Processes for the Northeastward Advance of the Summer Monsoon over the Western North Pacific

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Abstract

In the South China Sea (SCS) and western North Pacific (WNP), climatological summer monsoon sets in first in the SCS in mid-May, extends into the southwestern Philippine Sea in mid-June, and jumps suddenly northeastward to around 20°N, 150°E in late-July. The processes leading to this distinct northeastward advance of the summer monsoon are investigated using NCEP/NCAR reanalysis data for the period of 1979–1995. It is found that the cloud-radiation and wind-evaporation feedback plays an important role for the seasonal change in sea surface temperature (SST) in the WNP. Analysis shows that monsoon-induced changes in cloudiness and surface wind produce contrasting changes in surface short wave radiation, and latent heat flux between the convection and pre-convection region. The resultant SST tendency difference turns around the SST gradient east of the convection region in about one month, and induces the northeastward shift of the highest SST center. Following the ocean surface warming in the pre-convection region, the convective instability and low-level moisture convergence increase. East of the convection region, the reversal of the SST gradient tends to be associated with the transition of low-level winds from easterly to westerly. The results suggest that the SST change induced by the monsoon onset facilitates the northeastward extension of convection. It is speculated that the summer monsoon advance over the WNP may result from air-sea interactions.

1. Introduction

Two fundamental factors for the monsoon are seasonally varying solar radiation and land-sea thermal contrast (Ramage 1971; Webster 1987). With the seasonal north-south move of solar radiation, the rainy season commences in low latitudes and advances poleward. The observed monsoon advance, however, is not purely meridional. The summer monsoon advances northwestward over the Bay of Bengal and Indian

subcontinent (Indian Meteorological Department 1943; Rao 1976), and southeastward over Indonesia and northern Australia (Murakami and Matsumoto 1994). This feature is attributed to the land-sea distribution. Due to the difference in thermal inertia between land and sea, the land warms up more quickly than the ocean at the same latitude (Webster 1987). Because of the asymmetry of land distribution about the equator in the Asian and western Pacific sector, the land-sea thermal contrast induces a southeast-northwest march of convection (Meehl 1987).

Over the South China Sea (SCS) and western North Pacific (WNP), the summer monsoon marches northeastward (Wang 1994; Murakami and Matsumoto 1994) with three distinct stages observed in mid-May (pentad 28), mid-June (pentad 34), and late-July (pentad

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Fig. 1. 240 W/m^2 contours of OLR (a), locations of 850 hPa monsoon trough (b), and 5880 m contours of 500 hPa geopotential height (c). From top to bottom are for, respectively, pentad 26–29, pentad 32–35, and pentad 39–42. The labels on the contours denote the pentad. The data used are climatological pentad means. Adopted from Wu and Wang (2001).

41) (Wu and Wang 2000, 2001). This unusual monsoon advance is indicated by the northeastward extension of 240 W/m^2 contour of Outgoing Longwave Radiation (OLR) (Fig. 1a), and the northeastward move of the monsoon trough (Fig. 1b). In the current study, the 240 W/m^2 of OLR is taken as the threshold for monsoon rainfall (Murakami 1980; Graham and Barnett 1987). Different variables and criteria for monsoon onset can be found in literature. Recently, Qian and Yang (2000) argued that the upper tropospheric water vapor band brightness temperature is a better field than OLR for representing convective precipitation. For deep convection in the tropics, the brightness temperature and OLR are consistent. In the extratropics, it is still debatable whether the brightness temperature is superior to OLR. The monsoon trough on Fig. 1b is determined as the convergence zone between monsoon westerly winds and easterly trade winds. The eastward extension of monsoon convection observed in mid-June (pentad 33–34) occurs when the monsoon rainfall extends to the southwestern Philippine Sea (Fig. 1a), and the monsoon

trough moves quickly across the Philippine Islands (Fig. 1b). The northeastward extension after mid-July occurs as the monsoon rainfall penetrates to 165°E at pentad 41, and the monsoon trough shifts quickly northeastward at pentad 42. Obvious changes are observed in the WNP subtropical high: a northward lift in mid-June and a northward jump and split of the center at pentad 42 (Fig. 1c).

The summer monsoon progress over the WNP cannot be explained by the land-sea thermal contrast as the Indian monsoon. What is the reason for the unusual northeastward monsoon march over the WNP? The zonal asymmetry in sea surface temperature (SST) between western and eastern Pacific along $10^\circ\text{--}20^\circ\text{N}$ is thought to be a leading factor for the development of the WNP summer monsoon (Murakami and Matsumoto 1994). Ueda et al. (1995), and Ueda and Yasunari (1996) indicated the importance of the development of a warm SST tongue around 20°N and $150^\circ\text{--}160^\circ\text{E}$ in early-July for the abrupt jump of convection to around 25°N , 150°E in late-July. The northeastward monsoon advance over the WNP seems to follow

