Interannual Variation of the Late Spring–Early Summer Monsoon Rainfall in the Northern Part of the South China Sea

Tsing-Chang Chen and Wan-Ru Huang*

Department of Geological and Atmospheric Sciences, Iowa State University, Ames, Iowa

Ming-Cheng Yen

Department of Atmospheric Sciences, National Central University, Chung-Li, Taiwan

(Manuscript received 7 July 2010, in final form 1 February 2011)

ABSTRACT

Major rainfall (≥60%) in the northern part of the South China Sea (between North Vietnam and Taiwan) during May–June (the mei-yu season—the first phase of the Southeast–East Asian monsoon) is produced by rainstorms originating over the northern Vietnam–southwestern China region and the northern part of the South China Sea. As observed in this study, the occurrence frequency of rainstorms and rainfall contribution by these rainstorms undergoes a distinct interannual variation, in-phase with those of monsoon westerlies in northern Indochina and sea surface temperature (SST) anomalies over the NOAA Niño-3.4 region ΔSST (Niño-3.4). This in-phase relationship between monsoon westerlies and the ΔSST (Niño-3.4) anomalies is a result of the filling (deepening) of the subtropical Asian continental thermal low in response to the ΔSST (Niño-3.4) warm (cold) anomalies. Accompanied with this response is a slight southward (northward) shift of the North Pacific convergence zone (NPCZ), which extends from southern China to the North Pacific east of Japan. Thus, a favorable environment that meets the Charney–Stern instability criterion in initiating rainstorm genesis is enhanced (suppressed) by the intensification (weakening) of the monsoon shear flow formed by the midtropospheric northwesterly flow around the northeast periphery of the Tibetan Plateau and the monsoon westerlies. The meridional shift of the NPCZ established an elongated anomalous convergence (divergence) zone of water vapor flux along rainstorm tracks to increase (reduce) the rain-producing efficiency of rainstorms. Consequently, this interannual rainfall variation between northern Vietnam and Taiwan is primarily caused by rainstorm genesis and rain-producing efficiency.

1. Introduction

The summer monsoon rainfall in East–Southeast Asia is characterized by the mei-yu rainband around the northwestern rim and the monsoon trough rainband (stretching from northern Indochina, across the northern part of the South China Sea, to the Philippine Sea) around the southwestern rim of the western Pacific subtropic high (e.g., Chen and Murakami 1988). Analyzing station observations, Ramage (1952) suggested the monsoon life cycle in the northern part of Southeast Asia and the southern part of East Asia is established by a transition from the early summer mei-yu regime into the late summer tropical cyclone season through a break in monsoon rains during late June and early July. A depiction of this monsoon life cycle with Taiwan rainfall is presented in Fig. 1a [a modification of Chen et al.’s (2004) Fig. 6a]. The transition of monsoon rainfall regimes is caused by the sequential passage of the mei-yu rainband in early summer, the western Pacific subtropic high in midsummer, and the tropical cyclone activity in late summer. Dynamically, the northward migration of these monsoon elements is caused by the coupling of the Asian monsoon circulation with the eastward propagation of the global intraseasonal mode (Chen and Murakami 1988). This coupling is reflected by a distinct out-of-phase intraseasonal

* Current affiliation: Research Fellow, Guy Carpenter Asia-Pacific Climate Impact Centre, School of Energy and Environment, City University of Hong Kong, Hong Kong, China.

Corresponding author address: Tsing-Chang (Mike) Chen, Atmospheric Science Program, Department of Geological and Atmospheric Sciences, 3010 Agronomy Hall, Iowa State University, Ames, IA 50011.

E-mail: tmchen@iastate.edu

DOI: 10.1175/2011JCLI3930.1

© 2011 American Meteorological Society
oscillation between the mei-yu rainband along the Yangtze River Valley and the monsoon rainband across the South China Sea (Chen et al. 2000). In addition to this intraseasonal oscillation, it was observed by Samel et al. (1995) that interannual variation of rainfall in East Asia is centered over the Yangtze River Valley and southeastern China. In the former region, rainfall intensifies when the interaction between major elements of the eastern Asian monsoon circulation causes the temperature gradient across the mei-yu rainbelt to increase. The rainfall in southeastern China intensifies, when the monsoon continental thermal low moves to the north.

According to Ramage’s (1952) observation, the monsoon rainfall in Southeast–East Asia is primarily produced by fronts during the first phase of the monsoon life cycle (the mei-yu season) and by tropical cyclones during the revival phase. Using the global analyzed precipitation, including the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) Project (Xie and Arkin 1997), the Global Historical Climatological Network (GHCN) station rainfall (Easterling et al. 1996), and Japan Meteorological Agency (JMA) 6-h surface analysis maps, Chen et al. (2004) showed the mei-yu rainband coincides with major frontal activity from southern China to southern Japan during the mei-yu season. Following the identification approach of different weather systems across Taiwan adopted by Wang and Chen (2008), rainfall produced by these weather systems was measured by the rain gauge networks available—the Automatic Rainfall and Meteorological Telemetry system (Chen et al. 1999) and conventional surface stations of the Weather Bureau and the Environmental Protection Administration in Taiwan. A histogram of rainfall averaged over the southwestern region (21.7°–24.3°N, 120°–120.8°E) of Taiwan is shown in Fig. 1b. Over half of the rainfall is produced by rainstorms during the active monsoon phase, while slightly less than half of the rainfall is produced by tropical cyclones during the revival monsoon phase.

Performing an empirical orthogonal function analysis of station rainfall over China, Samel et al. (1995) identified two regions of large interannual variability—the Yangtze River Valley and southeastern China. Can this interannual variation of monsoon rainfall be reflected by the activity of rain-producing weather systems? Although some extensive analysis is lacking, a severe heavy rainfall event and a wet mei-yu season that occurred along the Yangtze River Valley in 1991 (Wang et al. 2000) were observed, following the development of a major warm ENSO episode. This unusual rainfall season over the Yangtze River Valley seems to suggest an answer to the question raised above. In contrast, the possible interannual variations of rain-producing weather systems in the northern part of the South China Sea (SCS) and southeastern China and their causes have not been explored so far. As inferred from contributions of different weather systems during the mei-yu season (shown in Fig. 1b), the interannual variation of monsoon rainfall during the mei-yu season is likely attributed to the interannual variations of rainstorm populations and rain-producing efficiency between the southwestern China–northern Indochina region and southern Japan.

