Interannual Variation of the Summer Rainfall Center in the South China Sea

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ABSTRACT

A northwest–southeast-oriented summer monsoon trough exists between northern Indochina and northwestern Borneo. Ahead of this the South China Sea (SCS) trough is located at a convergent center west of the Philippines, which provides an environment favorable for rain-producing synoptic systems to produce rainfall over this center and form the SCS summer rainfall center. Revealed from the x – t diagram for rainfall, this rainfall center is developed by multiple-scale processes involved with the SCS trough (TR), tropical depression (TY), interaction of the SCS trough with the easterly wave/tropical depression (EI), and easterly wave (EW). It is found that 56% of this rainfall center is produced by the SCS trough, while 41% is generated by the other three synoptic systems combined. Apparently, the formation of the SCS summer monsoon rainfall center is contributed to by these four rain-producing synoptic systems from the SCS and the Philippines Sea. The Southeast Asian summer monsoon undergoes an interannual variation and exhibits an east–west-oriented cyclonic (anticyclonic) anomalous circulation centered at the western tropical Pacific east of the Luzon Strait. This circulation change is reflected by the deepening (filling) of the SCS summer monsoon trough, when the monsoon westerlies south of 15°N intensify (weaken). This interannual variation of the monsoon westerlies leads to the interannual variation of the SCS summer monsoon rainfall center to follow the Pacific–Japan oscillation of rainfall. The rainfall amount produced over this rainfall center during the weak monsoon season is about two-thirds of that produced during the strong monsoon season. The rain-production ratio between TR and TY + EI + EW is 60:38 during the strong monsoon season and 47:49 during the weak monsoon season.

1. Introduction

As shown in Fig. 1b, the climate system over the northern part of Southeast Asia and East Asia during the warm season (May–September) belongs to the southwest monsoon regime (Ramage 1971). The warm-season monsoon in this region exhibits an active–break–revival cycle (Ramage 1952; Chen et al. 2004). In the southern part of East Asia, onset of the active monsoon phase occurs in mid-May, and the break monsoon phase starts in late June. Chen et al. (2011) observed a major rainbelt stretching from northeastern Vietnam across

southern China and Taiwan to southern Japan. Rainfall along this rainbelt is primarily contributed by rainstorms originating from northern Vietnam and southern China and the northern part of the South China Sea (SCS). In contrast, the existence of a rainfall center west of the Philippines over the entire summer monsoon season was identified by Chen and Chen (1995) prior to the South China Sea Monsoon Experiment (SCSMEX; Lau 1997). Since the rainfall centers and the rainbelt occur along the southern China coast during the active phase of the summer monsoon season, can the SCS warm-season monsoon rainfall center develop by certain rain-producing synoptic systems?

In the lower-tropospheric circulation of the Southeast Asian summer monsoon exists the SCS monsoon trough. This trough may deepen or fill, and its trough line may also rotate *cyclonically* along the island chain from northwestern Borneo to the Luzon Islands. Rainfall

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FIG. 1. Summer (mid-May–31 Aug) climatology for Southeast Asia over the 1979–2016 period for (a) $[V(200\text{ hPa}), P_T](15/5-31/8)$, (b) $[V(850\text{ hPa}), P_T](15/5-31/8)$, (c) $[(\chi, \mathbf{V}_D)(200\text{ hPa}), P_T](15/5-31/8)$, and (d) $(\chi_Q, \mathbf{Q}_D, P_T)(15/5-31/8)$. The color scale for P_T and the vector scale for $\mathbf{V}_D$ and $(\mathbf{Q}_D)_T$ are shown on the top of each relevant panel.

often is produced ahead of or over the SCS monsoon trough. Tropical cyclones may move across the Philippines (e.g., [Corporal-Lodangco et al. 2016](#)) and produce rainfall over the SCS summer rainfall center. The westward-propagating tropical easterly waves can also propagate across the Philippines (e.g., [Chen and Weng 1996, 1998](#)) and generate rainfall ahead of its trough line west of the Philippines. The westward-propagating easterly wave/tropical depression across the Philippines may sometimes interact with the SCS troughs to produce rainfall. Can an SCS summer rainfall center be established by the rain generated from these synoptic systems? So far, the formation mechanism of the SCS summer rainfall center has not been explored from the synoptic perspective in the past two decades. Because this rainfall center is an important part of the Southeast Asian summer monsoon, the formation mechanism of this rainfall center should play a crucial role in the development and maintenance of this monsoon circulation.

The monsoon rainbelt from the northern Vietnam/southern China region to southern Japan during the active monsoon phase undergoes an interannual variation in concert with the interannual variations of monsoon westerlies and rainstorm activity ([Chen et al. 2011](#)). It is likely that the SCS summer rainfall center also undergoes an interannual variation in phase with the

strength of monsoon westerlies through the activity of rain-producing synoptic systems across the SCS rainfall center. In other pre-SCSMEX analyses, [Lau and Yang \(1997\)](#) determined that the SCS monsoon onset date varies interannually with the strength of monsoon westerlies. This observation was further confirmed by [Zhou and Chan \(2007\)](#) with global reanalyses of NCEP and ECMWF data for the period of 1958–2002. Early monsoon onset usually leads to more monsoon rain production than late monsoon onset. The interannual variation of a monsoon is not only indicated by the strength of monsoon westerlies, but also reflected by the activity of rain-producing synoptic systems. The SCS trough deepens during the strong monsoon but fills during the weak monsoon ([Chen et al. 2017a](#)). The tropical cyclone activity and track across the Philippines (e.g., [Harr and Elsberry 1991](#); [Corporal-Lodangco et al. 2016](#), among others) are modulated by the meridional ridge location of the western North Pacific subtropical anticyclone, as shown by the Pacific–Japan (P–J) oscillation ([Nitta 1987](#)). During the summer monsoon season, the activity of the westward-propagating tropical easterly waves across the Philippines is modulated by the anomalous circulation cell coupled with the P–J oscillation ([Chen and Weng 1996, 1998](#)) to strengthen or weaken the SCS monsoon rainfall center. The

effectiveness of the interaction between tropical waves and the SCS trough is also likely affected by the variation of the SCS monsoon circulation through the deepening/filling of the SCS trough.

The major tasks of this study are 1) to search for the formation mechanism of the SCS summer monsoon rainfall center west of the Philippines and 2) to determine the cause of this rainfall center's interannual variation. This study is organized in the following manner. The rainfall data used in this study are derived from several sources. A simple approach is adopted to merge these sources into a uniform dataset for the 1979–2016 period in [section 2](#). Both the National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS) ([NCEP 2003, 2016](#)) and ECMWF-Interim reanalyses ([Dee et al. 2011](#)) are utilized for analyzing the water vapor transport and depicting the circulation structure and synoptic systems. The identification of rain-producing synoptic systems and the rainfall maintenance illustrated by the water vapor budget are also described in [section 2](#). The climatology and hydrological system of the Southeast and East Asian summer monsoons and the formation and maintenance of the SCS summer monsoon rainfall center by different synoptic systems are presented in [section 3](#). Interannual variations of the SCS summer monsoon circulation and the SCS rainfall center are presented in [section 4](#). In this section, the cause of these interannual variations is illustrated through roles played by the hydrological environment and circulations of rain-producing synoptic systems in the interannual variation of the SCS summer rainfall center and its maintenance. A summary of the formation and maintenance mechanisms of the SCS summer monsoon rainfall center and the causes of interannual variation of the SCS summer rainfall center contributed by different synoptic systems are provided in [section 5](#). Some future studies for the monsoon climate over this region developed from the perspective of weather systems are also suggested.

2. Data and identification of rain-producing disturbances and maintenance of rainfall

a. Data

Three data sources are utilized for the analysis in this study: rainfall, reanalysis, and a daily surface analysis map. Details for these data sources are provided in [Table 1](#). The analysis is performed for 38 summers during 1979–2016; thus, the data need to be compiled uniformly over this long time period. This basic requirement is met by the last two data sources, reanalysis and a daily surface analysis map, but not rainfall.

Images and blackbody brightness temperature T_{BB} from regional satellite and daily surface analysis maps issued by different weather agencies are used to verify the rain-producing systems identified by the streamline charts prepared with the reanalysis data. Data produced from two reanalyses (NCEP GFS and ERA-Interim) are used for this purpose. Streamline charts generated with former reanalysis data match the T_{BB} and/or rainfall distributions more closely in detail than the latter reanalysis. Note that the $0.5^\circ \times 0.5^\circ$ GFS reanalysis became available in 2006. Before this year, streamline charts were prepared with the ERA-Interim.

