Genesis and Development of Spring Rainstorms in Northern Southeast Asia: Southwest China–Northern Indochina and the Northern South China Sea

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ABSTRACT

During May and June, the monsoon rainfall in northern Southeast Asia is primarily produced by rainstorms. At the mature stage, these storms, coupled with a midtropospheric subsynoptic-scale trough, produce rainfall $\geq 50 \text{ mm } (6 \text{ h})^{-1}$ and exhibit a cyclonic surface vortex. With a scale $\sim O(10^2) \text{ km}$, rainstorms during the period of 1979–2016 are identified with station and satellite observations, along with assimilation data. Several dynamic processes of rainstorm geneses are disclosed by an extensive analysis. 1) Maximum occurrence of rainstorm geneses is located in the midtroposphere of two regions (northern Vietnam–southwestern China and the northern South China Sea), but eventually penetrates downward to the surface. 2) The environment favorable for rainstorm genesis is a southwest–northeast-oriented narrow trough formed by the confluence of the midtropospheric northeasterly around the eastern Tibetan Plateau and the lower-tropospheric monsoon southwesterlies. Because the criterion for Charney–Stern instability is met by the shear flow of this narrow trough, rainstorms are likely initiated by this instability. 3) The majority of rainstorm geneses occurs during the evening over the land and the morning at sea. This timing preference is caused by the modulation of the clockwise rotation of the East Asia continent circulation in response to the diurnal variation of the land–sea thermal contrast. These new findings from this study offer not only a new perspective for the genesis mechanism of the late spring–early summer rainstorms in northern Southeast Asia, but also a new initiative to develop medium-range forecasts for these rainstorms.

1. Introduction

The warm-season rainfall variation over south China and Taiwan undergoes an active rainfall period (May and June) and a typhoon (TY) active period (mid-July to early October) separated by a midsummer (late June–early July) dry spell (Ramage 1952), as illustrated by Fig. 1a. During the active monsoon phase, the frontal activity diminishes, but the monsoon southwesterly flow intensifies. Some midtropospheric disturbances accompanied by heavy rainfall moving along this monsoon southwesterly flow may produce heavy precipitation

over south China and southwest Taiwan. Caused by the northward migration of the western subtropical anticyclone, rainfall over south China and Taiwan significantly decline over the break monsoon phase. This northward advance of the subtropical anticyclone is followed by the intertropical convergence zone (ITCZ), the increase of TY occurrence frequency over the western tropical Pacific, and rainfall enhancement to reach its second maximum in late July.

Ramage's depiction of the warm-season rainfall variation for south China and Taiwan was adopted by the East Asian monsoon community to portray the East Asian monsoon life cycle; mei-yu–baiyu–break–typhoon season (e.g., Chang 1972; Lau and Li 1984; Ninomiya and Murakami 1987). Using multidata sources, this monsoon life cycle around the East China Sea and the northern part of the South China Sea (SCS), including southern China, Taiwan, Japan, and South Korea, was comprehensively analyzed by Chen et al. (2004). Despite

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FIG. 1. (a) Life cycle of the Southeast/East Asia summer monsoon [active phase (1 May–15 Jun), break phase (15 Jun–15 Jul), and revival phase (16 Jul–15 Sep)], depicted with the climatological 15-day-mean rainfall for Taiwan from April to October (adopted from Chen et al. 2004) and (b) the rainfall in Taiwan produced by different weather systems (including diurnal variations, typhoons, rainstorms, and fronts) for different phases of the summer monsoon life cycle (adopted from Chen et al. 2011).

this well accepted depiction of the East Asian warm-season monsoon life cycle, the IR image from the Geostationary Meteorological Satellite (GMS) for the eastward-propagating rainstorms in the northern SCS (Fig. 2) during the SCS Monsoon Experiment (SCSMEX; Lau 1997; Lau and Yang 1997) leads us to *infer* their crucial role in the rainfall contribution during the active monsoon phase in the southern part of East Asia. This inference was supported by our pilot studies (Fig. 1b; Chen et al. 2004; Wang and Chen 2008; Chen et al. 2011).