An effort was made in this study to explore the interannual variation of this rainstorm activity and its contribution to the monsoon rainfall during the mei-yu season over the northern part of Southeast Asia and the southern part of East Asia for the period of 1979–2008. Because the analysis period covers three decades, the rainfall and reanalyses used were derived from multiple data sources described in section 2. The interannual variations of rainfall and population and rain-producing efficiency of rainstorms over the concerned region are presented in section 3. The cause of these interannual variations and its link

---

**Fig. 1.** (a) The East–Southeast Asian summer monsoon life cycle [active phase (1/5–15/6), break phase (16/6–15/7), and revival phase (16/7–15/9)], depicted with the climatological 15-day mean rainfall for Taiwan from April to October [adopted from Chen et al. (2004)], and (b) the rainfall for Taiwan contributed by different weather systems [including diurnal variation (localized convection), typhoons, rainstorms, and fronts] for different phases of the summer monsoon life cycle, measured with Wang and Chen’s (2008) approach and averaged only over the southwestern region (21.7°–24.3°N, 120°–120.8°E) of Taiwan.
with interannual variations of monsoon westerlies in this region and the associated anomalous monsoon circulation are analyzed in section 4. The interannual variation of the Asian monsoon is often depicted by the change in rainfall. Thus, the rainstorm contribution to the monsoon rainfall in Southeast–East Asia during the mei-yu season adds a new perspective to the interannual variation of the Asian monsoon. The mechanism of this monsoon variation is often tested with global climate models (Sperber et al. 2001). The role played by rainstorms in this monsoon variation offers a new dimension to this effort. A summary of this study and remarks about the climate simulations during the Asian monsoon region are provided in section 5.

2. Identification and composite rainfall produced from rainstorm and data

Interannual variation of monsoon rainfall in the northern Southeast Asia–southern East Asia region contributed by rainstorms for the 1979–2008 period is the focus of the present study. Thus, the data are essentially determined by how rainstorms are identified during late spring–early summer. The criteria to identify a rainstorm are outlined below.

(i) A rainstorm is synoptically coupled with (or ahead of) a midtropospheric (700–600 hPa) subsynoptic-scale trough (indicated by a thick short line in Fig. 2a), slightly west of this rainstorm’s genesis location, identified by a clear IR center in a Multifunctional Transport Satellite (MTSAT) image (Fig. 2d).

(ii) Vorticity of a rainstorm at its genesis location, usually in the midtroposphere, should have a value $\geq 2 \times 10^{-3} \text{ s}^{-1}$. This location is determined by backtracking from the identified rainstorm.

(iii) Rainfall in a rainstorm should reach an amount $\geq 50 \text{ mm h}^{-1}$ at the mature stage of its life cycle.

(iv) Underneath the midtropospheric subsynoptic-scale trough, a cyclonic vortex in the lower troposphere can be identified either with the surface analysis (Fig. 2b), the JMA surface analysis (Fig. 2c), or the QuikSCAT image (Fig. 2e) close to the date of rainstorm genesis.

(v) Regardless of its genesis location (either in the northern Indochina–southern China region or the northern South China Sea), the trajectory of a rainstorm must reach the vicinity or east of Taiwan, and its life span should be $\geq 2$ days.

(vi) As measured by the subsynoptic-scale disturbance in the midtroposphere or by the cyclonic vortex in the lower troposphere, the horizontal scale of a rainstorm is $\leq O(10^3 \text{ km})$ (synoptic-scale disturbance), but $\geq O(10^2 \text{ km})$ (mesoscale convective system).

The identification genesis of a rainstorm at 1200 UTC 18 May 2008 is shown in Fig. 2 as an example.

Criteria (ii)–(vi) should be identified with analyses of surface and upper-air winds, while all criteria [(i)–(vi)] should be assessed by the following rainfall data sources and rainfall proxy—surface [World Meteorological Organization (WMO)/GHCN] station and satellite [e.g., Tropical Rainfall Measuring Mission (TRMM)] rainfall measurements, global precipitation analyses [e.g., CPC morphing technique (CMORPH)], rainfall proxy [e.g., Geostationary Operational Environmental Satellite (GOES) precipitation index (GPI)], and satellite images generated with IR or blackbody brightness temperature ($T_{BB}$) measurements. The data sources used for these two groups of criteria are designated as dataset I and II, respectively. Information regarding acronyms, spatial and temporal resolutions, spatial domain, available time period, and references of analyzed datasets are provided in Tables 1 and 2. Because the analysis period covers three decades, it is difficult (if not impossible) to compile all available data sources in a uniform manner to facilitate the identification of rainstorms. Therefore, it was required that the horizontal scale of a rainstorm occur between $O(10^3 \text{ km})$ and $O(10^2 \text{ km})$. Thus, the identification and depiction of a rainstorm need the data of horizontal resolution $\leq O(10^2 \text{ km})$. To properly use these available datasets to identify rainstorms over three decades, the manual identification procedure is adopted, which is a time-consuming task.

Since rainfall produced by a rainstorm ($P_{RS}$) is crucial information needed to examine the major theme of this study, the $P_{RS}$ distribution is prepared using the following steps:

1) The identified rainstorm may start to merge with an East Asian front 2–3 days on average after its genesis. As long as the vortex of this rainstorm can be recognized, the rainfall associated with this vortex is included in the $P_{RS}$ distribution.

2) The time for most rainstorms to propagate from their genesis locations either over northern Vietnam or the northern part of the South China Sea to Okinawa is about 5 days. The daily location of an average rainstorm center on a given day of its life cycle is the average of all rainstorms (belonging to the same genesis time window of either evening or early morning) with a given day in their life cycles.

3) Daily rainfalls produced by all rainstorms with their centers located around the daily-averaged center (red dot) on the same given day of their life cycles are summed up.

4) The $P_{RS}$ distribution is then obtained by piecing together the composite daily rainfall of all rainstorms.
obtained in step 3. The process of this step will be illustrated later by examples shown in section 3b(2).

5) The daily rainfall histogram of the composite rainstorm on a given daily is the area average of the daily ensemble rainfall coverage of this day obtained in step 3. This daily rainfall histogram is used to define the intensity of the composite rainstorm life cycle.

3. Interannual variation of rainfall: Contribution of rainstorms

After the rainstorms were identified by criteria presented in section 2, the backtracking procedure was used to track/locate genesis locations of all rainstorms from 1979 to 2008 (Fig. 3a). They are marked by red dots over the northern Indochina–southwestern China region and by blue dots over the northern part of the South China
Sea. The rainstorm genesis exhibits a location/time preference basically regulated by the day–night alternation of the land–sea thermal contrast. Land is warmer than water in the afternoon and evening but cooler than water in the early morning. Thus, most rainstorm genesis occurs over land in the former time period and over sea in the later time period. The purpose for keeping this distinction of genesis location/time is because it will be used to illustrate the environmental conditions of the rainstorm genesis over land and sea later in section 4. Trajectories of these rainstorms are depicted by color lines corresponding to their genesis locations. As indicated by their trajectories, all rainstorms move from genesis locations, across the northern South China Sea, South China, and Taiwan, to southern Japan. The primary rain-producing weather system in the northern part of Southeast Asia and the southern part of East Asia is a rainstorm during late spring–early summer (May–June). Following the procedure outline in section 2, the \( P_{RS} \) distribution during this phase of the monsoon life cycle is shown in Fig. 3b. It is no surprise to see that the \( P_{RS} \) distribution coincides with rainstorm trajectories.