the northeastward migration of the warm SST tongue (Wu and Wang 2001). Kawamura et al. (1998, 2001) suggested that the east-west gradient in summertime SST anomalies between the SCS and tropical western Pacific is intimately linked to the anomalous tropical convection near the Philippines. According to Lindzen and Nigam (1987), the SST gradient can affect low-level winds through differential surface heat fluxes and turbulent mixing in the atmospheric boundary layer. As such, it may be an important factor that determines where monsoon westerly winds can penetrate in the oceanic WNP monsoon region. A different view was presented in Kawamura and Murakami (1998), and Murakami et al. (1999). They proposed that the intensification of the monsoon trough at the height of the WNP summer monsoon (late-July) is indirectly caused by the remote influence of the Asian heat low. To understand the role of SST change, we need to diagnose how changes in SST, and SST gradient relate to the monsoon advance. This is one of the motivations for the present study.

An unresolved question is why the warm SST tongue moves northeastward in the WNP. Ueda and Yasunari (1996) indicated that the warm SST tongue around 20°N, 150°–160°E in early-July develops due to the appearance of a weak wind region around 25°N, 140°–160°E in late-June that is a part of the atmospheric response to the convection near the Philippines. Kawamura and his collaborators suggested that a longitudinal asymmetry of surface latent heat flux anomalies is responsible for the east-west gradient in summertime SST anomalies between the SCS and tropical western Pacific (Kawamura et al. 1998, 2001). Wu and Wang (2001) noticed that the seasonal change in SST in the WNP could involve cloud-radiation and wind-evaporation feedback. However, the processes for those feedbacks were not demonstrated clearly. This is another motivation for the present study.

The almost simultaneous and sudden changes observed in both convection and atmospheric circulation suggest that the monsoon onset may involve internal atmospheric dynamics. For example, Lau and Yang (1996) and Krishnakumar and Lau (1998) proposed the role of symmetric instability of the basic flow in May in the sudden jump of the equatorial Inter-tropical

Convergence Zone (ITCZ) to 10°N in mid-May. The present study, however, will not address the mechanism for the sudden monsoon onset. Instead, the objective of the present study is to understand the processes that are responsible for the northeastward advance of the summer monsoon over the WNP. The processes may involve the SST effects on the atmosphere, and the monsoon feedback on the SST change. Previous studies (Ueda and Yasunari 1996; Wu and Wang 2001) demonstrated the role of SST change in destabilizing the atmosphere. The present study emphasizes the reversal of the SST gradient east of the convection region, and its relation to the northeastward extension of convection. In particular, this study focuses on how monsoon-induced changes in cloudiness and circulation modify surface heat fluxes and turn around the SST gradient east of the convection region. In addition, this study elaborates the relative contributions of changes in solar radiation and latent heat flux to changes in the SST tendency. The organization of the text is as follows. Section 2 provides a brief description of data and methods. Section 3 discusses the SST change and its plausible effects. In section 4, monsoon-induced changes in surface heat fluxes and their impacts on the SST change will be discussed. Section 5 summarizes the major results and provides some discussion.

2. Data and methods

Daily mean OLR data on $2.5^\circ \times 2.5^\circ$ grids were derived from National Ocean and Atmospheric Administration (NOAA) satellite observations (Gruber and Krueger 1984) from 1975 to 1995, with 9 months of data missing in 1978. Daily mean wind, vertical p -velocity, temperature, and specific humidity at selected levels were extracted from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996), for the period of 1979–1995. These data have a horizontal resolution of $2.5^\circ \times 2.5^\circ$. The NCEP/NCAR reanalysis also provided skin temperature, air temperature and specific humidity at 2 m, wind at 10 m, surface heat fluxes, total cloudiness, and clear sky solar radiation at surface on Gauss grids of T62, with a horizontal resolution of about $1.9^\circ \times 1.9^\circ$. Over the open ocean, the skin temperature is fixed at the Reynolds SST (Rey-

nolds and Smith 1994) in the NCEP/NCAR reanalysis. The surface heat fluxes and total cloudiness belong to the C-type data in the reanalysis and have uncertainties. The uncertainties could be reduced when climatological means are used.

For all the variables, pentad means were constructed for each year and then climatological pentad means were calculated as multi-year means. The instability index used in this study is defined as the difference in hypothetically saturated equivalent potential temperature between 1000 hPa and 700 hPa (divided by the pressure difference) (Wu and Wang 2001). Surface moisture divergence was calculated using daily mean surface wind (10 m), and specific humidity (2 m) for each year, and climatological pentad means were obtained by multi-year averaging.

