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# Associations between the Western North Pacific Monsoon and the South China Sea Monsoon①

Lu Riyu (陆日宇)

*Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100080*

Chan-Su Ryu

*Department of Earth Science, Chosun University, Kwangju, Korea*

Buwen Dong

*CGAM, Department of Meteorology, University of Reading, Reading, United Kingdom*

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ABSTRACT

Based on the interannual variability of convection over the tropical western North Pacific (WNP), a region of 130°—160°E, 10°—20°N, a composite analysis is performed on the fields of surface temperature, outgoing longwave radiation and 850 hPa zonal wind. The composite results show that the weaker (stronger) WNP convection is related to the El Niño (La Niña)-pattern sea surface temperature (SST) anomalies in the preceding winter and in spring. A comparison with previous results indicates that a similar spatial and temporal distribution of SST anomalies is also associated with the onsets of both the WNP and South China Sea (SCS) monsoons.

The composite results also show that the weaker (stronger) convection over the WNP corresponds to the easterly (westerly) anomalies that extend westward from the WNP into the Bay of Bengal. A numerical experiment by an atmospheric general circulation model shows a similar result. In addition, during weaker (stronger) convection summer, the convection over the WNP and lower-level zonal winds over the SCS exhibit a small (large) extent of seasonal evolution.

**Key words**: South China Sea, Lower-level zonal flow, Tropical western North Pacific, Monsoon, El Niño

## 1. Introduction

The East Asian summer monsoon is significantly influenced by the convective activity over the tropical western North Pacific (hereafter WNP) (Huang and Li, 1987; Nitta, 1987; Kurihara, 1989; Huang and Sun, 1992; Murakami and Matsumoto, 1994; Ueda and Yasunari, 1996; Kawamura and Murakami, 1998; Lu, 2001a). It has been pointed out that an atmospheric Rossby wave is generated by anomalous convective activities over the WNP and propagates to the extratropics, and influences the interannual and intraseasonal variations of the East Asian summer monsoon.

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On the other hand, some evidence suggests that there is interaction between the East Asian summer monsoon and South China Sea (SCS) monsoon (Chen et al., 2000c). There are in-phase intraseasonal oscillations between the Meiyu latitudinal location and the intensity of westerlies over the SCS. In addition, rainfall activity along the Meiyu front is stronger (weaker) when the SCS monsoon is break (active).

As a possibly precursory signal of the Indian and East Asian summer monsoons, the SCS monsoon has been widely studied. Generally, these studies focused on the onset of SCS monsoon (Chen et al., 2000a; Chen et al., 2000c; Dai et al., 2000; He et al., 2000; Jian et al., 2000; Liu and Ding, 2000; Liu et al., 2000; Mu and Li, 2000; Sun and Chen, 2000; Wu et al., 2000; Zhu and Xu, 2000, among others). The onset of SCS monsoon generally happens around the fourth pentad of May and is prior to the Indian summer monsoon and East Asian summer monsoon.

Several studies showed that the SCS monsoon exhibits a significant intraseasonal variation (Chen and Chen, 1995; Mu and Li, 2000). Although around its onset the SCS monsoon does not show association with the convective activity over the WNP, the intraseasonal variation of the SCS monsoon after its onset may be influenced by the WNP monsoon, particularly around the onset of the latter. According to the theory of Gill (1980), when there is a heat source or anomaly over the WNP, there would be a cyclonic anomaly northwest of the WNP. Gill's model generally can well explain the atmospheric circulation in association with a heat source in the tropics. Thus, the WNP monsoon would influence the SCS monsoon by changing the atmospheric circulation. In this study we will focus on the association between the SCS monsoon and WNP monsoon. This study will also be helpful for the better understanding of the mechanisms responsible for the intraseasonal variations of East Asian summer monsoon, which has been shown to be associated with both the WNP monsoon and the SCS monsoon.

The convective activity over the WNP starts to be enhanced around pentad 32 according to the result of Murakami and Matsumoto (1994), who suggested the WNP summer monsoon (signified by them as WNPM). Wu and Wang (2000) also examined the onset of WNPM, and showed the onset is around pentad 40, much later than that shown by Murakami and Matsumoto (1994). The difference is mainly due to their different definitions on WNPM onset. Apparently, the onset of WNPM is much later than that of SCS monsoon. However, as a huge heating source with great interannual and intraseasonal variabilities, after onset, the WNP monsoon may exert its influence westward to the SCS monsoon, as well as northward to the East Asian summer monsoon.

