig. 6 the 200-hPa streamlines, geopotential height (left column), and wind speed (right column). In the upper troposphere, the Asian high is the most prominent circulation system. To the east of the Asian high is an oceanic trough. In the northern flank of the high is the westerly jet stream, and in the southern flank of the high the easterly jet is strong over the Indian Ocean. Two pentads prior to the monsoon transition (Fig. 6a), we can see a region of wind divergence with low wind speed between the Asian high and the oceanic trough. This region coincides with the region of low OLR shown by the box in Fig. 3b. It is also roughly the MCZ investigated by Sato et al. (2005). One pentad later, the 200-hPa geopotential height decreases near 20°–30°N from India eastward to the mid-NP Ocean (Fig. 6b). In contrast, the geopotential height increases north of ~30°N and in the NP Ocean south of ~20°N. It is essentially a weakening of the Asian high along the ridge and a strengthening in the northern and southern flanks of the ridge, causing the meridional gradient of the geopotential height to decrease (increase) south (north) of ~45°N, where the increase of the geopotential height attains a maximum. Corresponding to the height change is a strong anomalous anticyclonic circulation to the north of the ridge. The EAJS weakens and shifts northward.

During the monsoon transition the Asian high continues to weaken (Fig. 6c) and the EAJS weakens and shifts northward. There is a strong anomalous anticyclonic circulation centered over the Maritime Continent. The northeasterly flow over the Maritime Continent and southern Philippines is greatly enhanced. Since the low-level flow in these regions is southwesterly, the change in the upper troposphere implies an enhanced vertical wind shear, suppressed downstream deep convection, and a large increase of OLR over the eastern Indian Ocean. The suppressed convection was investigated in the region under the summertime tropical easterly jet stream in South Asia (Sathiyamoorthy et al. 2004) and the western Pacific warm pool (Lin and Mapes 2004). Their results showed that the strong wind shear swept the cloud tops and was unfavorable for cloud growth above ~300 hPa.

To further investigate the development of convection preceding the abrupt monsoon transition over the subtropical mid-NP Ocean, we show in Fig. 7 the vertical velocity and PV along the 15°–25°N latitude band (left column) and 155°–165°E longitude zone (right column). Two pentads prior to the rapid monsoon transition

Fig. 5. The lagged correlation coefficient between the OLR in the SWNP and 2.5° grid boxes. Negative lag denotes the former lags the latter. The data of 16 onset years with a pentad-mean temporal resolution is used. Regions with a 99% confidence level are shaded.
(Fig. 7a), ascending air over the ocean occurs mainly around 20°N between 150°E and 180. It is clear that the location of ascending air coincides with that of the high PV center in the middle and upper troposphere. One pentad later (Fig. 7b), the tropospheric PV and ascending motion strengthen and expand westward. During the monsoon transition (Fig. 7c), the ascending motion continues to strengthen and expands rapidly westward to 120°E.

It has been proposed that intrusion of the upper-level high vorticity destabilizes the air below and promotes convection (Dixon et al. 2003; Hoskins et al. 1985; Kiladis 1998; Thorpe 1985). Funatsu and Waugh (2008) examined the connection between the upper-level PV intrusion and convection by analysis of NCEP–National Center for Atmospheric Research (NCAR) reanalyses and satellite-observed OLR, as well as mesoscale model simulations. Consistent with theoretical expectations, they found that the upper-level PV initiates and supports convection by destabilizing the lower troposphere and causing upward motion downstream of the intrusive high PV. The results discussed in the previous paragraph are consistent with the hypothesis that convection in the subtropical western NP occurs downstream of the intrusive PV of the upper troposphere.