The change in latent heat flux can be due to that in wind speed, air-sea humidity difference, and high-frequency disturbance. To understand relative importances of those factors, contributions due to changes in wind speed and air-sea humidity difference were estimated based on surface wind (10 m) and humidity (2 m) data as conventionally with a constant exchange coefficient of 1.5×10^{-3} . When computing the contribution due to the change in wind speed (or air-sea humidity difference) at a specific time, only the change in wind speed (or air-sea humidity difference) at that time was considered. The contributions were calculated for each year based on pentad means and then averaged over multi-years to get their climatological pentad means. Note that the latent heat flux from the NCEP/NCAR reanalysis was originally calculated using instant atmospheric wind and humidity at $\sigma = 0.995$ level. The sum of the two estimated contributions based on the above procedure differs from the total change in latent heat flux. The discrepancy can be attributed to the effect of high frequency disturbances (period < 5 days), the difference of height of atmospheric elements used, and the change in the exchange coefficient.

3. SST change and its effects

3.1 SST change

The warm SST tongue (SST over 29°C) features a northeastward move in the SCS and WNP region. North of 10°N , SST reaches 29°C

Fig. 2. (a) Climatological mean SST 29°C contour and location of the highest SST center indicated by bold open circle with the number denoting the first pentad when the highest SST is observed. The three shaded boxes denote the regions for the area mean curves in Figs. 2b, 2c and Fig. 6. (b) Climatological pentad mean SST ($^{\circ}\text{C}$) over $6^{\circ} \times 6^{\circ}$ boxes centered at 12.5°N and 116°E (solid curve), 16°N and 127.5°E (dashed curve), and 18°N and 142.5°E (dotted curve) as indicated by shaded boxes in Fig. 2a. (c) Similar to (b) except for OLR (W/m^2). The vertical dashed lines in (b) and (c) denote the time for the three onsets.

first in the SCS (Fig. 2a). Before pentad 25, the center of the highest SST is located in the western equatorial Pacific (not shown). The SST center shifts to around 12.5°N and 116°E at pentad 26. The SST center stays there until

pentad 34 when the SST around 16°N and 127.5°E becomes the highest. At pentad 38, the SST center starts to move further eastward. During pentad 42–46, the SST center stays near 18°N and 142.5°E .

In concurrent with the shift of the SST center from the SCS to the western Philippine Sea at pentad 34, the SST gradient between the SCS and Philippine Sea reverses. This reversal can be seen in the time series of area mean SSTs shown in Fig. 2b. Mid-June is the time when SST in the western Philippine Sea becomes higher than that in the SCS, which is concurrent with the quick drop of OLR to below 240 W/m^2 near 16°N and 127.5°E (Fig. 2c). This change is due to a continuous rising of SST in the Philippine Sea and flattening and then decreasing of SST in the SCS during mid-May to mid-June. Notice that SST near 12.5°N and 116°E starts to flatten at the SCS monsoon onset (Fig. 2c). The distribution of the SST change from mid-May to mid-June shows distinct west-east contrast along 7° – 17°N (Fig. 3a). In one month, the SST decreases by about 0.2°C in the central SCS and increases by about 0.6°C in the western Philippine Sea.

Another change in the SST gradient is observed in early- to mid-July when the SST gradient between the western and northeastern Philippine Sea reverses and the center of the highest SST shifts to east of 135°E (Fig. 2a). This switch is reflected in the crossing of area mean SST curves in the western and northeastern Philippine Sea (Fig. 2b). This reversal in the SST gradient precedes the drop of OLR to below 240 W/m^2 near 18°N and 142.5°E (Fig. 2c). After mid-June, the SST in the western Philippine Sea turns to decrease, whereas the SST in the northeastern Philippine Sea keeps rising slowly. The distribution of the SST change from mid-June to late-July is characterized by a southwest-northeast contrast in the Philippine Sea (Fig. 3b). The SST decreases by about 0.2° – 0.4°C in the western Philippine Sea, and increases by about 0.4°C in the northeastern Philippine Sea, respectively.

3.2 Effects of SST change

The SST effects can be understood in terms of changes in local SST and SST gradient. The development of high SST can increase surface air temperature and humidity through surface

heat fluxes and turbulent mixing in the atmospheric boundary layer. The SST gradient can induce low-level moisture convergence through differential surface heat fluxes. The SST effects are supported by the coherent development and northeastward extension of high SST and instability region (Wu and Wang 2001). The scatter diagram shown in Fig. 4a demonstrates consistent changes in SST and the instability index. The correlation coefficient between the area mean SST and instability index over 10° – 20°N , 125° – 155°E is 0.97. Surface moisture convergence tends to increase with SST, especially when SST is below 29°C (Fig. 4b). After the monsoon onset (in the WNP, the monsoon sets in after the SST reaches over 29°C), the relation becomes weaker. The correlation coefficient between the area mean SST and surface moisture divergence over the above region is -0.95 .