Section 2 describes the data and model utilized in this study. Section 3 shows the relationship on the interannual timescale between the WNP monsoon and SCS monsoon. Section 4 is devoted to the different features of seasonal cycle between strong and weak WNP monsoon. A summary of main results is given in the final section.

## 2. Data and model

The National Centers for Environmental Prediction / National Center for Atmospheric Research (NCEP / NCAR) re-analysis data and satellite-observed outgoing longwave radiation (OLR) data are used in this study. Monthly and pentad averages for the 20-year period from 1979 to 1998 are used. Among the elements in NCEP / NCAR re-analysis data, the 850-hPa zonal wind and surface temperature are used in this study.

The model used in this study is the UK Universities' Global Atmospheric Modelling

Programme General Circulation Model (UGAMP GCM), which is based on the forecast model of the European Centre for Medium Range Weather Forecasts (ECMWF). It is a spectral model with a hybrid coordinate in the vertical, using a triangular truncation (T42L19). The physical parameterizations in the model are described in Slingo et al. (1994). 5 of 19 vertical levels of the model are within the lowest 150 hPa of the atmosphere. The land surface temperature and moisture content are calculated using a three-layer diffusive model. A no-flux boundary condition, which allows the surface temperature and soil moisture to respond fully to the forcing rather than being tied to the imposed climatology, is used at the bottom of the soil model. The use of the no-flux boundary condition improves the simulations on the climatology and interannual variability of the Asian summer monsoon circulation (Dong and Valdes, 1998).

The SCS monsoon is characterized by active convection and lower-level zonal winds, both of which are the objectives of many recent studies (Chen et al., 2000a; Chen et al., 2000b; Chen et al., 2000c; Dai et al., 2000; He et al., 2000; Jian et al., 2000; Mu and Li, 2000; Sun and Chen, 2000; Zhu and Xu, 2000). These two variables are also widely used to define the onset of SCS monsoon (e.g., Dai et al., 2000). In the present study, we also focus on these two variables.

## 3. Relationship in the interannual variability

### 3.1 *Composite analysis based on convection over the WNP*

During the past decade, there have been many studies on the interannual variation of the onset of SCS monsoon. Recently, Wu and Wang (2000) investigated the interannual variability of the onset of WNP monsoon. Therefore, we can compare the interannual variation of these two monsoons based on the results of these previous studies.

According to Chen et al. (2000c), the years for earlier SCS monsoon onset are 1981, 1984-1986, and 1996, and the years for later onset are 1983, 1987, 1991-1993, and 1995. These years are consistent with the extreme years of WNP monsoon (Fig. 4a of Wu and Wang (2000)), i.e., earlier (later) onset of SCS monsoon is followed by earlier (later) onset of WNP monsoon.

However, the definitions on the onsets of both SCS and WNP monsoons are somewhat subjective. Thus, the onset dates may be different in the different studies. The existence of such differences makes it difficult to clarify the association between the onsets of SCS and WNP monsoons. Therefore, in the following, we examine the interannual variability in the intensity of convection and zonal wind. Lu (2001b) investigated the differences in atmospheric circulation between weaker and stronger convection over the WNP, or more exactly, the region (110°—160°E, 10°—20°N), where the June-July-August (JJA) mean convection shows a great year-to-year variability. He chose the years within which the absolute values of the JJA mean OLR anomalies are greater than half of standard deviation, and performed a composite study based on these years. The years with a greater OLR (weaker convection) are 1979, 1980, 1983, 1987, 1993, 1995 and 1996 (7 years), and the years with a smaller OLR (stronger convection) are 1981, 1984, 1985, 1986, 1989, 1990 and 1994 (7 years). He found that the convection anomalies are significantly associated with the zonal displacement of the North Pacific subtropical high and the strength of zonal winds north and south of the high.

However, Lu (2001b) did not discuss the associated changes over the SCS. In this study, a similar approach is used to examine the association between the SCS and the WNP monsoons. A region (130°—160°E, 10°—20°N), rather than the region (110°—160°E, 10°—20°N) in Lu (2001b), is used as the WNP in this study, in order to distinguish clearly the SCS and WNP. Actually, Wu and Wang (2000) defined the WNP as the region of 120°—160°E, 10°—20°N, which is adjacent to the SCS. According to the interannual variation of JJA mean convection averaged over the WNP (not shown), the years for weaker convection are selected as: 1979, 1980, 1983, 1987, 1993, 1995, and 1998 (7 years), and the years for stronger convection as: 1981, 1984–1986, 1989, 1990, and 1997 (7 years). The criterion of selection is the same as that in Lu (2001b) except for the average region. These years are very similar to those in Lu (2001b), showing that the change in the average region does not result in a significant difference in the interannual variation of the JJA mean convection over the WNP. In addition, these years also bear some similarity with the years of extreme SCS monsoon onset (Chen et al., 2000c), i.e., the weaker convection over the WNP is associated with the later onset of SCS monsoon, and vice versa.