The major contribution to the total amount of rainfall (\( P_T \)) during the late spring–early summer along the mei-yu rainband (Fig. 3c) has long been considered by the East Asian monsoon community to be produced by mei-yu fronts/mesoscale convective systems (MCSs) (e.g., Ding and Sikka 2006). This rainband stretches from northern Vietnam, across southern China and Taiwan, to southern Japan. The distribution of the ratio \( P_{RS}/P_T \) is shown in Fig. 3d. The patterns of \( P_{RS} \geq 6 \text{ mm day}^{-1} \) and \( P_{RS}/P_T \geq 50\% \) bear a close resemblance. This pattern resemblance between \( P_{RS} \) and \( P_{RS}/P_T \) indicates the primary rainfall contribution to the mei-yu rainband in its western section during the mei-yu season (May–June) is derived from rainstorms, instead of fronts.

### a. Population and rainfall statistics of rainstorms

Monsoon rainfall in southern China exhibits interannual variability, out-of-phase with that along the Yangtze River Valley (Samel et al. 1999). Because part of the southern China rainfall in summer is the result of the mei-yu rainband, the monsoon rainfall in the western section of the mei-yu rainband during the first phase of the East–Southeast Asian summer monsoon life cycle likely undergoes an interannual variation too. As the major rain producer in the western section of the mei-yu rainband, the rainstorm’s population and rainfall may also exhibit interannual variation in accord with the mei-yu rainband. To substantiate this inference, histograms of the rainstorm population (\( N_{RS} \)) over the region with \( P_{RS} \geq 6 \text{ mm day}^{-1} \), superimposed with \( \Delta \text{SST} \) (Niño-3.4) [the SST (Niño-3.4) (MJ) departure from its average over the period of 1979–2008], are shown in Fig. 4a. Following the interannual variation of \( \Delta \text{SST} \) (Niño-3.4), \( N_{RS} \) is generally larger (smaller) than its long-term mean value (>8 per season), when \( \Delta \text{SST} \) (Niño-3.4) ≥ 0.5°C (≤–0.5°C) during late spring–early summer. The coincident interannual variations of \( N_{RS} \) and \( \Delta \text{SST} \) (Niño-3.4) are substantiated by the large correlation coefficient, 0.82 (with a significant confidence level of 90%), between these two variables. The former thermal condition of SST is defined as warm late spring–early summer, while the latter one is defined as cold late spring–early summer. The possible mechanism causing this \( N_{RS} \) interannual variation will be presented in

### Table 1. Dataset-I data: surface and upper-air wind fields.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Source</th>
<th>Spatial resolution</th>
<th>Spatial domain</th>
<th>Temporal resolution</th>
<th>Data period</th>
<th>Source information</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Reanalyses and Global Forecast System</td>
<td>Goddard Earth Observing System Data Assimilation System version 5 (GEOS-5)</td>
<td>0.5°(lat) × 0.667°(lon)</td>
<td>Global domain</td>
<td>3 h</td>
<td>1979–present</td>
<td>Rienecker et al. (2008)</td>
</tr>
<tr>
<td></td>
<td>National Centers for Environmental Prediction (NCEP)–Global Forecast System (GFS)</td>
<td>0.5°(lat) × 0.5°(lon)</td>
<td>Global domain</td>
<td>6 h</td>
<td>6/6 2004–present</td>
<td>Kanamitsu et al. (1991)</td>
</tr>
<tr>
<td></td>
<td>National Centers for Environmental Prediction reanalysis II (NCEP II)</td>
<td>2.5°(lat) × 2.5°(lon)</td>
<td>Global domain</td>
<td>6 h</td>
<td>1979–present to cover GEOS-5 reanalyses missing in 1988 and 1997</td>
<td>Kanamitsu et al. (2002)</td>
</tr>
<tr>
<td>b. Surface observation</td>
<td>Quick Scatterometer (QuikSCAT)</td>
<td>12 km × 12 km</td>
<td>Global domain (over ocean)</td>
<td>Twice daily</td>
<td>July 1999–present</td>
<td>Atlas et al. (2001)</td>
</tr>
</tbody>
</table>
section 4. The area-average rainfall produced by rainstorms during May–June over the region of $P_{RS} \geq 6$ mm day$^{-1}$ is shown in Fig. 4b. The three-decadal average of $P_{RS}$ over this defined region is 7.1 mm day$^{-1}$. It is clearly revealed from the contrast between the $N_{RS}$ and $P_{RS}$ histograms that the interannual variation of $P_{RS}$ closely follows that of $N_{RS}$.

If rainstorms are the major rainfall producer along the western part of the mei-yu rainband, how much $P_T$ is contributed by rainstorms? To answer this question quantitatively, a histogram of $P_T$ averaged over the region with $P_{RS} \geq 6$ mm day$^{-1}$ during May–June is shown in Fig. 4c; the interannual variations of the $P_{RS}$ and $P_T$ histograms are coincident. The long-term average of $P_T$ is 11.5 mm day$^{-1}$, about 4 mm day$^{-1}$ larger than $P_{RS}$, but the long-term average ratio $P_{RS}/P_T$ (Fig. 4d) indicates that $P_{RS}$ contributes slightly over 62% of $P_T$—most of the monsoon rain along the western part of the mei-yu rainband. However, our major concern is whether the interannual variation of $P_T$ along this part of the mei-yu rainband is primarily caused by that of rainstorms, and, in turn, interannual variations of rainstorm population and rain production. It is clearly revealed from Fig. 4d that the ratio $P_{RS}/P_T$ always exceeds 62% during warm years and drops below this average value during cold years. These statistics indicate the interannual variation of $P_T$ along the western section of the mei-yu rainband is heavily affected by rainstorms.

Since there are few surface stations available over the ocean, the rainfall data used in this study are derived from several different sources, as described in section 2. Using different data sources and analysis algorithms, the analyzed global precipitation generated by different agencies and projects may have analysis bias, particularly over the ocean, because of a lack of station...
measurement. However, to verify results of the analysis shown in Fig. 4, surface observations made in Hong Kong and Taiwan are analyzed and presented in Fig. 5. Both islands are located in the major path of rainstorms, but stations on these two islands can only detect part of the rainstorm population because of their size. Recall the horizontal scale of rainstorm varies between $O(10^2$ km) and $O(10^3$ km). Despite the small sizes of these two islands, the numerical numbers of rainstorm passage across them include not only those with their centers moving directly through, but also those whose rain can reach these islands. Thus, observations from these two islands may not provide interannual variations of $N_{RS}$, $P_{RS}$, and $P_T$ during May–June for all rainstorms. However, distinct signals of these interannual variations are well reflected by observations made by surface stations on these two islands.