In Fig. 8 we summarize changes of the upper-tropospheric circulation in the western NP prior to and during the monsoon transition. The EAJS east of Japan at 35°–45°N, 140°E–180° (marked U in the top panel) continuously decreases from 27 m s⁻¹ two weeks before the transition to 13 m s⁻¹ during the transition (middle panel). The weakened jet stream is related to the reduced gradient of geopotential height that can be seen in Fig. 6. Associated with this weakened jet south of 45°N, the eastern extension of the Asian high northwest of the

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**Fig. 6.** (left) The 200-hPa streamlines and geopotential height (m, shading) and (right) wind speed (m s⁻¹, shading). (a) The pentad-mean (day −6 to −10); (b),(c) differences between two pentads are shown. The results are 16-yr composites. Dots indicate regions where the differences are confident at the 90% level.
oceanic trough at 25°–35°N, 160°E–180° (marked PV in the upper panel) also undergoes changes. The solid curve in the middle panel of the figure shows that the PV at the isentropic surface of 350 K potential temperature increases rapidly two weeks prior to the transition and reaches a maximum at transition. The increase in vorticity signifies a weakened anticyclonic circulation at the exit of the jet stream east of Japan. This weakened anticyclonic cell induces an enhancement of divergence in the downstream region at 15°–25°N, 150°–170°E (marked

![Vertical cross sections of the pentad-mean vertical velocity (unit 0.01 Pa s⁻¹, shading) and the potential vorticity (contours) averaged over (left) 15°–25°N and (right) 155°–165°E. The PV contours are plotted between 0.35 and 0.50 PVU (=10⁻⁶ m² s⁻¹ K kg⁻¹); results are 16-yr composites for the vertical velocity and PV.](image-url)
DIV) that separates the Asian high to the west and the oceanic trough to the east and is downstream of the intrusive high PV. Consequently, convections in the DIV and the SWNP (marked OLR) strengthen and the OLR decreases. These changes are clearly shown in the lower panel of the figure. It is noticed that, during the monsoon transition, changes in DIV and OLR lag that of U and PV by ~5 days.

It should be noticed that during the two pentads prior to monsoon onset, the midoceanic region becomes more convective with decreasing OLR due to the intrusion of high PV, while it is still under the general influence of the subtropical high pressure. Only after the monsoon onset, the intrusion of high PV ceases, the western stretch of the NP high ridge shifts rapidly northward, and the subtropical western NP is within the domain of tropical influence (Fig. 3).

Using data from atmospheric GCM simulations and satellite observations, Ueda et al. (2009) suggested that slow tropospheric moistening over the subtropical regions of the western NP might trigger a threshold transition in mid-July. Wu et al. (2011) also demonstrated a large tropospheric moistening preceding the late July monsoon transition, using CloudSat data. Prior to the monsoon transition, the atmosphere above the boundary layer in the subtropical western NP is relatively dry. Extensive deep convection would be inhibited without an adequate supply of moisture. The transition period of ~2 pentads prior to the monsoon onset, as shown in Fig. 8, is the time required for the atmosphere to slowly build up the moisture to sustain enhanced convection. Figure 8 also shows that there are two or three fluctuations 2–4 pentads after the onset. They are much weaker than that during the monsoon transition. These fluctuations exhibit a simultaneous change in PV, U, DIV, and OLR, indicating a quick response of convection in the subtropical western NP to midlatitude forcing when there is adequate supply of moisture.

The series of changes of the upper-level circulation in the EA-WNP seems to suggest that the monsoon transition is connected to the reduced wind speed at the exit of the EAJS south of 45°N, east of 130°E. Zhang et al. (2006) showed a maximum diabatic heating of the troposphere and reduced meridional temperature gradient over western China and western Mongolia in mid-July, leading to a rapid westward shift of the core of the westerly jet in the 30°–45°N band (cf. Figs. 4 and 5 of Zhang et al. 2006). The westward shift of the jet core implies reduced wind east of 130°E in this latitude band prior to the monsoon transition, as shown in Fig. 8. The maximum inland heating in mid-July simply follows the seasonal procession of the sun with inland heating lagging the position of the sun by ~1 month. While the late July monsoon transition is connected to the maximum land heating, the exact timing of the monsoon onset might be related to the arrival of a wave packet in the “Silk Road pattern” and strengthening of the Bonin high. The Silk Road pattern refers to the propagation of stationary Rossby waves along the Asian jet in the upper troposphere, and the strengthening of the Bonin high at the arrival of the wave packet in late July corresponds to the termination of the baiu season, as demonstrated by Enomoto et al. (2003).