A composite analysis is performed on these 7 years of weaker convection and 7 years of strong convection. Figure 1 shows the composite difference in OLR between the years of weaker and stronger convection over the WNP (130°—160°E, 10°—20°N). The convection over the East Asian summer monsoon region is negatively correlated to the convection over the WNP, which is consistent with the previous studies. Although the composite analysis is performed according to the interannual variation of convection over the WNP, the composite difference extends westwards remarkably from the WNP to the SCS, suggesting the existence of consistence between the interannual variability in convection over the WNP and the SCS.

Figure 2 shows the composite difference of zonal velocity at 850 hPa between weaker and stronger convection over the WNP. Significant easterly difference appears exactly over the WNP and extends westwards into the Arabian Sea through the SCS, indicating that the SCS monsoon, which is closely associated with lower–level zonal winds, is possibly related to convection over the WNP. In addition, there is a westerly anomaly in the subtropical northwestern Pacific.

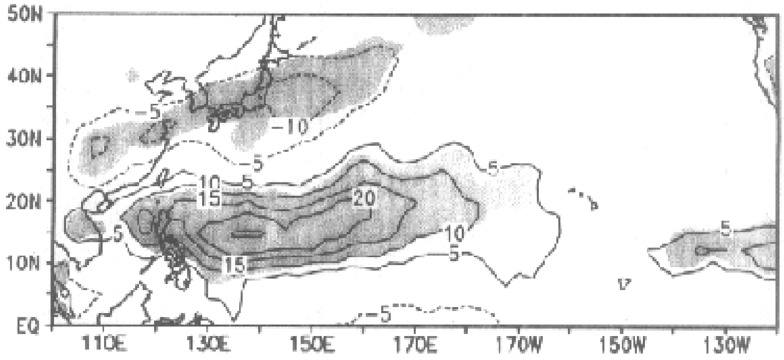


Fig. 1. Composite difference of OLR between the weaker and stronger convection over the tropical western North Pacific (130°—160°E, 10°—20°N). Unit is in W m$^{-2}$. Contour interval is 5, and zero contour is not shown. The shading illustrates the significance of differences at 95% level.

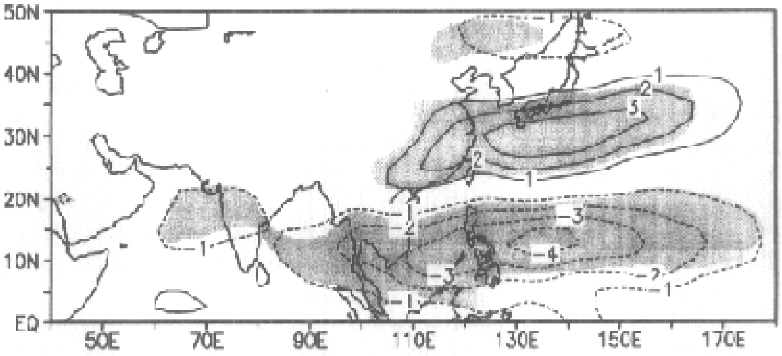


Fig. 2. Same as Fig. 1, but for 850 hPa zonal velocity. Unit in m / s. Contour interval is 1, and zero contour is not shown. Note that the longitude scope is also different.

### 3.2 *SST anomalies associated with the WNP monsoon and the SCS monsoon*

Figure 3 shows the precursory and simultaneous signals in SST differences between weaker and stronger convection over the WNP. For the convenience of comparing with previous results, the SST differences in preceding November, in January, in April and in July are given, respectively. In preceding November, there is a negative anomaly in the WNP, and a positive anomaly in the equatorial central and eastern Pacific (Fig. 3a). The anomalous SST pattern becomes slightly more significant in January in both the WNP and the equatorial central and eastern Pacific (Fig. 3b). In April, the negative SST anomaly in the WNP becomes more significant, while the positive anomaly in the equatorial central and eastern Pacific becomes less significant (Fig. 3c). The anomalous SST pattern becomes very weakly significant in July in both the WNP and the equatorial central and eastern Pacific, while there are significant positive anomalies in surface temperature west to the Philippines (Fig. 3d). Such a spatial and temporal distribution of SST anomalies is very similar to the results of Chen et al. (2000c) (their Fig. 3) and of Wu and Wang (2000) (their Fig. 5). This similarity indicates that the weaker WNP monsoon, and later onsets of WNP and SCS monsoons are all related to the El Niño–pattern SST anomalies.