Rainfall over land is derived from two data sources [World Meteorological Organization (WMO) station measurements and Asian Precipitation–Highly-Resolved Observational Data Integration toward Evaluation of Water Resources (APHRODITE)] and over both land and ocean from another three data sources [Tropical Rainfall Measuring Mission (TRMM),¹ Precipitation Estimation from Remote Sensing Information using Artificial Neural Network (PERSIANN), and GPCP]. Periods and regions covered by these data sources are shown in [Table 1](#). Different seasonal mean values at every grid point may be generated by different rainfall datasets. If these rainfall datasets are not calibrated, some unusual interannual variations of the SCS summer rainfall center may emerge. A simple calibration procedure introduced by [Chen et al. \(2017b\)](#) is adopted to make these rainfall datasets uniform over the analysis region, particularly the SCS. This procedure includes the following steps.

- 1) Over Japan for 1998–2007, the TRMM rain $P(\text{TRMM})$ is calibrated against the APHRODITE rainfall: $P(\text{calibrated TRMM}) \simeq 1.2P(\text{TRMM})$.
- 2) For 1998–2015 over the domain $95^\circ\text{--}140^\circ\text{E}$, $5^\circ\text{S--}20^\circ\text{N}$, the PERSIANN rainfall $P(\text{PERSIANN})$ is calibrated against $P(\text{calibrated TRMM})$: $P(\text{calibrated PERSIANN}) = 1.2P(\text{PERSIANN})$.
- 3) The GPCP rainfall is calibrated against the $P(\text{calibrated TRMM})$ data for 1998–2015 over the domain $95^\circ\text{--}140^\circ\text{E}$, $5^\circ\text{S--}20^\circ\text{N}$: $P(\text{calibrated GPCP}) = 1.2P(\text{GPCP})$.
- 4) For the 1979–2016 period, the calibrated P rainfall datasets are combined over their available periods of rainfall data to form a uniform rainfall dataset; $P(\text{calibrated GPCP})$ for 1979–82, $P(\text{calibrated PERSIANN})$ for

¹ According to [Huffman and Bolvin \(2015\)](#), the TRMM Microwave Imager was closed on 8 Apr 2015, but the 3B42 version of TRMM precipitation operates in parallel with the Integrated MultiSatellite Retrievals for GPM (Global Precipitation Measurement) (IMERG).

TABLE 1. Detailed information for the datasets used in this study.

Data	Source	Spatial resolution	Spatial domain	Temporal		Source information
				resolution	Data period	
Precipitation	Surface observation	WMO surface station	Global	3 h	1979–present	http://www.ncdc.noaa.gov/cdo-web/
	Regional satellite	APHRODITE (v1204R1)	Asia (land)	Daily	1951–2007	Yatagai et al. (2012)
	and observation	Geostationary Meteorological Satellite (GMS) T_{BB}	59.5°S–59.5°N, 80.5°E–160.5°W	3 h	1980–present	Meteorological Research Institute, Japan
		GMS/ <i>GOES-9</i> /MTSAT	70°S–70°N, 80°E–150°W	1 h	1979–present	Blersch and Probert (1991) Meteorological Services Centre Japan (1997)
Global gridded precipitation	TRMM (3B42v7)	0.25° lon × 0.25° lat	59.5°S–59.5°N, 180°–180°	3 h	1998–present	Huffman and Bolvin (2015)
	PERSIANN	0.25° lon × 0.25° lat	50°S–50°N, 180°–180°	Daily	1983–present	Ashouri et al. (2015)
Global reanalyses (1979–present)	GPCP (v2.2)	2.5° lon × 2.5° lat	Global	Daily	1979–present	Huffman and Bolvin (2015)
	NCEP GFS	0.5° lon × 0.5° lat	Global	6 h	2006–present	Global Climate and Weather Modeling Branch, EMC, NCEP (2003, 2016)
Daily surface analysis map	ERA-Interim	0.5° lon × 0.5° lat	Global	6 h	1979–present	Dee et al. (2011)
	NCEP SRRS			12 h	1979–present	SRRS (http://nomads.ncdc.noaa.gov/ncep/NCEP)
	Northern Hemisphere Surface Analysis NCEP Tropical Strip Surface Analysis and Observation		Northern Hemisphere Tropical (40°S–40°N)			
	JMA		Northwestern Pacific	6 h	1979–present	http://www.jma.go.jp/en/g3/
	BoM, Australia		35°S–30°N 70°E–180°	12 h	2000–present	http://www.bom.gov.au/australia/charts/archive/

1983–97, and P (calibrated TRMM) for 1998–2016 are combined.

Note that the calibration coefficients used in this procedure are generated by the scatter diagrams: calibrated rainfall versus calibrated TRMM rainfall. Results produced from this calibration procedure are presented in the online supplemental material (see supplement 1).

b. Identification of rain-producing synoptic systems and determination of rainfall days

The SCS summer monsoon rainfall center is defined over the area covered by a threshold value of rainfall $\geq 10 \text{ mm day}^{-1}$ in the SCS west of the Philippines. This rainfall center is formed by the accumulation of rain produced by the rain-producing synoptic systems over this center. The synoptic systems in the middle and high latitudes are operationally depicted by surface pressure and upper-air geopotential height. The magnitudes of these two variables for the tropical synoptic systems are an order smaller than those for the synoptic systems in the middle and high latitudes. In contrast, the wind speeds of synoptic systems in the tropical–subtropical region are comparable. For this reason, the 925- and 850-hPa winds are used to prepare streamline charts superimposed with precipitation. These streamline charts are used to identify the aforementioned four synoptic systems, which consist of the following three steps:

- 1) Preparation of the 925- and 850-hPa streamline charts superimposed with precipitation.
- 2) Verification of the identified system with the NCEP Service Records Retention System (SRRS) tropical strip surface analysis and observational charts, and daily surface and upper-air charts issued by the Japan Meteorological Agency (JMA) and the Australian Bureau of Meteorology (BoM).
- 3) The rainfall accumulation over the SCS summer rainfall center produced by the identified rain-producing synoptic systems should reach the threshold value $\geq 10 \text{ mm day}^{-1}$ mentioned above.

The major rain-producing synoptic systems identified by this study are the following:

1) SCS trough:

The trough (TR) is oriented from northern Indochina southeastward to northwestern Borneo. This trough line may rotate cyclonically from Borneo to the Luzon Islands. Rain is usually produced ahead of or over this trough. Sometimes, a cutoff low is formed by the SCS TR over the northern SCS, but it is still included by this SCS TR group.

2) Tropical storm/typhoon:

Tropical storms (TS) and typhoons (TY) are identified/archived by the Joint Typhoon Warning Center (JTWC). Rainfall is produced by the rainbelts spiraling around the named TS/TY, but no merger of a named TS/TY with a TR is identified

3) Easterly wave/tropical depression interacting with the SCS trough:

The westward-propagating easterly wave, including the tropical depression (TD) formed by the easterly wave (EW) east of the Philippines, interacts and merges with the SCS trough west of the Philippines. Some TD may propagate westward to form a link with the SCS trough, as achieved by EW. These interactions of EW and TD with the SCS TR are classified as the easterly interaction (EI) group.

4) Easterly wave:

The easterly wave may propagate westward across the Philippines or develop into a vortex before the Philippines, but it does not interact or merge with the SCS trough.

Typical examples of the six distinctive, rain-producing, synoptic systems classified are presented in the supplemental material (see supplement 2) to save space, but are grouped into four systems to facilitate analysis: SCS TR, TS/TY, EI, and EW. Typical synoptic structures/illustrations of these rain-producing synoptic systems will be presented in the next section. The SCS summer rainfall center is established by accumulation of rainfall produced by the classified rain-producing synoptic systems over this center's domain. The rainfall contribution by any one of these four synoptic groups is the rainfall accumulation as soon as the rain produced by the concerned synoptic system reaches the rainfall center over the time period until the rain from this synoptic system leaves this center. By this approach, about 97% of rain over the SCS summer rainfall center is contributed by the four groups of major rain-producing synoptic systems shown in [section 3](#).

c. Maintenance of precipitation: Water vapor budget

According to the water vapor budget equation, several hydrological processes are involved to maintain precipitation: temporal variation of the environmental precipitable water, convergence of water vapor flux, and evaporation. Nevertheless, the most crucial hydrological process for the maintenance of precipitation is convergence of water vapor flux. To explore this precipitation maintenance, the approximated water vapor budget may be written as

$$P \simeq -\nabla \cdot \mathbf{Q}, \quad (1)$$

where P and $\mathbf{Q}$ are precipitation and water vapor flux, respectively;

$$\mathbf{Q} = \frac{1}{g} \int_0^{p_s} (q\mathbf{V}) dp,$$

where q , $\mathbf{V}$, p , and p_s are specific humidity, wind vector, pressure, and surface pressure, respectively. Following the Helmholtz theorem, $\mathbf{Q}$ may be divided into the rotational $\mathbf{Q}_R$ and divergent $\mathbf{Q}_D$ components, that is,

$$\mathbf{Q} = \mathbf{Q}_R + \mathbf{Q}_D = \hat{k} \times \nabla \psi_{\mathbf{Q}} + \nabla \chi_{\mathbf{Q}}, \quad (2)$$

where $\psi_{\mathbf{Q}}$ and $\chi_{\mathbf{Q}}$ are streamfunction and potential function of water vapor flux, respectively. Thus, Eq. (1) may be rewritten as

$$P \simeq -\nabla \cdot \mathbf{Q} = -\nabla^2 \chi_{\mathbf{Q}}. \quad (3)$$

The major water vapor flux transported by the monsoon circulation and weather systems is primarily depicted by $(\psi_{\mathbf{Q}}, P)$. Precipitation maintenance is illustrated by the horizontal distribution of $(\chi_{\mathbf{Q}}, \mathbf{Q}_D, P)$ (Chen 1985).