To understand the *rain-producing process* during the active monsoon phase over southwest Taiwan, the Southwest Monsoon Experiment/Terrain-Influenced Monsoon Rainfall Experiment (SoWMEX/TiMREX) (Jou et al. 2011) was conducted over May and June for 2008. This active monsoon phase was separated into two periods (Davis and Lee 2012). Major rainfall during the *first* period (15 May–2 June) was primarily produced by afternoon convection over the elevated terrain of southwest Taiwan. The *second* period (3–30 June)

covered three rain spells; each lasted about 5–6 days over either the southwest plains or the elevated region (≥ 200 m). The postexperiment analyses focused on two spells (1–6 June) and (12–17 June). During the former spell, a series of three MCSs occurred south of a front moving across southern Taiwan: two rainfall events (3 and 4 June) followed by a mesoscale convective vortex (MCV; 5 June) on the southwestern coast plain of Taiwan (Lai et al. 2011). The latter spell included two heavy rainfall events. The first event exhibited an eastward propagation of MCS, while the second event showed a westward propagation of MCS (Wang et al. 2014). These efforts shed light on the synoptic conditions and physical processes in developing these heavy rainfall events over southwestern Taiwan during the *active* summer monsoon phase.

Despite efforts made by previous studies (Davis and Lee 2012; Lai et al. 2011; Wang et al. 2014), the genesis/formation mechanism of MCSs or MCV observed by these studies during the active summer monsoon phase is still beyond reach. The rainfall produced by these MCSs often leads to disastrous floods over some monsoon regions. To manage the water supply, reduce traffic hazards, and mitigate disasters caused by these heavy rainfall events, forecast improvement of these events is necessary. A better understanding of the genesis and development mechanisms facilitates our pursuit of this goal. It is inferred from the IR image of MCSs (Fig. 2) that the geneses of these rainfall events occur in the upstream region of monsoon westerlies during the active monsoon phase. This inference is confirmed by tracking rainfall at three locations along the monsoon westerlies [Hanoi, Vietnam (WMO 48820), Hong Kong (WMO 45004), and southwest Taiwan] during 15 May–20 June 2008 [Late Spring–Early Summer Rainstorm Experiment (LSESREX), Taiwan EAMEX Committee (2007)]. As inferred from Fig. 3, geneses of these rainfall events may occur over the southwestern China–northern Indochina region and the northern SCS. There is no formal definition for the heavy rain events during the active monsoon phase of East Asia, but they are commonly named “rainstorm” when the local rainfall rate is ≥ 50 mm $(6\text{ h})^{-1}$ (Tao and Ding 1981; Chen et al. 1998). This term is adopted by this study. An extensive analysis for rainstorm formation during the active summer monsoon phase is reported in this study. Our efforts include a backtracking procedure, genesis mechanism, and diurnal variation in the genesis over land and sea.

To achieve these goals, this manuscript is organized in the following manner. Section 2 describes an identification of downstream development and the backtracking procedure for rainstorms to their genesis

FIG. 2. The IR image for a series of MCSs/rainstorms between the Gulf of Tonkin and Taiwan at 0000 UTC 5 Jun 1998 during the South China Sea Monsoon Experiment (SCSMEX).

locations. Data utilized for this analysis are also presented. [Section 3](#) documents the occurrence statistics and geographic/spatial distribution of rainstorm genesis for the 1979–2016 period. After monsoon onset in the South China Sea during mid-May, the low-level monsoon westerlies intensify across Indochina. Following the development of the Tibetan anticyclone, the anticyclonic flow develops around the Tibetan Plateau. The confluence of these two monsoon flows forms a positive shear flow extending from the Yangtze River valley to southwest China, a favorable environment for rainstorm genesis. A detailed synoptic analysis related to the rainstorm genesis both in the southwestern China–northern Vietnam region and the northern South China Sea is presented in [section 4](#). Because the Tibetan anticyclone exists in the upper troposphere, rainstorm genesis occurs similar to the African easterly waves south of the African easterly jet in the midtroposphere (e.g., [Burpee 1972](#)). For this reason, the Charney–Stern instability criterion ([Charney and Stern 1962](#)) is applied to explore the possible genesis mechanism of the rainstorm. In response to the land–sea thermal contrast, the diurnal variation of the atmospheric circulation undergoes a clockwise rotation over the East Asian continent and adjacent oceans ([Chen 2006](#)). Because the rainstorm genesis exhibits a land–sea diurnal variation, the possible impacts on the diurnal variation of rainstorm genesis by diabatic heating and the regional

circulation are illustrated in [section 5](#). Finally, [section 6](#) presents a summary of new findings from this study. Future studies about rainstorm impacts in monsoon development, an intraseasonal mode on the rainstorm activity, and possible development of East Asian rainstorms from the eastward-propagating midtropospheric cyclones in South Asia are suggested.