Wu and Wang (2000) showed in their Fig. 7c that when the SSTs exhibit El Niño pattern in spring, suppressed convection appears around the Philippines in May, moves east of the Philippines in June, and further eastwards to around 150°–170°E in July. They pointed out that the suppressed convection in July indicates the delayed seasonal migration of convection over the WNP. We suggest that the suppressed convection around the Philippines in May is related to the delayed onset of SCS monsoon. Thus, the seasonal migrations of both SCS and WNP monsoons are all associated with similar precursory tropical SST signals, which also indicates the existence of association between the SCS and WNP monsoons.

### 3.3 *A numerical experiment*

To investigate whether the convection anomalies over the WNP influence the lower–level zonal flows over the SCS, we design a numerical experiment by UGAMP GCM in the following. The control simulations are with the AMIP climatological SST, averaged over 10 years (1979—1988). The last 10 years in total 11 years of integration are used as the model control climatology. The anomalous simulations are performed with negative SST anomalies of 1°C in the WNP (10°—20°N, 110°–160°E). There are six independent integrations with initial data from March 27 to April 1 of the 11th year in the control simulations, respectively.

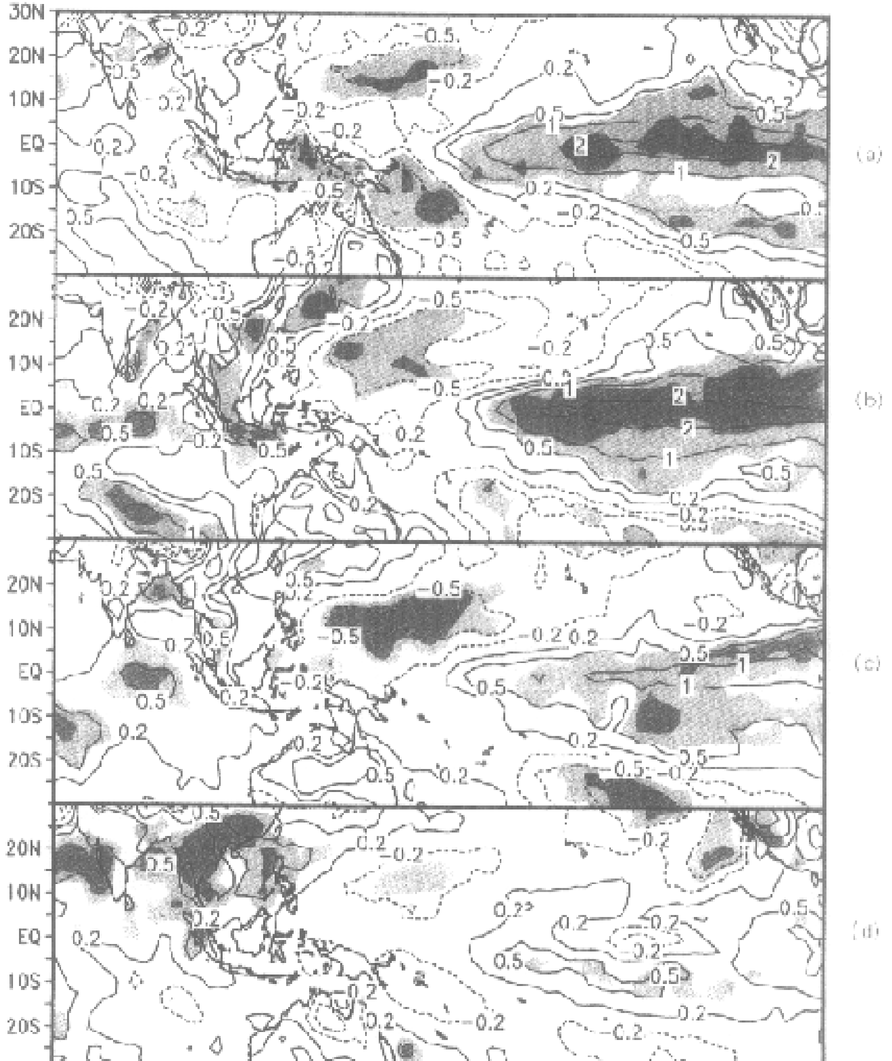


Fig. 3. The composite difference of SSTs from the preceding winter to summer between the weaker and stronger convection over the tropical western North Pacific in preceding November (a), January (b), April (c), in July (d). Contour lines are ± 0.2°C, ± 0.5°C, ± 1°C and ± 2°C, respectively The light and heavy shadings illustrate the significance of difference at 95% and 99% level, respectively.