3. Formation mechanism of the South China Sea summer rainfall center

a. Climatology and hydrological condition of the Southeast Asian summer monsoon circulation

The low-level Asian summer monsoon circulation in the subtropics and midlatitudes is characterized by the east–west juxtaposition of the Asian continental thermal low and the subtropical North Pacific anticyclone (Fig. 1b). Around this continental thermal low over Southeast Asia are embedded the northeast–southwest-oriented Bangladesh monsoon trough and the northwest–southeast-oriented SCS trough (red dashed line). Additionally, the southwesterly flow of the cross-equator anticyclonic shear flow and the southeasterly flow of the subtropical North Pacific anticyclone form the western tropical Pacific trough (red dashed line). Ahead of the SCS trough, the strong summer monsoon southwesterlies rotate cyclonically around the SCS and across the Philippines. The SCS summer monsoon rainfall center establishes west of the Philippines. Overlaying the Asian continental thermal low is the Tibetan anticyclone (Fig. 1a). This upper-level anticyclone extends eastward aloft over the western part of the subtropical North Pacific anticyclone.

The hydrological environment of the Asian monsoon circulation is depicted by the distribution of $(\chi_{\mathbf{Q}}, \mathbf{Q}_D, P)_T^2$

(Fig. 1d) alternating between the rainfall center and dry area. A short-wave train $(\chi_{\mathbf{Q}}, \mathbf{Q}_D, P)_T$ extends eastward from the Bay of Bengal to the Philippines Sea. One of the most significant east–west juxtapositions between the dry area and rainfall center is the east–west dipole of the dry area west and the SCS rainfall center east of the SCS trough line. As expected, the dry area is a divergent center of water vapor flux, while the SCS rainfall center is a convergent center of water vapor flux. Of interest is this short-wave train $(\chi_{\mathbf{Q}}, \mathbf{Q}_D, P)_T$ stretched from South Asia to the Philippines Sea, which resides along the tropical convergence zone of the water vapor flux between the Northern Hemisphere midlatitudes and the Southern Hemisphere tropics. In response to this low-level tropical convergence zone from tropical South Asia to the western tropical Pacific and the latent heat released by the rain produced by deep convection, the western tropical Pacific divergent center emerges in the upper troposphere (Fig. 1c). Note that this upper-tropical divergent center is spatially in quadrature with the Tibetan anticyclone. The Asian summer monsoon circulation is maintained by this quadratic relationship between this western Pacific tropical divergent center and the Tibetan anticyclone (Chen 2003).

b. Propagation of the rain-production synoptic systems across the SCS summer rainfall center

Shown in Fig. 1d, the dipole structure $(\chi_{\mathbf{Q}}, \mathbf{Q}_D, P)_T$ is spatially in quadrature with the SCS monsoon trough. What caused the creation of this east–west differential hydrological condition across the SCS? The x – t diagram of rainfall across the SCS summer rainfall center may provide some insight into the rain-producing synoptic systems. The histogram for daily rainfall accumulation over the SCS rainfall center is also used to identify rain-producing synoptic systems. The monsoon life cycle over India during the Monsoon Experiment (MONEX) in summer 1979 was well depicted by the intraseasonal mode (Krishnamurti and Subrahmanyam 1982). The SCS monsoon life cycle was also well portrayed by the intraseasonal mode during the MONEX summer (Chen and Chen 1995). Therefore, this summer is adopted as a sample season to illustrate how rain is produced by different rain-producing synoptic systems.

The x – t diagram for rainfall averaged over the latitudinal zone of 12°–18°N for summer 1979 is shown in Fig. 2a. Nakazawa (1988) observed the *multiple-scale system* of tropical convection and determined that the westward-propagating rainfall is produced by easterly disturbances and 10–24-day modes modulated by the eastward migrating intraseasonal mode. This observation is further confirmed by the histograms for daily rainfall accumulation over the SCS summer rainfall center over the area with a rainfall threshold value $\geq 10 \text{ mm day}^{-1}$, shown in Fig. 2b. These daily rainfall histograms include contributions from four

² The notation $(\)_T$ = total field variable $(\)$.

FIG. 2. (a) The x - t diagram for precipitation averaged over a latitudinal zone of 12° – 18° N during 1 May–31 Aug 1979, and (b) histogram of area-averaged daily precipitation over the SCS summer rainfall center encircled by the threshold rainfall $\geq 10 \text{ mm day}^{-1}$. Colored strips added to the rainfall histogram in (b) cover the rain periods caused by different synoptic systems indicated at the bottom of this histogram: TR, TS/TY, EI, and EW.

types of synoptic disturbances: TR, TS/TY, EI, and EW. The histograms for rainfall produced by these disturbances are marked by strips of different colors shown in the bottom of Fig. 2b. As noted previously in section 3a, the activity of the SCS trough is primarily confined within the SCS by the island chain around the SCS. A low-level convergent center is located ahead of the SCS trough and west of the Philippines as revealed from the contrast between the summer-mean SCS monsoon trough (Fig. 1b) and the $(\chi_Q, Q_D, P)_T$ distribution (Fig. 1d). This convergent center will facilitate and/or intensify the convection/rainfall produced by the four synoptic systems. The SCS trough moves along the island chain, but the other three types of synoptic distribution (TS/TY, EI, and EW) are westward-propagating synoptic systems.

c. Rain-producing synoptic systems

The four typical rain-producing synoptic systems depicted by the streamline chart superimposed with

rainfall are shown in Fig. 3 and verified against the NCEP SRRS chart (NOAA 2016):

- 1) SCS trough: The open trough appears very often over the SCS. Sometimes the deepening of the SCS trough may form a cutoff low within this trough. Because the SCS trough moves around the island chain, rain usually is produced over the trough west of the Philippines (Fig. 3a). As indicated by Fig. 2b, the rainfall produced by the SCS trough covers more summer days than the other synoptic systems.
- 2) TS/TY: The TS/TY season of the SCS covers primarily the period for July–September. Although TS/TY genesis may occur in May when the monsoon westerlies are unusually strong (Chen et al. 2017a), a majority of TS/TYs may have their geneses occur in the western tropical Pacific and move across the Philippines to the SCS (Fig. 3b). Thus, these TSs/TYs contribute some rain over the SCS rainfall center.
- 3) Vortex formed by the interaction of easterly waves with the SCS trough: The vortex generated by this type of interaction still produces rain west of the Philippines, as in the case shown in Fig. 3c.
- 4) Easterly wave: Easterly waves may propagate across the Philippines into the SCS by the tropical trade easterlies around the western Pacific subtropical anticyclone (Fig. 3d). Sometimes, the easterly wave may form a tropical depression around the Philippines, but rainfall is produced ahead of the SCS trough west of the Philippines.

Note that, except for TS/TY, the other two (EI and EW) synoptic systems are advected by the southwesterlies of the SCS trough within its movable domain around the SCS, but usually are confined by the island chain.

d. Contributions of the SCS summer rainfall center by different synoptic systems

The distribution of the SCS rainfall center is shown in Fig. 4a. Contributions from four groups of rain-producing synoptic systems (TR, TS/TY, EI, and EW) measured by the procedure described in section 2b over the 38 summers (1979–2016) are shown in Figs. 4c–f, respectively. The averaged rainy days, occurrence frequency, and duration for four groups of rain-producing synoptic systems are shown in Tables 2–4.³ The combined contributions of $(P_{TR} + P_{TY} + P_{EI} + P_{EW}) = P_C$ are displayed in Fig. 4b. The area-mean rainfall over the area covered over the rainfall threshold value $\geq 10 \text{ mm day}^{-1}$

³ Interannual variations of rainy days, occurrence frequency, and occurrence duration for each group of rain-producing synoptic systems are shown in the online supplemental material (supplement 3).