2. Identification of rainstorm and analysis data

It has been perceived by the East Asian meteorological community that rainstorms during the active monsoon phase develop from MCSs associated with the mei-yu–baiu front (e.g., [Lau and Li 1984](#); [Ninomiya and Murakami 1987](#)). Research has focused on the structure, development, and weather/rainfall impacts from these rainstorms, but rarely on the genesis mechanism and large-scale environment favorable for genesis. To explore the latter two issues, rainstorms should be correctly identified and backtracked to their genesis locations. Because of a lack of understanding in its genesis mechanism, rainstorms are easier to identify during their mature phase than their genesis phase. To avoid any possible error in search of genesis locations, rainstorms should be properly identified so a backtracking procedure can be performed. Some insight on the genesis mechanism can then be revealed from locations and the associated synoptic conditions.

FIG. 3. Rainfall [$\text{mm} (3 \text{ h})^{-1}$] histograms for 15 May–20 Jun 2008: (a) Hanoi, (b) Hong Kong, and (c) southwest Taiwan (22° – 24°N , 120° – 120.6°E). Dates of five identified rainstorms at these three locations are marked with red arrows along the time axis and by RN ($N = 1, \dots, 5$) on the rainfall histograms. These three locations are marked with red dots on the top geographical panel including genesis locations of five rainstorms (marked by blue crosses).

a. Identification criteria for a rainstorm

A typical rainstorm shown in Fig. 4 is utilized to illustrate the criteria used herein:

- 1) A rainstorm reaches a rainfall amount $\geq 50 \text{ mm} (6 \text{ h})^{-1}$ at the mature stage of its life cycle (Chen et al. 1998).
- 2) Relative vorticity of a rainstorm at its genesis location in the midtroposphere (700–500-hPa layer) should reach a threshold value of $\geq 2 \times 10^{-5} \text{ s}^{-1}$. This vorticity value becomes larger and penetrates downward, when a surface vortex appears at some later stage.
- 3) A rainstorm (indicated by a red oval in Fig. 4c) is coupled with a midtropospheric subsynoptic-scale trough (a thick short red line) (Fig. 4b). A cold, surgelike flow (blue arrow) and a narrow trough (red-dashed line) are located behind an upper-tropospheric short-wave trough (Fig. 4c).
- 4) Underneath the midtropospheric subsynoptic-scale trough, the rainstorm, a cyclonic vortex in the lower troposphere, is identified either by a surface

FIG. 4. A typical rainstorm case identified at 0600 UTC 2 Jun 2008: (a) 200-hPa streamline chart with isotachs (blue). (b) 700-hPa streamline chart superimposed with vorticity (red–blue), a rainstorm trough line (short red solid line), a shear line (red-dashed line), a front, and a cold surgelike flow (dark blue arrow). (c) Surface streamline chart superimposed with TRMM rainfall (gold), a shearline (red-dashed line), a front, a cold surgelike flow (dark blue arrow), and the low pressure system marked by a red L. (d) IR image of MTSAT. (e) Surface wind vectors of QuikSCAT. (f) JMA surface analysis chart. Scales for isotachs, vorticity, and TRMM rainfall are shown on the top-right-hand panels of (a)–(c), respectively. The rainstorm is marked with a red oval in (c)–(e).

FIG. 5. Temporal evolution of the second rainstorm identified in the summer of 2008 from its genesis at 0800 UTC 30 May–0000 UTC 3 Jun 2008 (when a great flood occurred in southwestern Taiwan): (a) IR image of MTSAT superimposed with front, (b) 925-hPa streamline chart superimposed with TRMM rainfall (blue) and front, (c) longitude–height cross section of (ζ , ζ_t) (ζ and ζ_t are vorticity and vorticity tendency, respectively) cutting through the latitude of the rainstorm, and x – t diagrams of (d) TRMM rainfall P , (e) ζ (600 hPa), and (f) ζ (850 hPa) over the period 15 May–19 Jun 2008. Genesis location of rainstorm R1 is marked with a blue cross on an orographic map of Southeast Asia on the top panels of (d)–(f). The location of the rainstorm is marked with R2 in (a)–(c) and all five identified rainstorms in May–June 2008 are marked with R1–R5 in (d)–(f). In (c), the contour interval of ζ is $2 \times 10^{-5} \text{ s}^{-1}$ and $\zeta_t > 2 \times 10^{-10} \text{ s}^{-2}$ ($< -2 \times 10^{-10} \text{ s}^{-2}$) shaded by red (blue). Trajectories of five rainstorms R1–R5 are indicated by blue dot lines in the three x – t diagrams (bottom row) of (d)–(f).

streamline chart (Fig. 4c) or by QuickSCAT surface winds (Fig. 4e), if available.