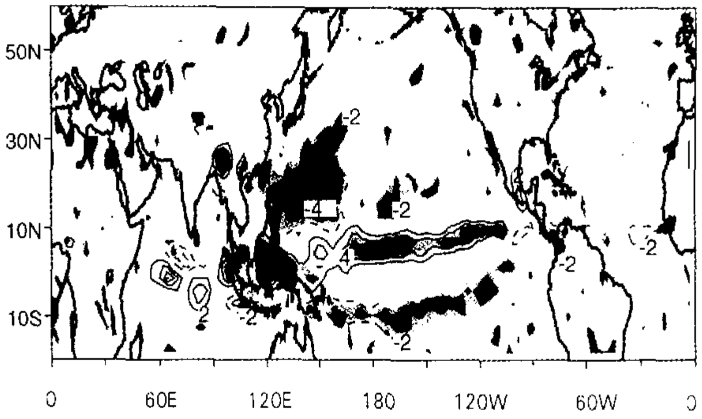


Fig. 4. The precipitation anomalies between the sensitive simulations with negative SST anomalies of 1°C in the region (10°–20°N, 110°–160°E) and the control. The shading illustrates the significance at 90% level (light shading) and 95% level (heavy shading).

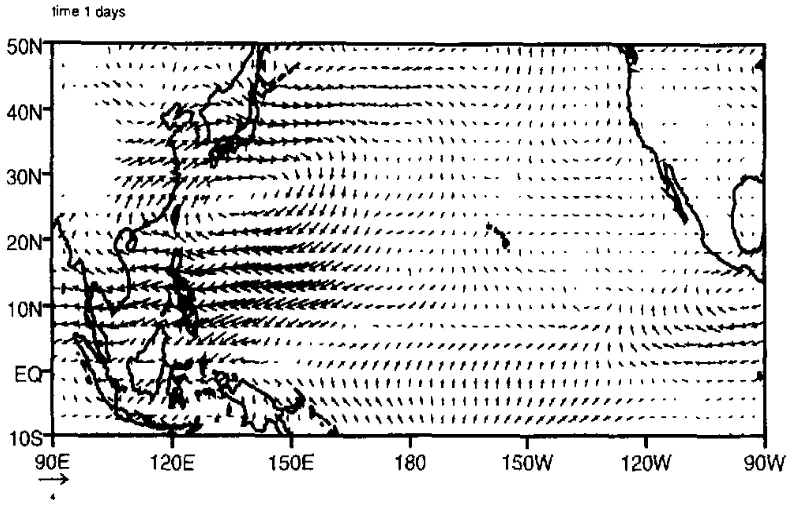


Fig. 5. The anomalies of horizontal winds at 850 hPa between the sensitive simulations and the control.

April and May are left around for spin–up, and the JJA mean differences between the sensitive and the control runs are examined. Figure 4 shows the difference in total precipitation. The precipitation is significantly less at the WNP due to the lower SSTs over there. Although the artificial SST anomalies are put in the SCS as well as in the WNP, there are no significant precipitation anomalies in the SCS. The less significant precipitation in the WNP indicates significantly weaker convection above there, since the convective rainfall predominates overwhelmingly total precipitation in the tropics. This result indicates that the artificial

SST anomalies in the WNP do change the convection over there.

Due to the lower SSTs and weaker atmospheric convection above, there are significant changes in the atmospheric circulation over the WNP. There is an anticyclonic anomaly at 850 hPa (Fig. 5), i.e., an easterly anomaly and a westerly anomaly in the tropical and subtropical western Pacific, respectively. These simulated anomalies of horizontal winds are consistent with the results of Gill (1980), and are in a fairly agreement with the composite difference in zonal wind based on observation data (Fig. 2). The easterly anomaly extends into the Bay of Bengal, i.e., extends remarkably westward comparing to the SST anomaly and resultant convection anomaly (Fig. 4), while the convection anomaly is concentrated over the WNP. The simulated results confirm that the convection anomalies over the WNP influence the lower–level zonal wind over the SCS.