Numerical values of $N_{RS}$ detected at Hong Kong (Fig. 5a) and Taiwan (Fig. 5e) are 5.8 and 5.4, respectively. These values are smaller than $N_{RS}$ over the region with $P_{RS} \geq 6$ mm day$^{-1}$ (Fig. 4a). This numerical difference in the rainstorm population between these two islands and the western part of the mei-yu rainband is attributed to the limited propagation of rainstorms across these two islands. For convenience, designate MR, HK, and TW as the mei-yu rainband, Hong Kong, and Taiwan, respectively. Interannual variations of $N_{RS}$ (HK) and $N_{RS}$ (TW) closely follow $\Delta SST$ (Niño-3.4). Thus, $P_{RS}$ (HK) (Fig. 5b) and $P_{RS}$ (TW) (Fig. 5f) should undergo a coincident interannual variation with $P_{RS}$ (MR). In addition, the average values of $P_{RS}$ (HK) and $P_{RS}$ (TW) are 8.3 and 10.2 mm day$^{-1}$, respectively, and larger than that of $P_{RS}$ (MR), despite the fact that numerical values of both $N_{RS}$ (HK) and $N_{RS}$ (TW) are smaller than $N_{RS}$ (MR). The averaged values of $P_T$ (HK) (Fig. 5c) and $P_T$ (TW) (Fig. 5g) are 13.7 and 15.7 mm day$^{-1}$, respectively, larger than $P_T$ (MR), but both $P_T$ (HK) and $P_T$ (TW) exhibit an interannual variation coincident with the $\Delta SST$ (Niño-3.4) index, as $P_{RS}$ (HK) and $P_{RS}$ (TW) do.

Average ratios $P_{RS}$ (HK)/$P_T$ (HK) and $P_{RS}$ (TW)/$P_T$ (TW) shown in Figs. 5d and 5h, respectively, are slightly over 60%, close to that of $P_{RS}$ (MR)/$P_T$ (MR) (Fig. 4d). During the warm (cold) $\Delta SST$ (Niño-3.4) May–June, $P_{RS}$/$P_T$ at both Hong Kong and Taiwan, $P_{RS}$ (MR)/$P_T$ (MR), are always larger (smaller) than their respective average values. This contrast shows the interannual variation of $P_T$ at these two islands is primarily caused by the rainstorms. Interannual variations of $N_{RS}$, $P_{RS}$, $P_T$, and $P_{RS}$/$P_T$ analyzed with surface observations at Hong Kong and Taiwan are consistent with those over the western part of the mei-yu rainband.
b. Rain-producing efficiency of rainstorms

One may infer from the coincident interannual variations of \( N_{RS} \), \( P_{RS} \), and \( P_T \) in the northern part of the South China Sea (Figs. 4 and 5) that interannual variations of the latter two variables follow rainstorm population. However, it is observed from previous studies (Nakamura et al. 2002; Chang et al. 2002) that precipitation along the North Pacific storm track undergoes an interannual variation following the jet stream in response to the tropical Pacific SST anomalies. These studies lead to the following concern. In addition to the change in the rainstorm population, are the interannual \( P_{RS} \) and \( P_T \) variations also related to some basic features of rainstorm activity in response to the change of the large-scale monsoon circulation? To clarify this concern, several features of rainstorm activity are presented in Fig. 6.

1) GENESIS LOCATION AND TRACK

During both warm and cold May–June, locations of rainstorm genesis and the ensuing tracks are shown in Figs. 6a and 6c, respectively. The average tracks of rainstorms in these two extreme climate conditions are portrayed by thick red lines in (b) and (d). One can find that rainstorm tracks during warm years are, on average, somewhat north of those during cold years. The northeasterward migration of rainstorm tracks is a reflection of the change in the Southeast–East Asian monsoon circulation in response to the tropical Pacific SST anomalies during May–June, mainly through a dynamic process instead of a hydrological outcome.

2) RAIN-PRODUCING EFFICIENCY

Following the procedure to prepare the rainfall distribution contributed by all identified rainstorms \( P_{RS} \), the \( P_{RS} \) distribution during warm and cold May–June, \( P_{RS}(\text{warm}) \) and \( P_{RS}(\text{cold}) \), are shown in Figs. 6b and 6d, respectively. Distributions of both \( P_{RS}(\text{warm}) \) and \( P_{RS}(\text{cold}) \) coincide with the corresponding rainstorm tracks. These rainfall distributions indicate the magnitude of \( P_{RS}(\text{warm}) \) is larger than that of \( P_{RS}(\text{cold}) \). It is clearly indicated by this comparison of daily rainfall amounts that the rain-producing efficiency is larger (smaller) during warm (cold) year. To summarize this comparison with average daily rainfall amount over the composite life cycle of rainstorm, \( P_{RS}(\text{total}) \), \( P_{RS}(\text{warm}) \), and \( P_{RS}(\text{cold}) \) are 7.1, 8.9, and 5.4 mm day\(^{-1} \), respectively.

Fig. 4. (a) Occurrence frequency of rainstorms \( N_{RS} \) over the region of \( P_{RS} \geq 6 \text{ mm day}^{-1} \) during May–June for the 1979–2008 period, superimposed with the \( \Delta S\text{ST} \) (Niño-3.4) index averaged over May–June and precipitation histograms of (b) rainstorm \( P_{RS} \), (c) total rainfall \( P_T \), and (d) the ratio, \( P_{RS}/P_T \), over the region of \( P_{RS} \geq 6 \text{ mm day}^{-1} \). Averaged values of all variables are also presented in each panel. The thermal condition of SST over the NOAA Niño-3.4 area (5°S–5°N, 170°–120°W) is determined by the following criteria: warm \( \Delta S\text{ST} \) (Niño-3.4) \( \geq 0.5\)°C and cold \( \Delta S\text{ST} \) (Niño-3.4) \( \leq -0.5\)°C. The warm and cold thermal conditions are colored red and blue, respectively, on histograms of all variables shown in this figure.

Fig. 4. (a) Occurrence frequency of rainstorms \( N_{RS} \) over the region of \( P_{RS} \geq 6 \text{ mm day}^{-1} \) during May–June for the 1979–2008 period, superimposed with the \( \Delta S\text{ST} \) (Niño-3.4) index averaged over May–June and precipitation histograms of (b) rainstorm \( P_{RS} \), (c) total rainfall \( P_T \), and (d) the ratio, \( P_{RS}/P_T \), over the region of \( P_{RS} \geq 6 \text{ mm day}^{-1} \). Averaged values of all variables are also presented in each panel. The thermal condition of SST over the NOAA Niño-3.4 area (5°S–5°N, 170°–120°W) is determined by the following criteria: warm \( \Delta S\text{ST} \) (Niño-3.4) \( \geq 0.5\)°C and cold \( \Delta S\text{ST} \) (Niño-3.4) \( \leq -0.5\)°C. The warm and cold thermal conditions are colored red and blue, respectively, on histograms of all variables shown in this figure.
The rain-producing efficiency can be measured with the following ratios: $P_{RS(warm)}/P_{RS} \approx 125\%$ and $P_{RS(cold)}/P_{RS} \approx 84\%$. The difference in rain-producing efficiency between warm and cold years is about 49%. The cause of this efficiency difference can be inferred from the water vapor budget analysis.

Rainfall $P$ is maintained primarily by the convergence of water vapor flux $\frac{\partial}{\partial t}Q + \nabla \cdot (VQ) = \nabla \cdot (\nabla p + \alpha g \nabla \theta)$, where $Q$ is the water vapor flux, $V$ is the velocity vector, $p$ is the pressure, and $\alpha$ is the expansion coefficient. Evaporation is neglected in this approximated water vapor budget equation because its contribution in maintaining rainstorm rainfall is relatively minor. Note that $P$ and $\nabla \cdot Q$ are localized hydrological processes, but water vapor flux is driven by the large-scale circulation embedded by the rainstorm.