The relationship between the SST gradient and low-level wind is indicated by the coherent evolution of the east-west SST difference and the 850 hPa zonal wind east of the convection region (Fig. 5), especially during April through October when the mean SST is over 28°C (Fig. 2b). The switch of the 850 hPa zonal wind from easterly to westerly in the region of 115° – 125°E , 5° – 15°N (Fig. 5b) occurs when the SST difference between the SCS and western Philippine Sea decreases quickly and changes to negative (Fig. 5a). With the switch of zonal winds near the Philippines, the monsoon trough intrudes into the southwestern Philippine Sea (Fig. 1b), convection develops there (Figs. 1a, 2c), and the subtropical high moves northward (Fig. 1c). Similarly, at a few pentads after the SST gradient reverses between the western and northeastern Philippine Sea (Fig. 5a), the 850 hPa zonal wind in between (130° – 140°E , 7.5° – 17.5°N) switches from easterly to westerly (Fig. 5b). With this switch of zonal winds, the monsoon trough shifts to the eastern Philippine Sea (east of 135°E) (Fig. 1b), deep convection develops around 20°N , 150°E (Figs. 1a, 2c), and the subtropical high splits and jumps northward (Fig. 1c).

4. Changes induced by the monsoon onset

The monsoon onset induces changes in cloudiness and surface wind. These changes occur

Fig. 3. Change of climatological mean SST ($^{\circ}\text{C}$) from pentad 27–28 to pentad 33–34 (a) and from pentad 33–34 to pentad 40–41 (b).

Fig. 4. Scatter diagrams of SST versus instability index over 10° – 20°N , 125° – 155°E (a) and SST versus surface moisture divergence over 10° – 20°N , 125° – 155°E (b). Refer to the text for the calculation of surface moisture divergence and the definition of the instability index.

Fig. 5. (a) Difference of area mean SST ($^{\circ}\text{C}$) between the SCS and western Philippine Sea (solid curve) and between the western Philippine Sea and northeastern Philippine Sea (dashed curve). See shaded boxes in Fig. 2a for the regions where the area means are calculated. (b) Area mean 850 hPa zonal wind (m/s) averaged over 115° – 125°E , 5° – 15°N (solid curve) and over 130° – 140°E , 7.5° – 17.5°N (dashed curve). The data used are climatological pentad means. The vertical dashed lines in (a) and (b) denote the time for the three onsets.

not only locally, but also away from the convection region. The induced changes in surface heat fluxes can affect the SST change. The processes for the SST change include surface heat flux, oceanic mixing, upwelling, and oceanic advection. Webster et al. (1998) indicated that in the western Pacific warm pool region, surface heat flux is a dominant factor in the seasonal change in SST. Based on the results of an ocean general circulation model (OGCM) forced by daily mean wind stress and heat flux derived from NCEP/NCAR reanalysis, Kawamura et al. (2001) made a heat budget analysis for the near-surface layer in the region of 5° – 15°N , 140° – 160°E . The seasonal climatology of the heat budget suggests that in boreal summer

the time tendency in the heat content in the WNP is mainly due to surface heat flux in the climatological sense. The contribution from horizontal and vertical convergence of heat transport is very small. In the present study, quantitative estimates are made for short wave radiation and latent heat flux, the dominant two terms in surface heat fluxes over the ocean. Oceanic mixing and upwelling is only inferred qualitatively, based on changes in surface wind and vorticity. The following discussion focuses on changes at the monsoon onset, since monsoon-induced changes are more easily identified at that time.

4.1 Mid-May change

Following the SCS monsoon onset, SST tendency in the SCS changes quickly from about $0.2^{\circ}\text{C}/\text{pentad}$ to negative (Fig. 6a). Near 16°N and 127.5°E , the SST tendency decreases from about $0.2^{\circ}\text{C}/\text{pentad}$ to $0.1^{\circ}\text{C}/\text{pentad}$. In the northeastern Philippine Sea, the SST tendency increases by about $0.1^{\circ}\text{C}/\text{pentad}$. The distribution of changes in SST tendency (Fig. 7a) corresponds well to that of changes in short wave radiation (Fig. 7b) and latent heat flux (Fig. 7c).

The decrease in SST tendency in the SCS is due to reduced short wave radiation (Figs. 6b, 7b) and enhanced latent heat flux (Figs. 6c, 7c). Short wave radiation has a larger contribution ($25 \text{ W}/\text{m}^2$) than latent heat flux (10 – $15 \text{ W}/\text{m}^2$). The local decrease in short wave radiation is caused by increased cloudiness (Fig. 8a) following the development of strong upward motion (Fig. 8b). The local increase in latent heat flux is contributed by the change in surface wind speed (Fig. 8c). The contribution from the change in air-sea humidity difference is negligible (Fig. 8d). High frequency disturbance contributes to the decrease in SST tendency in disturbed regions through inducing large amounts of evaporation, as well as oceanic mixing. The transition from anticyclonic to cyclonic vorticity (Lau and Yang 1997; Wang and Wu 1997) suggests that upwelling contributes to the local cooling.