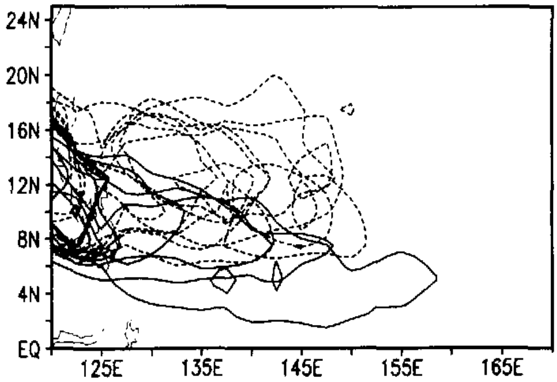
## 4. Different features in seasonal evolution

Figure 6 shows the contour line of 205 W $m^{-2}$ of composite OLR for climatology, stronger and weaker convection over the WNP, respectively, for each pentad from pentad 31 (May 31 to June 4) to pentad 49 (August 29 to September 2). Fig. 6a shows that in the climatological sense, convection activity is distinctly different over the WNP before and after pentad 40, which is the time for the onset of WNP monsoon climatologically. Here we adopted the result of Wu and Wang (2000), rather than that of Murakami and Matsumoto (1994), since the threshold value of OLR is somewhat larger in Murakami and Matsumoto (1994). After pentad 40 the convective activity shifts poleward remarkably over the WNP, and the contour line of 205 $Wm^{-2}$ encloses a larger area, indicating that the convection is enhanced.

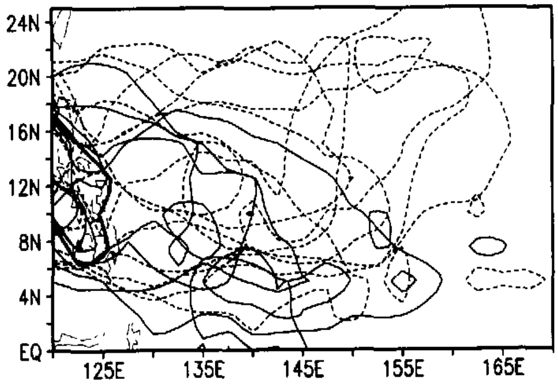
The composite seasonal evolution of convection activity over the WNP exhibits contrast features during the summers of stronger and weaker JJA mean convection. Fig. 6b shows the composite seasonal evolution of convection over the WNP during the summers of stronger JJA mean convection. It is similar to the climatological evolution (Fig. 6a), i.e., the convective activity shifts poleward and is enhanced after pentad 40. However, the extent of the shift in Fig. 6b is much more remarkable than the climatology. The seasonal evolution of convection during the summers of weaker convection also exhibits a jump around pentad 40 (Fig. 6c). This jump, however, is much weaker, in comparison with Fig. 6a for the climatology, especially with Fig. 6b for stronger convection.

Corresponding to the seasonal evolution of convection over the WNP, the 850 hPa zonal wind experiences a rapid change around pentad 40 over the SCS and WNP (Fig. 7). The climatological seasonal evolution (Fig. 7a) shows that the westerly increases after pentad 42 over the SCS and WNP. The seasonal evolution of zonal wind exhibits a distinct difference between stronger and weaker convection (Figs. 7b and c). During stronger convection summer, the westerly is much stronger over the SCS and WNP than that during weaker convection summer.

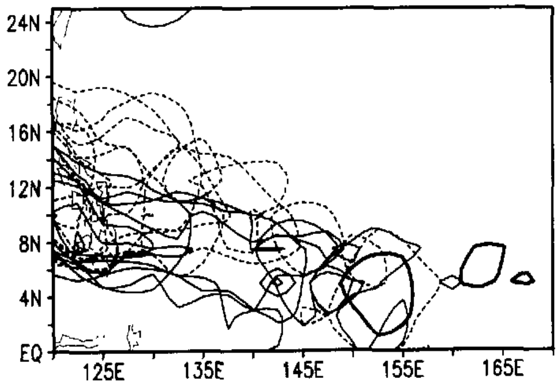
Figure 8 shows the seasonal evolution of zonal wind averaged over 5°–20°N, the latitudes of SCS, for stronger and weaker convection, respectively. Around late July, corresponding to rapid enhancing of convection over the WNP, the zonal wind increases swiftly over the SCS and WNP. Such increase in zonal wind is much more remarkable and occurs earlier for stronger convection than for weaker convection. In addition, the zonal wind over the SCS is considerably greater for stronger convection than for weaker convection, especially after late July.