FIG. 3. Sample synoptic systems depicted by the 850-hPa streamline charts superimposed with precipitation for (a) the SCS trough at 0600 UTC 20 Jun 2014, (b) TY Molave at 0000 UTC 18 Jul 2009, (c) the interaction of easterly waves with the SCS trough at 0000 UTC 1 Jul 2005, and (d) an easterly wave at 1200 UTC 12 Aug 2010. The color scale for rainfall is shown at the top right of every panel.

(red contour) for the four synoptic systems over the 38 summers is shown with histograms in Fig. 4g. The ratios between P_C , P_{TR} , P_{TY} , P_{EI} , P_{EW} and P_T are 97%, 56%, 17%, 12%, and 12%, respectively. Except for P_{TR} produced ahead or over the SCS trough, P_{TY} , P_{EI} , and P_{EW} are primarily produced by the westward-propagating disturbances from the western tropical Pacific. Clearly indicated by the ratio $(P_{TY} + P_{EI} + P_{EW})/P_{TR} \sim 73\%$, the SCS summer rainfall center is contributed to by synoptic systems propagating from both the east and west sides of this rainfall center. A similar conclusion can be drawn from the ratios of rainy days, occurrence frequency, and occurrence duration caused by (TY + EI + EW) against TR (Tables 2–4).

The maintenance of the SCS summer rainfall by the divergent circulation through the water vapor budget was illustrated by $(\chi_Q, Q_D, P)_T$ in Fig. 1d. Our concern here is how this rainfall center is maintained by the water vapor budget of the four rain-producing synoptic systems. To answer this concern, the composite (χ_Q, Q_D, P) distributions for these synoptic systems are prepared by the same approach used to illustrate their rainfall contributions (Figs. 4c–f) to the SCS rainfall center. These composite (χ_Q, Q_D, P) distribution charts for different synoptic systems (Fig. 5) are characterized by the following salient features:

- 1) $(\chi_Q, Q_D, P)_{TR}$ (Fig. 5a): A convergence of water vapor flux exists ahead of the summer monsoon trough. Because 56% of the SCS rainfall center is contributed by the SCS trough type, the composite $(\chi_Q, Q_D, P)_{TR}$ resembles the total $(\chi_Q, Q_D, P)_T$ (Fig. 1d), with an east–west dipole of divergent and convergent centers of water vapor flux, and the east–west juxtaposition of dry and rainy areas across the SCS. This special structure for $(\chi_Q, Q_D, P)_{TR}$ is spatially in quadrature with composite $(\psi_Q, P)_{TR}$ (Fig. 6a).
- 2) $(\chi_Q, Q_D, P)_{TY}$ (Fig. 5b): A center for $(\chi_Q, Q_D, P)_{TY}$ appears west of the Philippines, meridionally in quadrature with $(\psi_Q, P)_{TY}$ (Fig. 6b), located north of the $(\chi_Q, Q_D, P)_{TY}$ center. The location contrast between centers for $(\psi_Q)_{TY}$ and $(\chi_Q)_{TY}$ indicates that the TS/TY vortex facilitates the convergence of water vapor toward the SCS rainfall center and maintains 17% rainfall for the SCS rainfall center.
- 3) $(\chi_Q, Q_D, P)_{EI}$ (Fig. 5c): When the ridge line of the North Pacific subtropical anticyclone migrates northward to the coast of southern China, the SCS trough become east–west oriented toward the central Philippines, indicated by $(\psi_Q, P)_{EI}$ (Fig. 6c). The westward-propagating easterly wave across the Philippines may

FIG. 4. Distribution and contribution of the SCS summer monsoon rainfall center west of the Philippines from four rain-producing synoptic systems: (a) P_T , (b) P_C , (c) P_{TR} , (d) P_{TY} , (e) P_{EI} , and (f) P_{EW} . Note that $P_C = P_{TR} + P_{TY} + P_{EI} + P_{EW}$ and $P_T \approx P_C$. The averaged rainfall over the area covering the SCS summer monsoon rainfall center within a threshold rainfall value $\geq 10 \text{ mm day}^{-1}$ (red line) by different synoptic systems is shown by histograms in (g). The color scales for P_0 are shown in the upper-left side of (a)–(f). Percentages for P_T/P_C contributed by four rain-producing synoptic systems are indicated on the top of rainfall histograms belonging to these synoptic systems.

interact with the SCS trough and develop a converge center of water vapor flux west of the Philippines to produce and maintain 12% rainfall for the SCS rainfall center.

- 4) $(\chi_Q, Q_D, P)_{EW}$ (Fig. 5d): When the ridge line of the North Pacific subtropical anticyclone intrudes into the

northern SCS [implicated by $(\psi_Q, P)_{EW}$; Fig. 6d], the easterly wave can propagate westward across the Philippines. The convergent center can be developed ahead of the easterly wave trough west of the southern Luzon Island. Apparently, the short-wave train for $(\chi_Q, Q_D, P)_T$ between the Philippines and Indochina

TABLE 2. Averaged numbers of rainy days for the SCS summer rainfall center west of the Philippines caused by four different groups of synoptic systems: TR, TY, EI, and EW.

Synoptic system	Monsoon condition		
	Climate (day)	Strong monsoon (day)	Weak monsoon (day)
Total rainy days	89.0	92.1	89.2
TR ($\text{Day}_{TR}/\text{Day}_T$)	49.2 (55%)	58.0 (63%)	39.0 (44%)
TY ($\text{Day}_{TY}/\text{Day}_T$)	8.2 (9%)	8.3 (8%)	7.9 (9%)
EI ($\text{Day}_{EI}/\text{Day}_T$)	10.8 (12%)	10.4 (11%)	10.5 (12%)
EW ($\text{Day}_{EW}/\text{Day}_T$)	20.8 (24%)	15.4 (18%)	31.8 (36%)

TABLE 3. As in Table 2, but for average occurrence frequency (number per summer).

Synoptic system	Monsoon condition		
	Climate	Strong monsoon	Weak monsoon
TR	5.2	5.3	5.0
TY	3.2	3.6	2.9
EI	3.8	3.6	3.8
EW	7.2	5.4	9.1

TABLE 4. As in Table 2, but for average occurrence duration (days).

Synoptic system	Monsoon condition		
	Climate	Strong monsoon	Weak monsoon
TR	9.46	10.94	7.80
TY	2.56	2.31	2.72
EI	2.84	2.89	2.76
EW	2.89	2.85	3.49

draws certain contributions from $(\chi_Q, Q_D, P)_{EW}$. The convergent center can maintain 12% rainfall for the SCS rainfall center.

Two crucial features revealed from the four synoptic systems in producing and maintaining the SCS summer rainfall center west of the Philippines are reconfirmed by the water vapor budget analysis for these synoptic systems. They are the following:

- 1) Only the SCS trough migrating around the SCS is confined by the island chain west of the Philippines Sea. The rainfall produced and/or maintained ahead of the SCS trough coincides with the convergent center ahead of the SCS monsoon trough. This synoptic system contributed 56% rainfall for the SCS rainfall center.
- 2) The northwestward-moving TS/TY into the SCS, the interaction of the westward propagating easterly wave with the east–west oriented SCS trough, and the easterly wave moving across the Philippines develop the convergent center of water vapor flux west of the Philippines to produce and/or maintain 41% of the SCS summer rainfall center west of the Philippines.

The (χ_Q, Q_D, P) analysis for the four rain-producing synoptic systems provides a clear perspective that the SCS summer rainfall center is not directly developed from the interaction of the monsoon westerlies and orography over the upwind side of this orography perceived by the monsoon climatology. Instead, as shown in Fig. 2b, the geographic structure of the SCS facilitates the development of the multiple-scale convective system to produce and maintain rainfall for the SCS summer rainfall center west of the Philippines.

4. Interannual variation of the SCS summer rainfall center

a. Interannual variations of monsoon circulation and the SCS rainfall center

Histograms for rainfall averaged over the area covering the climatological summer rainfall by a threshold value $\geq 10 \text{ mm day}^{-1}$ are shown in Fig. 7c. A rainfall histogram $\geq 16.2 \text{ mm day}^{-1}$ (mean value) $+ 0.8\sigma_R$ ($=10 \text{ mm day}^{-1}$) is defined as a *wet* summer over the SCS summer rainfall center, while a rainfall histogram $< 16.2 \text{ mm day}^{-1} - 0.8\sigma_R$ is defined as a *dry* summer. The former (latter) rainfall histogram is colored blue (red), and the wet and dry summers over the SCS summer rainfall center are distinctively separated.