According to the approximated water vapor budget and the large-scale circulation embedded by the rainstorm. Apparently, the rain-producing efficiency of a rainstorm is vitally affected by the water vapor supply from the large-scale environmental flow.

Following the procedure in computing $P_{RS}$ histograms (Figs. 6e,f), the corresponding daily $-(\nabla \cdot Q)_{RS}$ histograms are shown in Figs. 6h–j. The chronicle evolution of the rainstorm hydrological cycle over its composite life cycle is revealed from the daily histograms of $P_{RS}$ and $-(\nabla \cdot Q)_{RS}$ over the composite life cycle of the rainstorm. The daily magnitude of $-(\nabla \cdot Q)_{RS}$ is somewhat smaller than that of corresponding $P_{RS}$. This discrepancy may be contributed by evaporation, which is part of the water vapor source neglected in the approximated water vapor budget. Comparing histograms of $-(\nabla \cdot Q)_{RS}$ shown in Figs. 6e–g, one finds $-(\nabla \cdot Q)_{RS(warm)} > -(\nabla \cdot Q)_{RS} > -(\nabla \cdot Q)_{RS(cold)}$. According to the approximated water vapor budget and

\begin{align*}
P \approx -\nabla \cdot Q,
\end{align*}

where $Q$ (water vapor flux) = \frac{1}{\rho} \int_{0}^{t} \nabla \cdot (\rho \nabla p + \rho \alpha g \nabla \theta) \, dt$.

Fig. 5. As in Fig. 4, but for (left) Hong Kong and (right) Taiwan.
FIG. 6. (a) Genesis locations (red/blue dots) and track (red/blue lines) during warm May–June for rainstorms and (b) $P_{RS}$ (warm) produced by rainstorms superimposed with average daily locations of rainstorm centers during warm May–June. (c), (d) As in (a), (b), respectively, but for cold May–June, histograms of average daily $P_{RS}$ for (e) all identified rainstorms, (f) warm and (g) cold May–June, and (h), (i), (j) as in (e), (f), (g), but for convergence of water vapor flux. Note that the average tracks in (b), (c) are portrayed by a thick red line connecting the average locations (red dots) of rainstorm centers over the composite life cycle of these rainstorms, which is depicted by the day number attached to those average locations. The center location of rainstorms included in the averaging process is covered by red-dashed ellipse and its peripheral area. The axes of these ellipses are one standard deviation of rainstorm distances from the average center parallel and perpendicular to the averaged track.
the contrast between daily \( -(\mathbf{V} \cdot \mathbf{Q})_{\text{RS}}(\text{warm}) \) and \( -(\mathbf{V} \cdot \mathbf{Q})_{\text{RS}}(\text{cold}) \) histograms, it is inferred that the rain-producing efficiency of rainstorms is related to the water vapor supply by the environment. This inference will be substantiated further in section 4b.

4. Possible mechanism

A summer cross-Pacific short-wave train around the North Pacific rim can be induced by warm sea surface temperature (SST) anomalies over the western tropical Pacific (Nitta 1987). An anomalous cyclonic cell adjacent to the western part of these SST anomalies is juxtaposed with an anomalous anticyclonic cell centered over northeastern Asia to form the so-called Pacific–Japan oscillation (PJO). The monsoon westerlies over northern Indochina may be intensified (weakened) by the PJO, while those along the Yangtze River Valley may be weakened (intensity) when the positive (negative) SST anomalies appear over the western tropical Pacific. Samel et al.’s (1999) out-of-phase interannual variation of rainfall between southern China and the Yangtze River Valley during summer is likely linked to the interannual variation of monsoon westerlies coupled with the PJO. The major rain-producing weather system before the break is the rainstorm, while those after the break are related to typhoons and the diurnal cycle. Thus, interannual variations of rainfall before and after the break should be related to different weather systems. Therefore, we shall focus on the search for the possible mechanism causing interannual variations of 1) rainstorm genesis and 2) rain-producing efficiency of a rainstorm.

a. Rainstorm genesis

The genesis mechanism of rainstorms during May–June was explored/illustrated from two perspectives: 1) synoptic environment and 2) dynamic instability. Therefore, the possible mechanism of the interannual variation in rainstorm genesis may be demonstrated through interannual variations of synoptic environment and dynamic intensity.

1) SYNOPTIC ENVIRONMENT

During May–June, the large-scale monsoon circulation over Southeast–East Asia consists of the Tibetan high in the upper troposphere and the continental thermal low in the lower troposphere surrounded by monsoon westerlies over Indochina and the ocean around the eastern seaboard of southern China. The synoptic environment favorable for the genesis of rainstorms is the formation of a strong shear by the midtropospheric northwesterly flow around the northeastern periphery of the Tibetan Plateau and lower-tropospheric monsoon westerlies across northern Indochina or the northern part of South China, as inferred from Fig. 7a. The northwesterly flow around the northeastern periphery of the Tibetan Plateau is generally the cold surge–like flow straddling a high–low couplet, coupled with an eastward-propagating synoptic-scale short wave in the upper troposphere. In contrast, after the onset of the Southeast–East Asian summer monsoon, monsoon westerlies persistently exist across Indochina, more stationary than the midtropospheric cold surge–like flow. Thus, the interannual variation of the environmental flow favorable for the formation of the monsoon shear line from northern Indochina to southern China may well be reflected by the interannual variation of monsoon westerlies.

As shown in Fig. 7a, strong 700-hPa monsoon westerlies appear over northern Indochina and extend northeastward to Japan. The variance of \( u(700 \text{ hPa}) \), \( \text{Var}[u(700 \text{ hPa})] \), in Fig. 7b exhibits a region of significant values over northern Vietnam. This variance maximum indicates the location of the \( u(700 \text{ hPa}) \) interannual variation. The time series of \( u(700 \text{ hPa}) \) (MJ) averaged over an area \((15°–25°N, 100°–115°E)\) around this variance maximum is shown in Fig. 7c by a dotted, thick solid line, along with \( \Delta \text{SST(Niño-3.4)} \), open circle, dashed thin line. Surprisingly, the interannual variation of \( u(700 \text{ hPa}) \) over northern Vietnam is in phase with \( \Delta \text{SST(Niño-3.4)} \). The contrast between \([u(700 \text{ hPa}), 15°–25°N, 100°–115°E], \Delta \text{SST(Niño-3.4)}\) (Fig. 7c) versus \( N_{\text{RS}} \) (Fig. 4a) clearly shows that rainstorm genesis occurs more (less) frequently when monsoon westerlies are strong (weak). The following two issues are raised from the coherent interannual variations between these three variables:

1) Why do interannual variations of \( N_{\text{RS}} \) and monsoon westerlies coincide?
2) How is the interannual variation of monsoon westerlies in the northern Indochina–southern China region coherently linked to \( \Delta \text{SST(Niño-3.4)} \)?