(a) Climatology

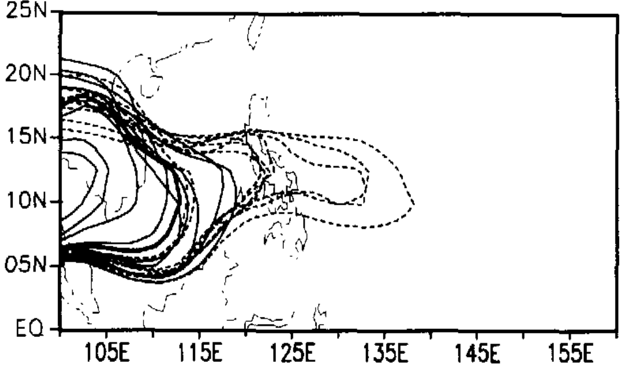


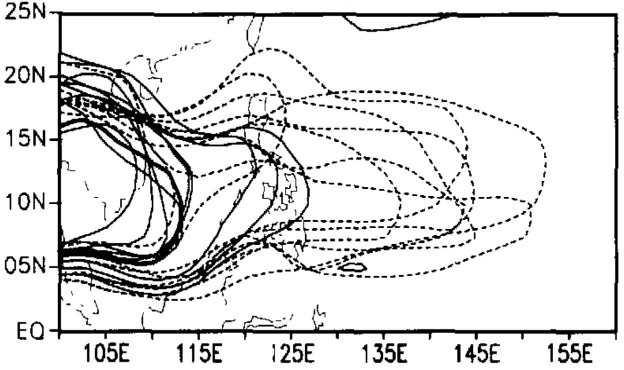
(b) For Strong Convection



(c) For Weak Convection

Fig. 6. Contour lines of 205 W $m^{-2}$ of pentad mean OLR for climatology (a), for the composites of weaker (b) and stronger (c) convection summers, respectively, for each pentad in summer, i.e., from pentad 31 (May 31 to June 4) to pentad 49 (August 29 to September 2). One contour line for each pentad. These contour lines delineate well the intensity and position of convective activity over the western Pacific in summer. Thin solid lines are for pentad 31 (May 31 to June 4) to pentad 39 (July 10 to 14), thick solid lines for pentad 40 (July 15 to 19), and dashed lines for pentad 41 to pentad 49 (August 29 to September 2).


(a) Climatology


(b) For Strong Convection

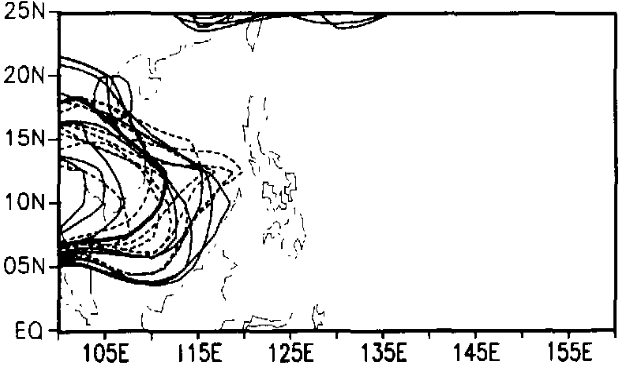

(c) For Weak Convection

Fig. 7. Same as Fig. 6, but for zonal velocity and contour lines of 5 m / s.