It was shown by our previous study (Chen et al. 2011) that the late spring–early summer rainfall produced by rainstorms along southern China, Taiwan, and southern Japan increases (decreases) when the monsoon southwesterlies intensify (weaken). Can the interannual variation for the SCS summer rainfall shown in Fig. 7c be affected by the monsoon westerlies?

Using the latitude of maximum monsoon westerlies $u_{\max}(850 \text{ hPa})$ at every longitude from 95° to 150°E , the x – t diagram of $u_{\max}(850 \text{ hPa})$ averaged over a latitudinal zone centered at the $u_{\max}(850 \text{ hPa})$ latitude is shown in Fig. 7a. The contrast among the x – t diagrams of $u_{\max}(850 \text{ hPa})$ and histogram for P_T clearly indicates that the strong (weak) Southeast Asian summer monsoon produced more (less) rainfall over the SCS summer monsoon rainfall center.

However, to measure the monsoon's intensity in a quantitative way, a monsoon index (MI) is defined in terms of the following three variables:

$u(850 \text{ hPa})$ = individual summer-mean 850-hPa zonal velocity,

$\bar{u}(850 \text{ hPa})$ = mean 850-hPa zonal velocity averaged over 38 summers(1979–2016), and

$\Delta u(850 \text{ hPa}) = u(850 \text{ hPa}) - \bar{u}(850 \text{ hPa})$.

The first two variables are used to compute the third variable and its variance.

The variance for $\Delta u(850 \text{ hPa})$ superimposed on the streamline chart prepared with $(\bar{u}, \bar{v})(850 \text{ hPa})$ is shown

FIG. 5. Distributions for composite $(\chi_Q, Q_D, P)_0$ chart for the four synoptic systems: (a) TR, (b) TY, (c) EI, and (d) EW. The vector scale for $(Q_D)_0$, the contour interval for $(\chi_Q)_0$, and the color scales for P_{TR} , P_{TY} , P_{EI} , and P_{EW} are shown on the top right of the four panels.

in Fig. 3d. The $u(850 \text{ hPa})$ field averaged over the maximum variance $[\Delta u(850 \text{ hPa})]$ area boxed by $8^\circ\text{--}11^\circ\text{N}$, $113^\circ\text{--}116^\circ\text{E}$ is located at the upstream side of the SCS summer rainfall center. This area-averaged $u(850 \text{ hPa})$ is defined as the SCS MI. As shown in Fig. 3b, the strong and weak monsoons can be defined by $\text{MI} \geq (5.9 + 0.8\sigma) \text{ m s}^{-1}$ and $\text{MI} \leq (5.9 - 0.8\sigma) \text{ m s}^{-1}$, respectively. Note that $\sigma =$ one standard deviation of the MI time series over 38 summers. It is clearly reflected by the contrast between histograms for MI (Fig. 3b) and P_T (Fig. 3c) that the wet (dry) SCS summer rainfall center is coincident with the strong (weak) Southeast Asian summer monsoon indicated by the MI index.

To further clarify how the wet (dry) SCS rainfall center is affected by the strong (weak) monsoon, the composite 850-hPa summer monsoon flow is depicted by the streamline charts superimposed with westerlies (red) and easterlies (blue) for the wet and dry SCS summer rainfall centers, respectively, in Fig. 8. For the wet summer (Fig. 8a), the ridge line of the western North Pacific subtropical migrates northward to northeastern China. The SCS trough line rotates northward, and the western tropical Pacific monsoon trough line and the equatorial anticyclonic shear line extend eastward. The eastward extension of the latter two tropical monsoon circulation elements is attributed to the eastward

intrusion of the monsoon westerlies to reach beyond 140°E . In contrast, the monsoon circulation during the dry summer (Fig. 8b) exhibits a pronounced change against the wet summer. The subtropical ridge line shifts southward to south of Taiwan, the SCS monsoon trough rotates westward, and the equatorial anticyclonic shear line also retreats westward.

The composite departure of the wet summer monsoon circulation from its climatological-mean summer monsoon circulation (Fig. 1b) is shown in Fig. 8c, while the composite departure for the dry summer monsoon circulation is shown in Fig. 8d. The salient features for the monsoon circulation contrast between the wet (strong) and dry (weak) condition are highlighted below:

- 1) Strong monsoon: The significant 850-hPa monsoon westerlies exhibit a clear eastward extension to reach beyond 140°E . On the other hand, the northwest migration of the ridge line of the North Pacific subtropical anticyclone results in easterly anomalies between Taiwan and Japan toward the Yangtze River. This change in the summer monsoon circulation is characterized by a short-wave train along the North Pacific rim. In fact, centers for the cyclonic (L) and anticyclonic (H) anomalous circulation cells emanating from the Philippines are coincident with

FIG. 6. As in Fig. 5, but for (a) $(\psi_Q, P)_{TR}$, (b) $(\psi_Q, P)_{TY}$, (c) $(\psi_Q, P)_{EI}$, and (d) $(\psi_Q, P)_{EW}$. Contour intervals and $(\psi_Q)_0$ color scales for P_0 are shown at the top of Fig. 6.

the convective cloud anomaly zones forming the P-J oscillation described by Nitta (1987).

- 2) Weak monsoon: Contrasting the change in the weak summer monsoon to the strong one, the southward shift and eastward retreat of the ridge line of the North Pacific anticyclone and the westward retreat of the sequential anticyclonic shear line result in tropical trade easterly anomalies across the Philippine archipelago and the westerly anomalies north of Taiwan. These two opposite zonal flow anomalies form an anomalous anticyclonic circulation cell in the East Asian subtropics connected to an anomalous cyclonic circulation cell to its northeast and other disturbances along the North Pacific to form a short-wave train, like the strong monsoon, but with an opposite phase.

The composite anomalous 850-hPa monsoon circulations for strong and weak monsoons shown in Figs. 8c and 8d, respectively, indicate that the mechanism causing this interannual variation of the Southeast and East Asian monsoon is the P-J oscillation (Nitta 1987).

b. Interannual variation of hydrological conditions

Because atmospheric water vapor primarily exists in the lower troposphere, the horizontal distribution of the water vapor flux should resemble the lower-tropospheric circulation. As shown in section 2d, the major water vapor flux

is contributed by the rotational component depicted by the streamfunction of water vapor flux ψ_Q and precipitation is maintained by the divergent component portrayed by the potential function of water vapor flux (χ_Q, Q_D, P). For strong and weak monsoons, the composite $(\psi_Q, P)_T$ distributions over the Asian monsoon region are shown in Figs. 9a and 9b, respectively. Contrast of the $(\psi_Q)_T$ structure between two extreme monsoon conditions resembles that for the composite 850-hPa monsoon circulation portrayed by the streamline charts (Fig. 8). In addition to $\Delta(\psi_Q)_T$, departure from the composite $(\psi_Q)_T$ in extreme monsoon conditions changes the climate average. Superimposed with the corresponding composite ΔP_T departure of composite rainfall, P_T also exhibits a systematic difference in their spatial patterns (Fig. 9). The rainfall center west of the Philippines and the rainbelt along the Philippines Sea monsoon trough are larger during the strong monsoon summer. In contrast, the rainfall distributions from southern China across Taiwan to southern Japan, and from the Malay Peninsula across the tropical SCS and Borneo to the Solomon Islands, are larger during the weak monsoon summer. The north-south juxtaposition of composite $\Delta(\psi_Q)_T$ and ΔP_T anomalies exhibits two interesting features:

- 1) The contrast between the north-south juxtapositions of $\Delta(\psi_Q)_T$ and ΔP_T is spatially in quadrature for both strong and weak monsoon conditions.

FIG. 7. (a) The $x-t$ diagram of the SCS summer monsoon westerlies averaged over a 3° latitudinal zone with $u_{\max}(850\text{ hPa})$ at its center, (b) MI generated by the summer-mean 850-hPa zonal wind $u(850\text{ hPa})$ averaged over the red box ($8^\circ\text{--}11^\circ\text{N}$, $113^\circ\text{--}116^\circ\text{E}$) shown in (d), (c) the histogram of P_T averaged over the domain of the SCS summer monsoon rainfall center (encircled by a golden line in Fig. 4a), and (d) the streamline chart {prepared with $(\bar{u}, \bar{v})$ (850 hPa) superimposed with the variance of $\Delta u(850\text{ hPa})$ [$=u(850\text{ hPa}) - \bar{u}(850\text{ hPa})$]. Note that $(\bar{u}, \bar{v})(850\text{ hPa}) = 38$ summer (1979–2016) – mean(u, v)(850 hPa) are summer-mean 850-hPa (zonal, meridional) winds. The maximum variance of $\Delta u(850\text{ hPa})$ in the South China Sea is encircled by the red box ($8^\circ\text{--}11^\circ\text{N}$, $113^\circ\text{--}116^\circ\text{E}$).