The decrease in the SST tendency in the western Philippine Sea is mainly caused by reduced short wave radiation (Figs. 6b, 7b) that results from the extension of increasing cloudiness from the subtropics (Fig. 8a). In the north-

Fig. 6. Climatological pentad means of the SST tendency ($^{\circ}\text{C}/\text{pentad}$) (a), surface net short wave radiation (W/m^2) (b), and surface latent heat flux (W/m^2) (c) over $6^{\circ} \times 6^{\circ}$ boxes centered at 12.5°N and 116°E (solid curves), 16°N and 127.5°E (dashed curves), and 18°N and 142.5°E (dotted curves) as indicated by shaded boxes in Fig. 2a. The vertical dashed lines in (a), (b), and (c) denote the time for the three onsets.

eastern Philippine Sea, the increase in SST tendency is due to increasing short wave radiation (Figs. 6b, 7b) following a decrease in cloudiness (Fig. 8a) and decreasing latent heat flux (Figs. 6c, 7c). The decrease in latent heat flux in the Philippine Sea is mainly due to weakening winds (Fig. 8c) as the monsoon trough pushes eastward (Fig. 1b) and the subtropical high retreats northeastward (Fig. 1c). The contribution due to the change in air-sea humidity difference is small and opposite (Fig. 8d). Reduced oceanic mixing due to decreasing wind speed may favor a rise in SST in the Philippine

Sea. Most of the Philippine Sea experiences an increase in surface anticyclonic vorticity (Figure not shown), which may contribute to the SST increase through downwelling.

The increasing short wave radiation occurs east of the convection region (Fig. 7b). To determine how much the change in solar inclination contributes to the increase in solar radiation, the change in clear sky solar radiation at the surface (purely due to the change in solar inclination) was examined. The results indicate an increase of only $3\text{--}4 \text{ W}/\text{m}^2$ along 25°N , and a decrease of only $4\text{--}5 \text{ W}/\text{m}^2$ along 5°N from pentad 26–27 to pentad 28–29. Along 15°N , the change is less than $1 \text{ W}/\text{m}^2$. Thus, the observed increase in solar radiation is mainly due to the decrease in cloudiness (Fig. 8a) that corresponds to the change in vertical motion (Fig. 8b).

The positive change in vertical p -velocity (Fig. 8b) is consistent with the northeastward retreat of the WNP subtropical high (Lau and Yang 1997; Fig. 1c). Comparison of vertical motion before and after the monsoon onset shows a decrease in upward motion south of 17°N and an increase in downward motion north of 17°N (Figures not shown). Both changes are responses to the development of the rain band extending from the SCS to the subtropical WNP. Inspection of vertical sections demonstrates the presence of a local east-west circulation around mid-May (e.g., along 17.5°N), which is composed of low-level easterly south of the ridge of the subtropical high, upward motion over the SCS-Philippines, upper-level westerly north of the ridge of the South Asian anticyclone, and downward motion east of 140°E . The difference field of vertical circulation between pentad 28–29 and pentad 26–27 shows descending motion over the eastern Philippine Sea below 500 hPa, which is connected to the rising motion over the SCS-Philippines (Figure not shown). However, the direct cause for the compensating downward motion seems to be the intensification of a north-south vertical circulation between the subtropical high and the subtropical rain band.

4.2 Mid-June change

Following the mid-June monsoon onset in the southwestern Philippine Sea, the SST tendency near 16°N and 127.5°E changes quickly

Fig. 7. Change of climatological pentad means of the SST tendency ($^{\circ}\text{C}/\text{pentad}$) (a), surface net short wave radiation (W/m^2) (b), and surface latent heat flux (W/m^2) (c) from pentad 26–27 to pentad 28–29.

from $0.1^{\circ}\text{C}/\text{pentad}$ to less than $-0.05^{\circ}\text{C}/\text{pentad}$ (Fig. 6a). Around 18°N and 142.5°E , the SST tendency shows a decrease less than $0.05^{\circ}\text{C}/\text{pentad}$. The spatial distribution shows decreases in SST tendency in the SCS and most of the Philippine Sea, and increases to the northeast and southeast (150° – 165°E , 5° – 10°N) (Fig. 9a). The distribution of changes in SST tendency appears to result from a combination of changes in short wave radiation and latent heat flux (Figs. 9b, c).

The reduction of SST tendency in the western Philippine Sea is due to both reduced short wave radiation (15 – $20 \text{ W}/\text{m}^2$) (Figs. 6b, 9b) and enhanced latent heat flux (15 – $20 \text{ W}/\text{m}^2$) (Figs.

6c, 9c). The former is due to the increase in cloudiness (Fig. 10a) following the development of strong upward motion (Fig. 10b). The latter is contributed by an increase in surface wind speed (Fig. 10c), and high frequency disturbance. The contribution due to the change in air-sea humidity difference is very small (Fig. 10d). There may be positive contributions from oceanic mixing and upwelling because of the development of wind disturbance and cyclonic vorticity. The reasons for decrease in SST tendency in the northeastern Philippine Sea are similar except that changes in surface heat fluxes are smaller (Figs. 9b–c).