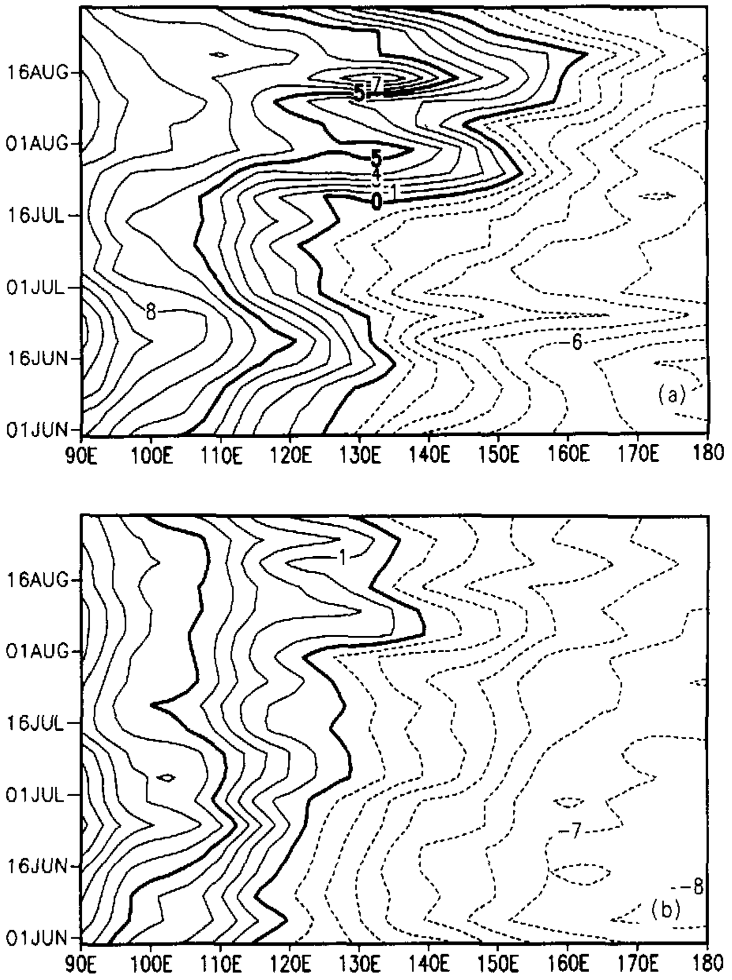


Fig. 8. Seasonal evolution of the zonal winds averaged over 5°–20°N. (a) for stronger convection, (b) for weaker convection. Unit is in m / s. The thick lines indicate the contour lines of zero and five, respectively.

## 5. Conclusions

Using the NCEP / NCAR reanalysis data and satellite–observed OLR data and a GCM, in this study we examine the relationship between the WNP monsoon and SCS monsoon. A composite analysis based on the JJA mean convection averaged over the WNP (130°—160°E, 10°—20°N) shows that the weaker (stronger) convection over the WNP is related to the El Niño (La Niña)–pattern SST anomalies in the preceding winter and in spring. A similar pattern of the tropical SST anomalies is also associated with the onsets of WNP and SCS monsoons, according to a comparison between the present and previous results. This similarity implies that there is a relationship between the SCS and WNP monsoons, and that both the monsoons are influenced by the precursory SSTs in the tropical Pacific.

The composite analysis also shows that the weaker (stronger) convection over the WNP corresponds to the easterly (westerly) anomalies in situ. These zonal flow anomalies extend westward into the Bay of Bengal. Ensemble simulations by using an AGCM show that the lower SSTs in the WNP lead to weaker convection in situ and lower–level easterly anomalies over the SCS and the Bay of Bengal, and thus confirm the results based on observation data.

The seasonal cycle shows a remarkable difference between summers with weaker and stronger convection over the WNP. During weaker convection summer, the seasonal evolution of convection over the WNP and 850 hPa zonal wind over the SCS is weak. During stronger convection summer, by contrast, the seasonal evolution of convection over the WNP and 850 hPa zonal wind over the SCS is much more remarkable.

The following description may be synthesized from the results obtained from this study. The preceding SST annomalies in the tropical Pacific influence the seasonal cycle of the SCS and WNP monsoons during summer. The strong heating over the WNP after the onset of WNP monsoon influences the convection and lower–level circulation over the SCS, which is consistent with the numerical results obtained by Gill (1980) from a simple model. The anomalous seasonal evolution is closely associated with the year–to–year anomalies of the SCS and WNP monsoons. The seasonal cycle has advanced during a summer of strong monsoon, and is delayed during a summer of weak monsoon.

This study shows that the SCS monsoon is associated with the WNP monsoon, which plays a crucial role in influencing the East Asian summer monsoon. Therefore, the association between the SCS and East Asian monsoons may be caused to some extent by the effect of the WNP monsoon.

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# 热带西太平洋夏季风和南海夏季风之间的联系

陆日宇　　Chan–Su Ryu　　Buwen Dong

## 摘　要

根据热带西太平洋(130°—160°E, 10°—20°N)上空对流的年际变化, 对表面温度、向外长波辐射、850 hPa 纬向风进行了合成分析。合成分析结果表明, 热带西太平洋上空的弱(强)对流对应着前冬和春季厄尔尼诺(拉尼娜)型的海温异常。与以前的研究结果进行了比较, 说明上述海温异常的时空分布也与热带西太平洋和南海季风的爆发早晚相关联。合成分析结果还表明, 热带西太平洋上空的弱(强)对流对应着从热带西太平洋向西伸展到孟加拉湾的东风(西风)异常。数值模拟也得到类似的结果。此外, 在对流弱(强)的夏季, 热带西太平洋上空的对流和南海低层纬向风均表现出弱(强)的季节演变特征。

**关键词**: 南海, 低层纬向风, 热带西太平洋, 对流, 厄尔尼诺