(i) Issue 1: \( N_{\text{RS}} \) versus monsoon westerlies

The relationship of interannual variations between \( N_{\text{RS}} \) and \( u(700 \text{ hPa}) \) in northern Vietnam is illustrated by the contrast between the pressure–time and latitude–time cross sections in Figs. 7d and 7e, and time series of \([u(700 \text{ hPa}), \Delta \text{SST(Niño-3.4)}]\) in Fig. 7c. Monsoon westerlies strengthen (weaken) and deepen (shallow) when \( \Delta \text{SST(Niño-3.4)} \) is \( \geq 0.5°C \) (\( \leq -0.5°C \)).\(^1\) The relationship between monsoon westerlies and \( \Delta \text{SST(Niño-3.4)} \) will be illustrated further in our answer to issue II, but the relationship between \( N_{\text{RS}} \) and monsoon westerlies is discussed here. When monsoon westerlies strengthen (weaken) and

\(^1\) \( \Delta() = (() - \overline{()}) \), \( \overline{()} \) = long-term average value of (). The long-term average in this study covers the period of 1979–2008.
deepen (shallow), the monsoon shear flow formed by monsoon westerlies and midtropospheric northwesterlies around the northeastern periphery of the Tibetan Plateau is intensified (weakened). Consequently, the intensification (weakening) of this shear north of monsoon westerlies facilitates (hinders) rainstorm genesis and leads to the increase (decrease) of $N_{RS}$.

(ii) Issue 2: Interannual variation of the monsoon circulation is reflected by monsoon westerlies

The interannual variation of monsoon westerlies should be an indicator of the interannual variation of monsoon circulation. Thus, the relationship between interannual variations of $\Delta u(700 \text{ hPa}, (15^\circ-25^\circ N, 100^\circ-115^\circ E))$ and $\Delta SST(\text{Niño-3.4})$ is explored in terms of warm and cold composite charts of eddy streamfunction anomalies at 700 hPa, $\Delta \theta_E(700 \text{ hPa})$, shown in Figs. 8b and 8c. As indicated by a heavy arrow, $\Delta u(700 \text{ hPa})$ in northern Vietnam and southern China increases (decreases) when $\Delta SST(\text{Niño-3.4}) \geq 0.5^\circ C$ ($\leq -0.5^\circ C$). It is inferred from the coincident interannual variations of $\Delta u(700 \text{ hPa}, (15^\circ-25^\circ N, 100^\circ-115^\circ E))$ and $\Delta SST(\text{Niño-3.4})$ time series shown in Fig. 7c that the anomalous monsoon circulations depicted by Figs. 8b and 8c are the response of the monsoon circulation to $\Delta SST$ anomalies.

To substantiate the aforementioned inference, the long-term-averaged ($\psi_E, \mathbf{V}$) charts covering major parts of the northern Indian Ocean and the Pacific are shown in Fig. 8a. The 700-hPa summer circulation over the Pacific and Indian oceans is characterized by the North Pacific anticyclone and the Asian continental thermal low north of the equator and the South Pacific anticyclone and the
Indian Oceanic anticyclone south of the equator. The South Asian monsoon westerlies, the East Asian monsoon southwesterlies, and the Pacific trade winds are indicated by arrows. The warm composite SST anomalies (Fig. 8b) are signified by the eastern tropical warm tongue surrounded by cold SST anomalies in the North and South Pacific and juxtaposed in the west with the warm SST anomalies over a large part of the tropical Indian Ocean, particularly in the western section. Compared to $\psi_E(700 \text{ hPa})$ (Fig. 8a), the response of the

Fig. 8. (a) Eddy of streamfunction at 700 hPa, $\psi_E(700 \text{ hPa})$, superimposed with isotachs (red) averaged over May–June for the 1979–2008 period. Based on the index of $\Delta$SST (Niño-3.4, MJ) shown in Fig. 7c, composite $\psi_E(7000 \text{ hPa})$ anomalies [$\Delta\psi_E(700 \text{ hPa})$] superimposed with $\Delta$SST anomalies for (b) warm and (c) cold events determined by $\Delta$SST (Niño-3.4) anomalies averaged over May–June. Contour intervals of $\psi_E$ in (a) and $\Delta\psi_E$ in (b),(c) are $10^6 \text{ m}^2 \text{ s}^{-1}$ and $3 \times 10^5 \text{ m} \text{ s}^{-1}$, respectively. (bottom right) Scales of $\psi_E(700 \text{ hPa})$ and $\Delta\psi_E(700 \text{ hPa})$ are shown in (a) and (b),(c), respectively. Arrows are added on $\psi_E(700 \text{ hPa})$ and $\Delta\psi_E(700 \text{ hPa})$ to indicate the flow directions.
lower-tropospheric circulation to SST anomalies during the
warm phase depicted by the composite $\Delta \psi_E$ (700 hPa)
anomalies is the poleward migration of oceanic anticyclones in the
Indian Ocean and in both the North and South Pacific. Portrayed with the lower-tropospheric
flow, the response is reflected by the tropical westerly
anomalies in the Pacific and the tropical easterly anom-
alies in the Indian Ocean, a divergence of anomalous
tropical flows over the Maritime Continent, as the surface
tropical wind changes during the warm ENSO events
(Philander 1990). The monsoon westerlies in South Asia
move northward and the monsoon southwesterlies east
of Japan are intensified. The spatial pattern of com-
posite $\Delta SST$ anomalies during the cold phase is re-
versed (Fig. 8c); the response of the lower-tropospheric
circulation to SST change during the cold phase is
opposite to the warm phase. This response is reflected by
the strengthening of the Pacific trade easterlies in the
east, the South Asian monsoon westerlies in the west, and
the convergence between monsoon westerlies and the
Pacific trade easterlies over the tropical western Pacific.
The monsoon southwesterlies in East Asia also weaken,
as indicated by the $u[700 \text{ hPa}, (15^\circ-25^\circ N, 100^\circ-115^\circ E)]$.

2) DYNAMIC INSTABILITY

Rainstorm genesis generally occurs in the mid-
troposphere (700–600 hPa) over northern Vietnam–
southwestern China and the northern part of the
South China Sea (Fig. 3) because the environmental flow
over these two regions usually meets the Charney–Stern
instability criterion (Charney and Stern 1962) during
May–June. As inferred from Fig. 7, the strengthening
(weakening) and deepening (shallowing) of monsoon
westerlies facilitate (hinder) the formation of the mon-
soon shear line and rainstorm genesis. In view of this
possible relationship between rainstorm genesis and the
environmental flow embedded by the monsoon shear
line, it is strongly suggested that the Charney–Stern in-
stability criterion is much easier to satisfy when mon-
soon westerlies are strong.

The Charney–Stern criterion includes two elements:

a) a sign change in the meridional gradient of potential
vorticity ($q$) $\partial q/\partial y$, and

b) a maximum gradient of potential temperature $\theta$
$\partial \theta/\partial y$ north of the location of the $\partial q/\partial y$ sign change
and south of the region of the negative $\partial q/\partial y$.