Over 20° – 25°N and 155° – 160°E (Ueda and

Fig. 8. Change of climatological pentad means of the total cloudiness (%) (a), 700 hPa vertical p -velocity (0.01 Pa/s) (b), and surface latent heat flux (W/m^2) due to the change in wind speed (c) and surface air-sea humidity difference (d) from pentad 26–27 to pentad 28–29.

Yasunari 1996), the SST tendency increases by about $0.04^\circ\text{C}/\text{pentad}$ (Fig. 9a) due to reduced latent heat flux (Fig. 9c). The change in short

wave radiation is small in this region (Fig. 9b). Ueda and Yasunari (1996) pointed out the role of reduced evaporation for the SST increase in

Fig. 9. Similar to Fig. 7 except from pentad 32–33 to pentad 34–35.

early-July around 20°N and 155°E . The decrease in latent heat flux in this region may be mainly caused by the suppression of high frequency disturbances that may also reduce oceanic mixing. Suppression of these disturbances could be a response to the development of convection in the southwestern Philippine Sea through Rossby wave propagation. Changes in low-level winds in mid-June display a clear wave train propagating northeastward from the Philippines with a low-level anticyclone located around 25°N and 140°E (Ueda and Yasunari 1996; Wu and Wang 2001). East of the anticyclone where the subtropical high is weakened (Ueda and Yasunari 1996), reduced surface winds suppress the disturbances.

The increasing short wave radiation in the subtropics (Fig. 9b) is consistent with decreasing cloudiness (Fig. 10a). The decrease in cloudiness occurs at higher latitudes than the upward motion (Fig. 10b), which is related to the northward tilting of the observed anticyclonic wind change with height (Wu and Wang 2001). At this time, the sun is near its northernmost latitude and the change in clear sky short wave radiation between 10° – 25°N is small ($<1 \text{ W/m}^2$). Thus, the spatial pattern of changes in short wave radiation is due to changes in cloudiness. The enhanced downward motion along 20° – 25°N (Fig. 10b) is consistent with the northward lift of the subtropical high (Fig. 1c). Inspection of north-south vertical circulation

Fig. 10. Similar to Fig. 8 except from pentad 32–33 to pentad 34–35.

(e.g., along 135°E) indicates that at mid- and low-levels the enhanced downward motion is due to the development of convection in the southwestern Philippine Sea, and at the upper-levels it is mainly due to the start of Baiu in Japan.

5. Summary and discussion

5.1 Summary

The present study based on NCEP/NCAR re-analysis data from 1979 to 1995 demonstrates that the monsoon onset induces important

Fig. 11. Schematic diagrams for major changes at the mid-May (a) and mid-June (b) onset. The solid (dashed) lines denote the location of the monsoon trough before (after) the onset. The closed solid (dashed) curves indicate the subtropical high before (after) the onset. The light (heavy) shadings signify regions of increase (decrease) of cloud and/or evaporation. The dashed arrows indicate the direction for the extension of low-level westerly winds.

changes in surface heat fluxes by altering cloudiness and surface wind. The increased cloudiness reduces locally the incoming short wave radiation. The increased wind speed and intrusion of monsoon disturbances enhances evaporation and oceanic mixing. The development of cyclonic vorticity induces upwelling. Northeast of the convection region, the compensating downward motion in relation to the move of the subtropical high increases the incoming short wave radiation. The decrease in wind speed, and suppression of high frequency disturbance in association with the move of the monsoon trough and subtropical high reduce evaporation and oceanic mixing. The major changes at the two monsoon onsets in mid-May and mid-June are summarized in Fig. 11.

The monsoon-induced contrasting change in

surface heat fluxes between the convection region and the pre-convection region to the northeast induces a similar contrast for the change in SST tendency. This leads to the reversal of the SST gradient east of the convection region in about one month after the monsoon onset. As a result, the warm SST tongue shifts eastward, which induces the destabilization of atmosphere. The switch of the SST gradient enhances the low-level moisture convergence northeast of the convection region through differential surface heat fluxes and turbulent mixing in the atmospheric boundary layer (Lindzen and Nigam 1987). The turn-around of the SST gradient is almost simultaneous with the switch of low-level zonal wind from easterly to westerly east of the convection region. The relationship seems to be consistent with that for

the interannual anomalies (Kawamura et al. 2001). Thus, the SST change effected by the monsoon onset seems to set up the atmosphere in a favorable state for the northeastward extension of convection.