- 2) For strong and weak monsoons, the meridional juxtaposition of composite $\Delta(\psi_Q)_T$ and ΔP_T anomalies are reversed, as described by the P-J oscillation depicted by Nitta (1987) with the correlation of convective clouds.

Figures 9c and 9d lead us to question *how* the ΔP_T anomalies for this P-J-like oscillation are maintained. Although the magnitude for Q_D is an order smaller than the magnitude for Q_R , precipitation is maintained by the

FIG. 8. (a) Composite 850-hPa streamline charts superimposed with zonal winds $u(850 \text{ hPa})$ for strong monsoon years and (b) weak monsoon years. (c),(d) Departures of (a) and (b), respectively, from the climatological 850-hPa summer monsoon circulation shown in Fig. 1b.

convergence of water vapor flux through the approximated water vapor budget, $P \simeq \nabla \cdot \mathbf{Q}_D$. The distributions for $(\chi_Q, \mathbf{Q}_D, P)_T$ and its departure $\Delta(\chi_Q, \mathbf{Q}_D, P)_T$ from the climatological mean will be used to illustrate the maintenance of the monsoon rainfall, and the ΔP_T anomalies for both the strong and weak monsoon conditions.

Revealed from the $(\chi_Q, \mathbf{Q}_D, P)_T$ distributions for the strong and weak monsoons shown in Figs. 10a and 10b, respectively, the $(\chi_Q)_T$ field exhibits a $(\mathbf{Q}_D)_T$ convergent center of water vapor flux over this Asian monsoon region between the Northern and Southern Hemispheres, and the Eastern and Western Hemispheres outside this monsoon for both the strong and weak monsoons. Note that the $(\chi_Q)_T$ field encircling this monsoon (Figs. 10a,b) is spatially in quadrature with the $(\psi_Q)_T$ fields shown in Figs. 9a and 9b. The $\Delta(\chi_Q, \mathbf{Q}_D, P)_T$ distribution departures from the climatological-mean $(\chi_Q, \mathbf{Q}_D, P)_T$ (Fig. 1d) for both the strong and weak monsoons shown in Figs. 9c and 9d, respectively, exhibit the following salient features between $\Delta(\psi_Q, P)_T$ and $\Delta(\chi_Q, \mathbf{Q}_D, P)_T$:

- 1) For both the *strong* and *weak* monsoons, the $\Delta(\chi_Q)_T$ field is spatially in quadrature with the corresponding $\Delta(\psi_Q)_T$ field.
- 2) The positive (negative) ΔP_T anomalies during both extreme monsoon seasons are maintained

(suppressed) by convergent (divergent) water vapor flux $\Delta(\mathbf{Q}_D)_T$.

During the strong monsoon season, the *cyclonic* tropical–subtropical cell of the P-J oscillation (Fig. 9c) enhances the convergence of water vapor flux from the rainbelts north and south of this anomalous circulation cell to maintain the rainfall center located in the southern half of this P-J circulation cell (Fig. 10c). On the contrary, the *anticyclonic* tropical–subtropical circulation cell (Fig. 9d) for the P-J oscillation strengthens the divergence of water vapor flux to send the water vapor flux from the SCS rainfall center west of the Philippines along the Philippines Sea monsoon trough to enhance the rainfall along the northern and southern peripheries for this P-J circulation cell.

c. Contributions from four rain-producing synoptic systems to interannual variation of the SCS summer rainfall center

The rainfall contributions from four different rain-producing synoptic systems were presented in Fig. 4. Can these rainfall contributions undergo interannual variations by the modulation of the Asian summer monsoon through the P-J oscillation? The composite rainfall distributions of $P_T, P_C, P_{TR}, P_{TY}, P_{EI}$, and P_{EW}

FIG. 9. Composite $(\psi_Q, P)_T$ charts for (a) strong and (b) weak monsoon years. (c), (d) Departures of (a) and (b), respectively, from the summer (mid-May–August)–mean $(\psi_Q, P)_T$ chart [shown in the online supplemental material (supplement 4)]. Contour intervals for $(\psi_Q)_T$ and $(\Delta\psi_Q)_T$ and color scales for P_T and ΔP_T are shown at the top of (a) and (c).

for both the strong- and weak-monsoon summers are shown in Fig. 11. As revealed from the contrast between distributions of P_T , P_C , P_{TR} , P_{TY} , and P_{EI} during the strong monsoon (Figs. 11a–f) and the weak monsoon (Figs. 11g–k), the contributions of the former group are larger than those of the latter group. However, the contribution of P_{EW} to the SCS summer rainfall center in the strong summer monsoon season is smaller than in the weak summer monsoon. As observed by our previous study (Chen and Weng 1998), the anomalous cyclonic (anticyclonic) circulation cell of the P–J oscillation in the tropical–subtropical North Pacific region facilitates (hinders) the westward propagation of easterly waves across the Philippines. Estimated by the present study, on average, there are 9.1 easterly waves during the weak monsoon summer, but only 5.4 occur during the strong monsoon summer. As shown in Fig. 8, the anomalous cyclonic (anticyclonic) circulation cell of the P–J oscillation in the tropical–subtropical North Pacific region is associated with the strong (weak) monsoon. This contribution contrast of P_{EW} between these two extreme monsoon conditions is likely attributable to the population of easterly waves moving across the Philippines.

To make a more quantitative comparison from rainfall contributions for the four rain-producing synoptic systems during the strong and weak summer monsoons, histograms of P_T , P_C , P_{TR} , P_{TY} , P_{EI} , and P_{EW} averaged over the area of the SCS summer rainfall (P_T) center defined by a threshold value $\geq 10 \text{ mm day}^{-1}$ are shown in Figs. 11m and 11n for these two extreme monsoon conditions, respectively. This comparison is further illustrated with the averaged values of all concerned variables in these two panels. Interesting features emerging from them are highlighted below:

- 1) For any monsoon climate conditions, P_C is about 2%–4% smaller than P_T . This discrepancy may be attributed to the computation bias or some less crucial hydrological processes.
- 2) Revealed from the contrast between Figs. 11f and 11i, P_{EW} during the strong monsoon condition is smaller than P_{EW} during the weak monsoon condition. This contrast was caused by a population difference of easterly waves across the Philippines during the strong and weak monsoons.
- 3) Rainfall contributions to the SCS summer rainfall center by the SCS monsoon trough P_{TR} west of the

FIG. 10. As in Fig. 9, but for $(\chi_Q, Q_D, P)_T$. Contour intervals for $(\chi_Q)_T$ and $\Delta(\chi_Q)_T$, vector scales for $(Q_D)_T$ and $\Delta(Q_D)_T$, and color scales for P and ΔP are shown at the tops of (a) and (c).

Philippines, and by the synoptic systems from the Philippine Sea (P_{TY}, P_{EI}, P_{EW}) exhibit an interesting contrast:

- $(P_{TY} + P_{EI} + P_{EW})$ (strong monsoon): P_{TR} (strong monsoon) = 38%:60%, but $(P_{TY} + P_{EI} + P_{EW})$ (weak monsoon): P_{TR} (weak monsoon) = 49%:47%.

The ratios for rainy days caused by (TY + EI + EW) against TR during the strong and weak monsoons (Tables 2–4) exhibit very similar contrasts as the ratios of rainfall amounts, respectively. Additionally, during the weak monsoon, the synoptic systems from the Philippine Sea are more effective to produce rainfall over the SCS summer rainfall center west of the Philippines.

- P_{TR} (strong monsoon): P_T (strong monsoon) = 60%:100%, but P_{TR} (weak monsoon): P_T (weak monsoon) = 47%:100%.

Clearly, P_{TR} is more effective in producing rainfall over the SCS summer rainfall center west of the Philippines during the strong monsoon season than during the weak monsoon season.