The composite latitude–height cross sections of $\partial q/\partial y$
and the latitudinal distribution of $\partial \theta/\partial y$ at the surface at
two longitudes cutting through northern Vietnam (at
104$^\circ$E) and the northern part of the South China Sea
(at 114$^\circ$E) are shown in Figs. 9a and 9d, respectively, when
monsoon westerlies are strong. The strength of monsoon
westerlies is *defined* by their magnitude indicated by the
$u[700 \text{ hPa}, (15^\circ-25^\circ N, 100^\circ-115^\circ E)]$ time series. Mon-
soon westerlies are defined as strong (weak) when the
$u[700 \text{ hPa}, (20^\circ N, 104^\circ E)]$ magnitude is larger (smaller)
than its mean value, plus one positive (negative) standard
deviation. It is clearly revealed from Figs. 9a and 9d that
there is a sign change of $\partial q/\partial y$ occurs at 600 hPa close to $20^\circ N$
when monsoon westerlies are strong and a maximum $\partial \theta/\partial y$
appears slightly north of the location where $\partial q/\partial y$ changes
signs. Apparently, the Charney–Stern instability criterion
is met by the environmental flow over northern Vietnam
(104$^\circ$E) and the northern South China Sea (114$^\circ$E), when
monsoon westerlies are strong (Figs. 9a,d). In contrast,
when monsoon westerlies are weak, the Charney–Stern
instability criterion can be met by the composite latitude–
height cross section of $\partial q/\partial y$ and the latitudinal dis-
tribution of $\partial \theta/\partial y$ at the surface (Figs. 9b,c), but these
meridional gradients are much smaller in magnitude.

The comparison of the Charney–Stern instability cri-
terion between strong and weak monsoon conditions
can be quantitatively assessed by the difference of this
instability criterion in Figs. 9c and 9f. It becomes clear
that the strong (weak) monsoon westerlies in northern
Indochina strengthens (weakens) the monsoon shear so
that the Charney–Stern instability criterion of the en-
vironmental flow can (cannot) be met and rainstorm
genesis can occur more (less) frequently.

b. Environment enhancing the rain-producing
efficiency

According to Chen (1985), the water vapor flux may
be separated into divergent ($Q_x$) and rotational ($Q_y$)
components,

$$ Q = Q_D + Q_R. $$

These two components of water vapor flux can be
expressed with the potential function ($\chi_Q$) and stream-
function ($\psi_Q$) of water vapor flux,

$$ Q_D = \nabla \chi_Q \quad \text{and} \quad Q_R = k \times \nabla \psi_Q. $$

Thus, the divergence of water vapor flux may be writ-
ten as

$$ \nabla \cdot Q = \nabla^2 \chi_Q. $$

The major content of water vapor resides in the lowest
layer of the atmosphere. Thus, the water vapor transport
is driven primarily by the lower-tropospheric circulation.
Because water vapor is a scalar variable, spatial patterns for
both $\chi_Q$ and $\psi_Q$ resemble those of velocity potential ($\chi$)
and streamfunction ($\psi$) in the lower troposphere.
The onset of the Asian monsoon and the active monsoon phase in the South, Southeast, and part of East Asia occur during May–June. During this time period, the global water vapor converges toward the Asian monsoon hemisphere so the global $\chi_0$ field exhibits a wave-1 structure around a latitude circle (not shown), but its Asian monsoon hemisphere portion is shown in Fig. 10a. The convergent center of $(\chi_0, Q_D)$ is located over the Indochina–South China Sea–Philippine Sea region. The distribution of rainfall radiates from the

![Figure 9](image-url)
Asian monsoon region eastward along the North Pacific convergence zone, the South Pacific convergence zone, and the intertropical convergence zone.

During warm (cold) May–June, the tropical SST anomalies exhibit positive (negative) anomalies over the tropical Indian Ocean centered at 60°E and the National Oceanic and Atmospheric Administration (NOAA) Niño-3.4 area as shown in Fig. 8b (Fig. 8c). The response of the global divergent circulation in the lower troposphere to these two tropical SST anomaly centers during warm May–June is reflected by convergent centers of $\Delta(x_0, Q_D, P)$ over the tropical Indian Ocean and the central-eastern tropical Pacific and a divergent region covering the Asian monsoon and Australia (Fig. 10b). As inferred from this low-level divergent circulation, the coupled east–west circulation in the tropics possesses the upward branches over the positive SST anomaly regions and a downward branch over the monsoon region as the tropical east–west circulation depicted by Streten and Zillman (1984). The downward branch of this east–west circulation suppresses weather development and convective activity and reduces rainfall. The reversed spatial structure of global divergent circulation and tropical east–west circulation occur during cold May–June, as inferred from Fig. 10c. The rainfall increases over the Asian monsoon region because the upward branch of the tropical east–west circulation appears there.

With the large-scale divergent circulation change resembling the $\Delta(x_0, Q_D, P)$ fields presented in Figs. 10b and 10c, in response to the SST anomalies over the tropical Indian Ocean and the NOAA Niño-3.4 region, it seems unusual to find the interannual rainfall variation $\Delta P$ over the major $P_{RS}$ region (Fig. 2b) is opposite to those over the rest of the Asian monsoon region. The disparity of these monsoon rainfall variations leads to two basic issues of the regional hydrological cycle:

1) What causes the disparity of the interannual rainfall variation between the major $P_{RS}$ region and the remaining Asian monsoon region?
2) How does the interannual variation of $(x_0, Q_D)$ facilitate or hinder the rain-producing efficiency of rainstorm?

1) ISSUE 1

Extending southward along 105°E from southwestern China to the eastern coast of Vietnam, a shallow monsoon trough indicated by a dashed blue line is discernible on the 700-hPa streamline charts (Figs. 7a,b). Overlapping with the favorable region of rainstorm genesis in southwestern China (Fig. 3a), this trough facilitates rainstorm genesis. Based on the tropical east–west circulation inferred from the $\Delta(x_0, Q_D)$ patterns (Figs. 10b,c), it is expected that the genesis/development of synoptic disturbances in the Asian monsoon region should be suppressed (enhanced) by the downward (upward) branch of this east–west circulation. However, it was indicated by both the height–$t$ diagram of $u(15°–25°N, 100°–115°E)$ (Fig. 7d) and the $y–t$ diagram of $u(700 \text{ hPa}, 100°–115°E)(\text{MJ})$ (Fig. 7e) that monsoon westerlies increase (reduce) during warm (cold) May–June in the northern Vietnam–southwestern China region. As indicated by the anomalous circulations in response to the tropical Pacific SST anomalies during May–June (Figs. 8b,c), the interannual variation of monsoon westerlies (Figs. 7d,e) is caused by anomalous anticyclonic (cyclonic) cell centered at the northern part of the South China Sea and the Philippine Sea. During warm (cold) May–June, the aforementioned anticyclone (cyclonic) cell exhibits two centers over the head Bay of Bengal and the northern part of the South China Sea; therefore, an anomalous trough (ridge) appears between these two centers. An anomalous convergent (divergent) zone of $\Delta(x_0, Q_D)$ and the accompanied

![Fig. 10](image-url)
positive (negative) zone of $\Delta P$ are collocated ahead of the aforementioned trough (ridge). This interannual variation of the large-scale divergent circulation forms a favorable (unfavorable) environment to enhance (reduce) the rain-producing efficiency of rainstorms.