Based on the above results, it is speculated that the distinct northeastward progress of the summer monsoon over the WNP involves important air-sea interactions. The processes leading to the northeastward advance of the summer monsoon are illustrated using a schematic diagram shown in Fig. 12. Following the SCS monsoon onset in mid-May, over the SCS there is increase in cloudiness, surface wind speed, and high-frequency disturbance. In contrast, northeast of the convection region cloudiness decreases, and surface wind is weakened. These monsoon-induced changes in cloudiness and surface wind cause a longitudinal difference in changes in short wave radiation and evaporation. This leads to a contrasting change in SST between the SCS and Philippine Sea. As a result, the SST gradient across the Philippines turns around and the warm SST center shifts eastward to the western Philippine Sea in mid-June. This is followed by increases in convective instability and low-level moisture convergence over the western Philippine Sea, which favors the development of deep convection. When some trigger for convection arrives (for example, intra-seasonal oscillation), the convection region extends eastward, and the summer monsoon onsets over the southwestern Philippine Sea. After the convection occurs, over the western Philippine Sea cloudiness and surface wind speed increase and high-frequency disturbance develops, whereas northeast of the convection region (20° – 30° N and 150° – 160° E) high-frequency disturbance is suppressed. These changes induce a southwest-northeast pattern of changes in short wave radiation and evaporation. This results in a contrasting change in SST between the southwestern and northeastern Philippine Sea. As such, the warm SST center moves northeastward and the SST gradient between the western and northeastern Philippine reverses. This leads to the development of high convective instability and low-level moisture convergence over the northeastern Philippine Sea, which sets up the atmosphere for the extension of monsoon rain to around 20° N, 150° E in late-July.

5.2 Discussion

Under seasonally varying solar radiation, the land-sea thermal contrast and the SST gradient are fundamental factors for the change in large-scale circulation. While the continental scale land-sea thermal contrast controls the tropospheric temperature gradient and dominates the broad-scale monsoon evolution, the SST gradient is important in modifying the low-level circulation and thermodynamics and contributes to the regional monsoon features. Especially in the oceanic WNP monsoon region where the mean SST is high in boreal summer, low-level winds may be sensitive to changes in the SST gradient. Previous studies indicate a close relation between low-level convergence and convection when the SST is in excess of the threshold (Graham and Barnett 1987). In the WNP, the convection sets in after the SST reaches over 29° C on the pentad mean time scale (Figure not shown), a value much higher than the threshold derived from monthly mean observations (28° C, e.g., Graham and Barnett 1987). Thus, changes in the SST gradient may play an important role for the extension of monsoon convection within the large-scale background set up by the land-sea thermal contrast.

The proposed mechanism for the extension of convection to the northeastern Philippine Sea is similar to the maturing process of the WNP summer monsoon suggested by Ueda and Yasunari (1996). They focused on the SST change around 20° N, 155° E and its impact on the atmospheric stability. The present study contrasted changes in surface heat fluxes between the convection and pre-convection region, and stressed the change in the SST gradient. Murakami et al. (1999) inferred that the SST is not really a matter of importance for the mature phase of the WNP summer monsoon. Their inference was based on the small change of *in situ* 1000 hPa temperature compared to the large decrease of 1000 hPa geopotential height along 25° N across the WNP from pentad 40 to pentad 46 (see their Fig. 4). The contrasting early-July SST pattern between typical and atypical years (corresponding to occurrence of non-occurrence of convection around 20° N, 155° E in late-July) shown by Ueda and Yasunari (1996) (see their Fig. 5), however, indicates the importance of the SST change. Although the differences in actual SST are small (a few tenth of 1° C), the

Fig. 12. Schematic diagram illustrating the processes for the northeastward advance of the summer monsoon over the western North Pacific. The thin arrows indicate the sequence of changes. The thick dashed arrows in the left denote the direction for the time evolution. The dashed horizontal arrows represent the role of the trigger for convection.

SST pattern is quite different between the two types of years. Because of the high mean SST in the WNP, a small change in the SST can reverse the SST gradient and effects the atmosphere.

As to the reason for the seasonal SST change, surface heat flux is stressed in this study. Although it is likely that surface heat flux is the major reason for the seasonal SST change in the WNP, discrepancies exist between net sur-

face heat flux and the SST tendency. For example, the SST tendency in the SCS changes its sign in mid-May, whereas net surface heat flux maintains its positive value until early June. The SST tendency in the western Philippine Sea changes to negative in mid-June, whereas net surface heat flux remains positive. The discrepancies indicate that oceanic processes also affect the seasonal SST change. The role of oceanic mixing and upwelling may be inferred from the development of monsoon disturbances and the transition of low-level vorticity from cyclonic to anticyclonic at the monsoon onsets. Previous studies indicated the role of oceanic advection (Chu et al. 1993; Masumoto and Yamagata 1991). In view of the fact that oceanic mixing and upwelling depends upon surface wind, and that oceanic current upon atmospheric wind forcing, those oceanic processes reflect the monsoon feedback on the SST change as well. Quantitative estimates of those processes require reliable ocean data.