Finally, the rainfall contributions to form the SCS summer rainfall center west of the Philippines by four different rain-producing synoptic systems provide a new insight into the formation mechanism of this

rainfall center. As shown by the x - t diagram of P (12°–15°N) in Fig. 2, this rainfall center is formed by the multiple-scale process (Nakazawa 1988) anchored west of the Philippines rather than simply being formed by the interaction of the monsoon southwesterlies with the mountains along the western part of the northern Luzon Islands.

d. Maintenance for interannual variation of the SCS summer rainfall center by four synoptic systems

Shown in Fig. 11, $P_T, P_C, P_{TR}, P_{TY}, P_{EI}$, and P_{EW} undergo significant interannual variation between the strong and weak monsoons. We denote departures of these rainfall variables from their corresponding climatological mean values as $\Delta P_T, \Delta P_C, \Delta P_{TR}, \Delta P_{TY}, \Delta P_{EI}$, and ΔP_{EW} , respectively. The interannual variations for the water vapor transport and rainfall maintenance for ΔP_O are given by $\Delta(\psi_Q, P)_O$, and $\Delta(\chi_Q, Q_D, P)_O$, respectively, where the subscripted closed set of parentheses indicate any of the rainfall types. However, a concern is raised here with *how* interannual variations of the rainfall contributions from different synoptic systems to the SCS summer rainfall center are maintained and coupled with the P-J pattern for $(\psi_Q, P)_T$ between the strong and weak summer monsoons.

FIG. 11. As in Fig. 4, except (a)–(f) and (m) are for strong monsoon years, and (g)–(l) and (n) are for weak monsoon years. Percentages of P_O/P_T contributed by four rain-producing synoptic systems in (m) and (n) are located at the top of the rainfall histograms belonging to these synoptic systems.

To save space, distributions for $(\psi_Q, P)_{TR}$, $(\psi_Q, P)_{TY}$, $(\psi_Q, P)_{EI}$, and $(\psi_Q, P)_{EW}$ for the two extreme summer monsoons are presented in the supplemental material (see supplement 4). Nevertheless, departures of $(\psi_Q, P)_0$, $\Delta[(\psi_Q, P)_{TR}, (\psi_Q, P)_{TY}, (\psi_Q, P)_{EI}, (\psi_Q, P)_{EW}]$ for strong and weak monsoons are displayed in Figs. 12a–d and 12e–h, respectively. Indicated by $(\psi_Q)_T$ in Fig. 9, the SCS monsoon trough deepens (fills) during the strong (weak) monsoon. In fact, this circulation change between the two extreme monsoons is also reflected by all four synoptic systems: As shown in Figs. 12a–d (Figs. 12e–h), all $\Delta\psi_0$ are negative (positive) over the SCS, when the monsoon is strong (weak). Thus, during the former (latter) monsoon condition, $\Delta\psi_0$ is positive (negative) over the SCS summer monsoon rainfall center, except ΔP_{EW} . This exception is caused by the weakening (strengthening) of tropical trade easterlies across the Philippines. The combinations $\Delta[(\psi_Q, P)_{TR} + (\psi_Q, P)_{TY} + (\psi_Q, P)_{EI} + (\psi_Q, P)_{EW}]$ for the strong and weak monsoons are presented in Figs. 13a and 13b, respectively. Interestingly, an oscillation emerges from the well-depicted P-J pattern of ΔP_C meridionally in quadrature with the $\Delta(\psi_Q)_C$ pattern, resembling those shown in Figs. 9c and 9d, respectively. Implicated by the contrast between Figs. 12 and 13, the P-J pattern/oscillation is a reflection of the monsoon climate change organized by the four rain-producing synoptic systems between the two extreme monsoon conditions. On the other hand, it may be argued the changes of these four synoptic systems are modulated by the P-J oscillation.

During the strong monsoon, the rainbelts, including the SCS summer rainfall center west of the Philippines and the rainfall center along the Philippine Sea summer monsoon trough, are enhanced/intensified and juxtaposed with the weakened rainbelts north and south of the former ones (Fig. 10a). This reversed meridional juxtaposition of anomalous rainbelts during the weak monsoon is shown in Fig. 10b. Note that the $\Delta P_T \sim \Delta P_C [= \Delta(P_{TR} + P_{TY} + P_{EI} + P_{EW})]$ anomaly distribution, the convective cloud anomaly pattern shown by Nitta (1987), is in quadrature with the $\Delta(\psi_Q)_T \simeq \Delta(\psi_Q)_C$ anomaly pattern. However, to understand how the ΔP_C anomaly pattern is maintained, the $\Delta[(\chi_Q, Q_D, P)_{TR}, (\chi_Q, Q_D, P)_{TY}, (\chi_Q, Q_D, P)_{EI}, (\chi_Q, Q_D, P)_{EW}]$ values for both the strong and weak monsoons are shown in Figs. 14a–d and 14e–h, respectively. For comparison, distributions of the four groups for $(\chi_Q, Q_D, P)_0$ for the strong and weak monsoons are presented in the supplemental material (see supplement 2). During the strong monsoon, the rainbelts, including the SCS summer rainfall center west

of the Philippines and the rainfall center along the Philippine Sea summer monsoon trough, are enhanced/intensified and meridionally juxtaposed with two weakened rainbelts north and south of the former ones (Fig. 10a). This reversed meridional juxtaposition of anomalous rainbelts during the weak monsoon is shown in Fig. 10b.

During the strong monsoon, how do $\Delta(P_{TR} + P_{TY} + P_{EI} + P_{EW})$ and $\Delta[(\chi_Q, Q_D)_{TR}, (\chi_Q, Q_D)_{TY}, (\chi_Q, Q_D)_{EI}, (\chi_Q, Q_D)_{EW}]$ anomalies contribute to maintain these $\Delta(\chi_Q, Q_D, P)_C$ anomalies shown in Fig. 10a? The positive enhanced/intensified central rainbelt is primarily contributed by ΔP_{TR} supplemented by positive ΔP_{TY} and ΔP_{EI} , and maintained by anomalous convergent flux $\Delta[(Q_D)_{TR}, (Q_D)_{TY}, (Q_D)_{EI}]$ superimposed on the positive $\Delta[(\chi_Q)_{TR}, (\chi_Q)_{TY}, (\chi_Q)_{EI}]$ center. The rainfall center along the Philippine Sea monsoon trough is largely contributed by $\Delta(P_{TR}, P_{TY})$ and maintained by $\Delta[(\chi_Q, Q_D)_{TR}, (\chi_Q, Q_D)_{TY}, (\chi_Q, Q_D)_{EI}, (\chi_Q, Q_D)_{EW}]$. The negative ΔP_T anomaly belts north and south of the major central monsoon rainfall center are contributed not by westward propagating easterly waves but by the suppressed convective rainfall activity, when the easterly waves are inactive. In contrast, during the weak monsoon, it is shown in Figs. 14e–g that the $\Delta(P_{TR}, P_{TY}, P_{EI})$ anomaly centers become negative, except for ΔP_{EW} anomalies over the SCS summer rainfall center west of the Philippines, and the two rainbelts north and south of the suppressed central rainbelt (Fig. 14h). For the relationship between the P-J pattern depicted by $\Delta(\psi_Q, P)_T(\text{strong})$ and contributions by $\Delta[(\psi_Q, P)_{TR}, (\psi_Q, P)_{TY}, (\psi_Q, P)_{EI}, (\psi_Q, P)_{EW}]$, the P-J pattern formed by $\Delta P_T/\Delta P_C$ is a synthesized pattern formed by $\Delta(P_{TR}, P_{TY}, P_{EI}, P_{EW})$ and maintained by $\Delta[(\chi_Q, Q_D)_{TR}, (\chi_Q, Q_D)_{TY}, (\chi_Q, Q_D)_{EI}, (\chi_Q, Q_D)_{EW}]$.

5. Concluding remarks

The Asian summer rainfall centers usually appear in the upwind side of the coast. From the climatological perspective, these rainfall centers are considered established by the interaction of monsoon westerlies with the orography along the coastal line (e.g., Xie et al. 2006). Because a summer rainfall center exists in the SCS west of the Philippines, an effort is made in this study to explore how this rainfall center is formed. To serve this purpose, the x – t diagram for the 1979 summer rainfall through the SCS summer rainfall center west of the Philippines was shown in Fig. 2. This x – t diagram revealed this rainfall center is formed by the multiple-scale processes (Nakazawa 1988), primarily by the rainfall produced by four groups of rain-producing synoptic systems:

FIG. 12. Composite (a) $\Delta(\psi_0, P)_{TR}$, (b) $\Delta(\psi_0, P)_{TY}$, (c) $\Delta(\psi_0, P)_{EI}$, and (d) $\Delta(\psi_0, P)_{EW}$ for strong monsoon years. (e)–(h) As in (a)–(d), but for weak monsoon years. Contour intervals of $\Delta(\psi_0)_0$ are shown in the upper-left corners of (a) and (b), while the color scales for ΔP_0 are shown in the lower-right corners for (e) and (f).