2) ISSUE 2

A histogram of $-(V \cdot Q)_T$ superimposed with $P_T$ for the region of $P_{RS} \geq 6 \text{ mm day}^{-1}$ is shown in Fig. 11a, while the same for $-(V \cdot Q)_{RS}$ superimposed $P_{RS}$ is shown in Fig. 11b. These two quantities during warm and cold May–June are colored red and blue, respectively; their interannual variations are coincident with a correlation coefficient of 0.89 at a confidence level of 90%. Magnitude ratios of average $-(V \cdot Q)_T/P_T$ and $-(V \cdot Q)_{RS}/P_{RS}$ are 90% and 93%, respectively. Interannual variation magnitudes of $-(V \cdot Q)_T$ and $-(V \cdot Q)_{RS}$ may be measured statistically by their standard deviations—both are 1.5 mm day$^{-1}$. Because $-(V \cdot Q)_{RS}$ is part of $-(V \cdot Q)_T$, it is implied by such close statistical values that the interannual variation of $-(V \cdot Q)_{RS}$ is largely modulated by $-(V \cdot Q)_T$. The latter is determined by the large-scale divergent circulation through $-(V^2x_Q)_T$ because rainstorms are embedded in the large-scale divergent circulation $-(V \cdot Q)_{RS}$ and in turn the rain-producing efficiency of the rainstorm is affected by this circulation through $-(V^2x_Q)_T$.

5. Concluding remarks

The north–south extent of the eastern seaboard of Southeast–East Asia under the influence of the summer monsoon is over 4000 km, covering the tropics and midlatitudes. This long north–south distance causes the monsoon life cycle in East Asia to lag behind the monsoon life cycle in Southeast Asia. The out-of-phase intraseasonal oscillation of summer monsoon rainfall between the northern South China Sea and the Yangtze River Valley is a reflection of this phase lag (Chen et al. 2004). Thus, the rain-producing weather systems in different latitudinal zones may be different not only in the timing of their activity but also in their nature. It has long been regarded by the Asian monsoon community that rainfall along the mei-yu rainband during May–June is produced by mei-yu fronts. It was shown in this study that a major part of the May–June monsoon rainfall in the northern South China Sea, over northern Vietnam, across Taiwan, and into southern Japan is produced by rainstorms over the ocean.

It was shown by Samel et al. (1999) that the summer rainfall in South China exhibits an out-of-phase interannual variation with that in the Yangtze River Valley. An effort was made in this study to explore whether the rainfall over the northern part of the South China Sea undergoes an interannual variation during May–June, and the cause of this interannual variation. The major findings of this study are summarized as follows:

1) Interannual variation of May–June monsoon rainfall:

A distinct interannual variation emerges from the occurrence frequency of rainstorm genesis $N_{RS}$ rainfall produced by rainstorms $P_{RS}$ and total rainfall $P_T$ over the region ($P_{RS} \geq 6 \text{ mm day}^{-1}$). Interannual variations of $N_{RS}$, $P_{RS}$, and $P_T$ coincide with the interannual variation of SST over the NOAA Niño-3.4 region $\Delta SST$ (Niño-3.4); ($N_{RS}$, $P_{RS}$, $P_T$) are larger (smaller), when $\Delta SST$ (Niño-3.4) $\geq 0.5^\circ C$ (warm) $[\leq -0.5^\circ C$ (cold)].

2) Possible cause of ($N_{RS}$, $P_{RS}$, $P_T$) interannual variations: In-phase interannual variations between ($N_{RS}$, $P_{RS}$, $P_T$) and monsoon westerlies in northern Vietnam were observed. It is inferred from the in-phase...
interannual variations of monsoon westerlies over there and $\Delta$SST (Niño-3.4) that interannual variations of $(N_{RS}, P_{RS}, P_T)$ are a reflection of the interannual variation of the monsoon circulation’s impact on the genesis occurrence frequency and rain-producing efficiency of rainstorms.

\textbf{a. Genesis occurrence frequency}

\textbf{1) SYNOPTIC PERSPECTIVE}

The formation of the monsoon shear line between monsoon westerlies across northern Indochina and the midtropospheric cold surge–like northwesterly around the northeastern periphery of the Tibetan Plateau may be facilitated (hindered) by the intensification (weakening) and deepening (shallowing) of the monsoon westerlies.

\textbf{2) DYNAMIC INSTABILITY}

The Charney–Stern instability criterion was shown to be the dynamic mechanism responsible for the rainstorm genesis, which consists of a sign change of meridional potential vorticity $\partial q/\partial y$ coupled with a maximum meridional gradient of potential temperature $\partial T/\partial y$ north of the sign change of $\partial q/\partial y$. The Charney–Stern instability criterion is much easier (more difficult) to meet by the monsoon shear when monsoon westerlies are strong (weak).

\textbf{b. Rain-producing efficiency}

The response of the global divergent circulation in the lower troposphere to the SST anomalies in the tropical Indian and Pacific oceans leads to the development of the tropical east–west circulation with its upward (downward) branches over these two regions and a downward (upward) over the Asian monsoon region during warm (cold) May–June. On the other hand, the monsoon trough, extending southward from southwestern China to the coast of Vietnam, deepens (fills) during warm (cold) May–June, so that an anomalous convergent (divergent) zone of water vapor flux appears ahead of the deepened (filled) monsoon trough and is coincident with rainstorm tracks. The water vapor supply to this region is enhanced (depleted) so the rain-producing efficiency of rainstorms increases (decreases) during warm (cold) May–June.

These findings have two important implications:

\textbf{1) Seasonal prediction of the Southeast–East Asian climate:} The $\Delta$SST (Niño-3.4) index is operationally issued by the Climate Prediction Center, Washington, D.C. The in-phase relationship between $(N_{RS}, P_{RS}, P_T)$ and $\{u(\text{monsoon westerlies}), \Delta$SST (Niño-3.4)] may provide an excellent tool to monitor the interannual variation of the Southeast–East Asian monsoon.

\textbf{2) The seasonal predictions of South Asian monsoons performed by global climate model simulations have been under careful scrutiny (Sperber et al. 2001), but this research effort has not been expanded to cover the Southeast–East Asian monsoon. Findings about the interannual variations of monsoon rainfalls in Southeast–East Asia and their cause may provide a way to examine the performance of global climate models over this region.}

\textbf{Acknowledgments.} This study is partially sponsored by the Cheney Research Fund and the NSF Grant ATM-0836220 and is greatly improved with comments and suggestions offered by two reviewers. The effort made by Ming-Cheng Yen is supported by the Grant NSC99-2111-M-008-012. Technical assistance provided by Paul Tsay is crucial to the revision of this manuscript.

\textbf{REFERENCES}