4. Changes of monsoon circulation in South and Southeast Asia

Regions of high pressure in the upper atmosphere correspond to large heating below. In May the center of
the Asian high is centered over Indochina where the mountainous terrain is heated by solar radiation that induces strong deep convection with added latent heating. In June solar heating in northern India and the Tibetan Plateau is strong, inciting the summer monsoon in India and the Bay of Bengal. The center of the Asian high is then shifted to the Tibetan Plateau. As the solar heating of land continues to increase in July, the Asian high strengthens and expands greatly, especially westward and northward where there are expansive arid lands. We have demonstrated in the previous section that the late July monsoon transition in the EA-WNP region is related to the northward shift and weakening of the EAJS south of 45°N that, in turn, is caused by an increase of the upper-tropospheric geopotential height at the northern flank of Asian high and a decrease of the geopotential height at the ridge of the Asian high. The former is likely related to the strong solar heating of the midlatitude deserts in July when heating attains a maximum (cf. Fig. 5 of Zhang et al. 2006), while the latter is related to a weakening of the South Asian monsoon circulation.

To investigate changes in the South Asian monsoon circulation during the late July EA-WNP monsoon transition, we show in Fig. 9 the southerly winds in the longitude–pressure plane, averaged over 15°–25°N, and wind speeds (m s⁻¹: color bar) and streamlines at the 1000-hPa pressure level, Terrain marked by black bars. Green shadings indicate regions where the differences are confident at the 90% level.

**FIG. 9.** Differences between consecutive pentads of southerly winds in the pressure–longitude plane (contours, averaged over 15°–25°N) and wind speeds (m s⁻¹: color bar) and streamlines at the 1000-hPa pressure level, Terrain marked by black bars. Green shadings indicate regions where the differences are confident at the 90% level. but decreases between 110° and 130°E nearly throughout the troposphere (Fig. 9a). Near the surface there is a weak anomalous (i.e., change) cyclonic circulation (also shown in Fig. 4a for the 850-hPa circulation). At the monsoon transition (Fig. 9b), the near-surface anomalous cyclonic circulation strengthens and moves westward to the SWNP. This is when the midoceanic low OLR region expands rapidly to the SWNP. The southerly flow decreases in the entire 15°–25°N latitude zone west of ~120°E (Fig. 9b). It is reasonable to expect that the monsoon circulation in southern Asia, including India and Indochina, decreases at the late July EA-WNP monsoon transition. In the South China Sea and Philippine Sea, the southwesterly wind strengthens, feeding more humid air from the tropics to the SWNP (lower section of Fig. 9b).

We further show in Fig. 10 the pressure–latitude cross sections of meridional circulation and the westerlies along the longitudinal bands 80°–100°E, 100°–120°E, and 120°–140°E. The orographic lifting of flow is clearly seen in South Asia and the Bay of Bengal (top panel of Fig. 10a). The ascent of air continues to weaken since two pentads prior to the monsoon transition, indicating a weakening of the South Asian monsoon circulation. In Indochina and the South China Sea, southerly and upward flows also continue to weaken prior to the monsoon transition (Fig. 10b). East of 120°E the upward air motion increases at the monsoon transition in the tropical and subtropical regions of the western NP (Fig. 10c). In the lower troposphere the largest increase in westerly
wind is found in the southern South China Sea and the Philippine Sea during the monsoon transition (lower panels of Figs. 10b,c).