1) the SCS trough (TR) west of the island chain, 2) TS/TY, 3) interaction of tropical easterly wave/tropical depression with the SCS TR (EI), and 4) easterly waves (EW). With this classification of rain-producing synoptic systems, two major tasks are pursued: determining

1) the formation mechanism of the SCS summer-monsoon rainfall center west of the Philippines and 2) the cause for the interannual variation of this SCS rainfall center. New findings for these two tasks are summarized below.

FIG. 13. (a) Composite $\Delta(\psi_Q, P)_C$ (strong) is the sum of Figs. 12a–d, composite $[\Delta(\psi_Q, P)_{TR} + \Delta(\psi_Q, P)_{TY} + \Delta(\psi_Q, P)_{EI} + \Delta(\psi_Q, P)_{EW}]$ (strong) for strong monsoon years. (b) As in (a), but for composite $\Delta(\psi_Q, P)_C$ (weak), sum of Figs. 12e and 12f, composite $[\Delta(\psi_Q, P)_{TR} + \Delta(\psi_Q, P)_{TY} + \Delta(\psi_Q, P)_{EI} + \Delta(\psi_Q, P)_{EW}]$ (weak) for weak monsoon years. (c) As in (a), but for composite $\Delta(\chi_Q, Q_D, P)_C$ (strong), sum of Figs. 14a–d, composite $[\Delta(\chi_Q, Q_D, P)_{TR} + \Delta(\chi_Q, Q_D, P)_{TY} + \Delta(\chi_Q, Q_D, P)_{EI} + \Delta(\chi_Q, Q_D, P)_{EW}]$ (strong) for strong monsoon years. (d) As in (c), but for $\Delta(\chi_Q, Q_D, P)_C$ (weak), sum of Figs. 14e and 14f, composite $[\Delta(\chi_Q, Q_D, P)_{TR} + \Delta(\chi_Q, Q_D, P)_{TY} + \Delta(\chi_Q, Q_D, P)_{EI} + \Delta(\chi_Q, Q_D, P)_{EW}]$ (weak) for weak monsoon years. Contour intervals of $\Delta(\psi_Q)_C$ and color scale of ΔP_C are shown at the top of (a), while contour intervals for $\Delta(\chi_Q)_C$, vectors scale for $\Delta(Q_D)_C$, and color scales for ΔP_C are shown at the top of (c). The error variances for $(\chi_Q, Q_D, P)_C$ against $(\chi_Q, Q_D, P)_T$ (Figs. 9c,d and 10c,d) are in a range of 8%–9% over 100°–150°E, 10°S–40°N.

a. Formation mechanism of the SCS summer rainfall center

The low-tropospheric monsoon flow over Southeast Asia exhibits a well-developed northwest–southeast-oriented trough between northern Vietnam and northwest of Borneo (Fig. 1b). A convergent center of water vapor flux coincident with the SCS summer monsoon rainfall center is located ahead of the SCS monsoon trough and west of the Philippines (Fig. 1d). This convergent center facilitates the rainfall generation by the four groups of different rain-producing synoptic systems.

Climatologically, the rainy days over the SCS summer rainfall center are 89 days over the mid-May–August period. On average over this summer season, TR, TY, EI, and EW cover 49.2, 8.2, 10.8, and 20.8 days, respectively. Rainfall contributions from these four synoptic systems to the SCS summer rainfall center are 56% (P_{TR}), 17% (P_{TY}), 12% (P_{EI}), and 12% (P_{EW}).

Because the total rainfall P_T over the SCS summer monsoon rainfall center is primarily contributed by the four synoptic systems ($P_T \simeq P_{TR} + P_{TY} + P_{EI} + P_{EW}$), the maintenance of P_T should be responsible by these four rain-producing synoptic systems. This inference is substantiated by

$$(\chi_Q, Q_D, P)_T \simeq (\chi_Q, Q_D, P)_{TR} + (\chi_Q, Q_D, P)_{TY} + (\chi_Q, Q_D, P)_{EI} + (\chi_Q, Q_D, P)_{EW}$$

in Figs. 4 and 5.

b. Interannual variation of the SCS summer rainfall center

The SCS monsoon trough deepens (fills) when the SCS monsoon circulation intensifies (weakens), as indicated by the strength of the SCS monsoon westerlies. These circulation changes lead to an anomalous cyclonic

FIG. 14. As in Fig. 12, but for $\Delta(\chi_Q, Q_D, P)_0$. Contour intervals for $\Delta(\chi_Q)_0$ and vector scales for $\Delta(Q_D)_0$ are shown in the upper right-hand corners for (a) and (b), while color scales for ΔP_0 are shown in the lower left-hand corners for (e) and (f).

circulation cell between the Philippines and southern Japan when the monsoon is strong or an anomalous *anticyclonic* circulation when the monsoon is weak. The rainbelt, consisting of the SCS rainfall center and the rain center along the Philippines Sea trough, is enhanced

during the strong monsoon. In contrast, the northern rainbelt stretching from northern Vietnam across Taiwan to southern Japan and the southern rainbelt extending from central Vietnam across Borneo to the Solomon Islands are suppressed. The opposite changes

for these three rainbelts occur during the weak monsoon. The alternation of anomalous rainbelts is the P-J oscillation, which is meridional in quadrature with the P-J oscillation of an anomalous circulation cell.

The interannual variation of the SCS summer rainfall center west of the Philippines is shown in Fig. 11; $[P_C(\text{strong}) - P_C(\text{weak})]/[P_C(\text{climate})] \simeq 40\%$. Contributions to P_C by rainfall-produced synoptic systems originating west and east of the Philippines show significant contrast. $P_{TR}:(P_{TY} + P_{EI} + P_{EW})$ (strong monsoon) = 60%:38%, but P_{TR} (weak monsoon): $(P_{TY} + P_{EI} + P_{EW})$ (weak monsoon) = 47%:49%. A contrast of rainy days caused by P_{TR} and $(P_{TY} + P_{EI} + P_{EW})$ is very close to the ratio of rainfall amounts during the strong and weak monsoons, respectively. Additionally, the P-J pattern depicted by $\Delta(\psi_Q, \chi_Q)_T$ mainly is formed by $\Delta[(\psi_Q, \chi_Q)_{TR} + (\psi_Q, \chi_Q)_{TY} + (\psi_Q, \chi_Q)_{EI} + (\psi_Q, \chi_Q)_{EW}]$ for the corresponding monsoon condition only with the difference measured by error variance $< 9\%$ over the domain of $100^\circ\text{--}160^\circ\text{E}$, $10^\circ\text{S--}40^\circ\text{N}$. Apparently, the P-J pattern is basically formed by changes of the four groups of synoptic systems (during the strong and weak monsoons).

In summary, the SCS summer monsoon rainfall center is formed by contributions from the synoptic systems for TR, TY, EI, and EW over the low-tropospheric convergent center east of the SCS monsoon trough and west of the Philippines. The interannual variation of this summer monsoon rainfall center is caused by the responses of these four rain-producing synoptic systems to the interannual variation of the Southeast Asian monsoon circulation. A combination of these responses is reflected by the P-J oscillation of $\Delta(\psi_Q, P)_T$ and $\Delta(\chi_Q, Q_D, P)_T$.

With these new findings for the formation mechanism of the SCS summer monsoon rainfall center and its interannual variation, the following studies are suggested for future efforts:

- 1) Numerous studies demonstrated the interannual variation of TS/TY genesis and tracks, rainfall, and monsoon onset, etc., are caused by the ENSO through the P-J oscillation. Nevertheless, the present study observed that the interannual variation of the summer monsoon intensity is clearly indicated by the monsoon westerlies, which may not be always coincident with the Niño-3.4 index. Takahashi et al. (2015) also noted that rainfall variation in Thailand is not attributed to the ENSO cycle. Thus, it is important to understand how the interannual variation for the summer monsoon intensity occurs.
- 2) The $\Delta(\psi_Q, P)_T$ and $\Delta(\chi_Q, Q_D, P)_T$ anomalies of the P-J oscillation/pattern are formed by the $\Delta(\psi_Q, P)_T$ and $\Delta(\chi_Q, Q_D, P)_T$ anomalies of four concerned synoptic systems. This leads to a concern whether the

interannual variation in the activity of these four synoptic systems forms the P-J oscillation pattern or the P-J oscillation/pattern modulates the activity of these four synoptic systems to cause the interannual variation of monsoon climate change. The interactions between the synoptic and climate systems need a serious study.

- 3) Previous studies addressed the perspective of climate simulation that monsoon rains always form a rainfall center over the upwind side of the coast because of the interaction of the monsoon flow with coastal orography. On the contrary, the present study shows that the SCS summer monsoon rainfall center is contributed by synoptic systems from both sides of the Philippines over the climatological convergent center west of the Philippines. This finding requires future study to explore the role played by the weather systems in the development of a special climate system.

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