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References

- Chu, P.C., C. Li, D. Ko, and C.N.K. Mooers, 1993: A numerical simulation of seasonal circulation in the South China Sea. AGU Fall Meeting, San Francisco, California, 6–10 December 1993.
- Graham, N.E. and T.P. Barnett, 1987: Sea surface temperature, surface wind divergence, and convection over tropical oceans. *Science*, **238**, 657–659.
- Gruber, A. and A.F. Krueger, 1984: The status of the NOAA outgoing longwave radiation data set. *Bull. Amer. Meteor. Soc.*, **65**, 958–962.
- Indian Meteorological Department, 1943: *Climatological Atlas for Airmen*. Indian Meteorological Department, 100pp.
- Kalnay, E. and Coauthors, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471.
- Kawamura, R., M. Sugi, T. Kayahara, and N. Sato, 1998: Recent extraordinary cool and hot summers in East Asia simulated by an ensemble climate experiment. *J. Meteor. Soc. Japan*, **76**, 597–617.
- , T. Matsuura, and S. Iizuka, 2001: Interannual atmosphere-ocean variations in the tropical western North Pacific relevant to the Asian summer monsoon-ENSO coupling. *J. Meteor. Soc. Japan*, **79**, 883–898.
- and T. Murakami, 1998: Baiu near Japan and its relationship to summer monsoons over Southeast Asian and the western North Pacific. *J. Meteor. Soc. Japan*, **76**, 619–639.
- Krishnakumar, V. and K.-M. Lau, 1998: Possible role of symmetric instability in the onset and abrupt transition of the Asian monsoon. *J. Meteor. Soc. Japan*, **76**, 363–383.
- Lau, K.-M. and S. Yang, 1996: Seasonal variation, abrupt transition, and intraseasonal variability associated with the Asian summer monsoon in the GLA GCM. *J. Climate*, **9**, 965–985.
- and ———, 1997: Climatology and interannual variability of the southeast Asian summer monsoon. *Adv. Atmos. Sci.*, **14**, 141–162.
- Lindzen, R.S. and S. Nigam, 1987: On the role of sea surface temperature gradient in forcing low-level winds and convergence in the tropics. *J. Atmos. Sci.*, **45**, 2440–2458.
- Masumoto, Y. and T. Yamagata, 1991: Response of the western tropical Pacific to the Asian winter monsoon: The generation of the Mindanao Dome. *J. Phys. Oceanogr.*, **21**, 1386–1398.
- Meehl, G.A., 1987: The annual cycle and interannual variability in the tropical Pacific and Indian Ocean regions. *Mon. Wea. Rev.*, **115**, 27–50.
- Murakami, T., 1980: Empirical orthogonal function analysis of satellite-observed outgoing longwave radiation during summer. *Mon. Wea. Rev.*, **108**, 205–222.
- and J. Matsumoto, 1994: Summer monsoon over the Asian continent and western North Pacific. *J. Meteor. Soc. Japan*, **72**, 719–745.
- , ———, and A. Yatagai, 1999: Similarities as well as differences between summer monsoons over Southeast Asia and western North Pacific. *J. Meteor. Soc. Japan*, **77**, 887–906.
- Qian, W. and S. Yang, 2000: Onset of the regional monsoon over Southeast Asia. *Meteor. Atmos. Phys.*, **75**, 29–38.
- Ramage, C.S., 1971: *Monsoon Meteorology*. International Geophysics Series, Vol. 15, 296pp, Academic Press.

- Rao, Y.P., 1976: *Southwest Monsoon*. Synoptic Meteorology, Meteorological Monograph, No. 1., Indian Meteorological Department, 367pp.
- Reynolds, R.W. and T.M. Smith, 1994: Improved global sea surface temperature analysis using optimum interpolation. *J. Climate*, **7**, 929–948.
- Ueda, H., T. Yasunari, and R. Kawamura, 1995: Abrupt seasonal change of large-scale convective activity over the western Pacific in the northern summer. *J. Meteor. Soc. Japan*, **73**, 795–809.
- and ———, 1996: Maturing process of the summer monsoon over the western North Pacific—A coupled ocean/atmosphere system. *J. Meteor. Soc. Japan*, **74**, 493–508.
- Wang, B., 1994: Climatic regimes of tropical convection and rainfall. *J. Climate*, **7**, 1109–1118.
- and R. Wu, 1997: Peculiar temporal structure of the South China Sea summer monsoon. *Adv. Atmos. Sci.*, **14**, 177–194.
- Webster, P.J., 1987: The elementary monsoon. *Monsoons*, J. S. Fein and P. L. Stephens, Eds., 3–32.
- , V.O. Magana, T.N. Palmer, T.A. Tomas, N. Yanai, and T. Yasunari, 1998: Monsoons: Processes, predictability, and perspectives for prediction. *J. Geophys. Res.*, **103**, 14451–14510.
- Wu, R. and B. Wang, 2000: Interannual variability of summer monsoon onset over the western North Pacific and the underlying processes. *J. Climate*, **13**, 2483–2501.
- and ———, 2001: Multi-stage onset of the summer monsoon over the western North Pacific. *Clim. Dyn.*, **17**, 277–289.