5. Conclusions

In the search for causes of the rapid transition of the EA-WNP monsoon in late July, we investigated the upper tropospheric circulation. Based on OLR data, we found that there are 16 years in 1974–2010 with a clearly identifiable rapid change in the subtropical western NP from clear to cloudy conditions. By normalizing to the date of transition, we composited atmospheric conditions of those 16 summers and investigated the evolution of the atmospheric circulation during and prior to the rapid transition. We found that the transition of the EA-WNP monsoon is initiated by a weakening of the westerlies at the exit of the EAJS south of 45°N. This weakened jet stream incites a series of changes in downstream regions that lead to a rapid change in the lower-tropospheric circulation.

Two pentads prior to the rapid monsoon transition, the EAJS over the ocean south of 45°N, east of Japan weakens. This region is north of a sensitive region where the circulation turns from anticyclonic west of 160°E to cyclonic east of 160°E. Corresponding to the weakened jet stream, the potential vorticity increases in the region to the south. It signifies a weakened anticyclonic circulation in the eastern stretch of the Asian high west of 160°E and a weakened cyclonic circulation in the NP oceanic trough east of 160°E. These changes continue until the date of the rapid monsoon transition. Farther to the south in the subtropical midocean, the flow turns divergent. The divergent air in the upper troposphere pumps air upward, resulting in an increase of clouds and a reduction of OLR. The OLR in the SWNP to the west also decreases. The changes of divergence and OLR in these subtropical regions lag the changes of zonal wind and PV in the regions to the north by ~5 days. Under the influence of the NP high, the atmosphere in those subtropical regions is stable and clear prior to the monsoon transition. The series of changes described above is consistent with the existing theories and hypotheses that the intensity of the EAJS affects the location of the Bonin high and the withdrawal of the baiu (Enomoto et al. 2003; Suzuki and Hoskins 2009) and that upper-level

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**Fig. 10.** Pressure–latitude cross sections of winds (m s$^{-1}$) averaged over (a) 80°–100°E, (b) 100°–120°E, and (c) 120°–140°E: (top) average of days –6 to –10; (middle and bottom) differences between consecutive pentads (after minus before). Shadings are westerly winds; circulations are shown by vectors and streamlines, Terrain marked by black bars. Circles indicate regions where differences of the westerlies are confident at the 90% level.
PV anomalies initiate and support convection by destabilizing the lower troposphere and causing upward motion ahead on the PV tongue (Funatsu and Waugh 2008).

In association with the enhanced convection and decreased OLR in the subtropical regions, the low-level monsoon trough deepens over the Philippine Sea, and the western stretch of the NP high jumps northward by ~10° latitude to the south of Japan. Southwesterly monsoon flow in the South China Sea and Philippine Sea strengthens, whereas the OLR in Japan and Korea significantly increases and the baiu/Changma rainy season ends. The EA-WNP enters a new (the third) phase of summer monsoon. These changes suggest that the rapid monsoon transition in the lower troposphere is a response to the upper-tropospheric circulation.

Weakening of the EAJs south of 45°N prior to the rapid monsoon transition is caused by a reduction of the meridional gradient of geopotential height; the geopotential height increases in the extratropics with a maximum at ~45°N prior to the monsoon transition but decreases along the ridge of the Asian high. The former corresponds to the poleward shift of the jet axis (Zhang et al. 2006; Lin and Lu 2008) and is related to the intense solar heating of land with ~1-month lag of the sun’s position; while the latter is related to a weakened monsoon circulation (and latent heating) over India and the Bay of Bengal. Over the vast dry areas of inland Asia, solar heating varies smoothly with that of the sun’s position, but the South Asian monsoon circulation could fluctuate significantly within weeks. Therefore, it appears likely that in late July, when solar heating of inland Asia reaches a maximum, a weakened South Asian monsoon circulation might cause a weakened EAJS in the upper troposphere that induces a rapid transition of monsoon in the EA-WNP region. The exact timing of the rapid transition might also be connected to the development of the Bonin high, which is a result of the propagation of stationary Rossby waves along the EAJS (Enomoto et al. 2003).

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