- 5) Some MCS may be embedded/merged with a front, but the rainstorm in its initial stage is often located west of a JMA-analyzed front's tail (Fig. 4f).
- 6) Despite its genesis location, the trajectory of a rainstorm should reach the vicinity or east of Taiwan ($\sim 120^\circ\text{E}$) (Fig. 5a). The life cycle for this storm should be ≥ 2 days.
- 7) Measured by the size of the cyclonic vortex at the surface (Fig. 4e), the subsynoptic-scale disturbance in the midtroposphere (Fig. 4b), or the IR image of the Multifunction Transport Satellite (MTSAT)

(Fig. 4d), the horizontal scale for a rainstorm is smaller than a synoptic-scale disturbance [$\sim O(10^3)$ km], but larger than an MCS [$\sim O(10^2)$ km].

b. Backtracking procedure to determine genesis location of rainstorm

Rainstorms, recognizable weather disturbances, are not difficult to identify, except during their genesis stage. Several diagnostic tools with different data sources are applied to backtrack these storms to their initial locations. The first group of diagnostics consists of IR images for GMS/MTSAT (Fig. 5a), surface/925-hPa streamline

FIG. 5. (Continued)

charts along with precipitation (TRMM or GPCP; Fig. 5b), and the longitude–height cross sections of vorticity, which cuts through the genesis and propagation paths of the rainstorm (Fig. 5c). The second group of diagnostic tools are the longitude–time diagrams of rainfall (TRMM/GPCP; Fig. 5d) along the trajectory and vorticity at 600 hPa [ζ (600 hPa); Fig. 5e] and vorticity at 850 hPa [ζ (850 hPa); Fig. 5f] along the identified path of the rainstorms.

As arranged in Fig. 5, horizontal charts provide a clear view of the temporal–spatial evolution of a rainstorm. The spatial relationship between the rainstorm genesis location and Asian front on JMA surface maps provide a perspective different from the conventional view that rainstorms evolve from MCSs embedded in a mei-yu

front. The synoptic conditions of a rainstorm's genesis location and its downstream development may be depicted by streamline charts prepared with the NCEP GFS and CFS reanalyses. The QuikSCAT observation is formed with passes over the analyzed region twice daily. Thus, the existence of a rainstorm vortex may be confirmed with QuikSCAT surface wind vectors over the sea. Rainstorm genesis occurs in the midtroposphere (700–500 hPa). Therefore, the downward penetration of the rainstorm can be illustrated with longitude–height cross sections of vorticity and vorticity tendency. The propagation properties and downstream development of a rainstorm can be revealed from the longitude–time diagrams for precipitation P , ζ (600 hPa), and ζ (850 hPa).

TABLE 1. Dataset-I data: Surface and upper-air wind fields.

Dataset	Source	Spatial resolution	Temporal resolution	Data period	Source information
Reanalyses and Global Forecast System	NASA MERRA	1.25° (lon) × 1.25° (lat)	3 h	1979–present	Rienecker et al. (2008)
	NCEP GFS	0.5° lon × 0.5° lat	6 h	2006–present	Global Climate and Weather Modeling Branch, EMC, NCEP (2003, 2016)
Surface observation	NCEP CFSR	0.5° (lon) × 0.5° (lat)	6 h	1979–2011	Saha et al. (2010)
	QuikSCAT	12 km × 12 km 0.5° (lon) × 0.5° (lat)	Twice daily	1999–2009	Atlas et al. (2001)

with 6-h intervals. The genesis location for rainstorm R2 is also verified by the contrast between the longitude–time diagrams for ζ (600 hPa) and ζ (850 hPa) on the upstream side, since a ζ (600 hPa) center appears in Fig. 5e, as indicated by Fig. 5c, but does not have a corresponding ζ (850 hPa) center in Fig. 5f.

In addition to information revealed from Fig. 5, the following criteria are adopted to accurately identify the genesis location and time of rainstorm:

- 1) Following criterion 2 for the rainstorm identification in section 2a to track its genesis location, the embryo vortex should have maximum vorticity ($\zeta_{\max}$) $\geq 2 \times 10^{-5} \text{ s}^{-1}$ and maximum ζ_r (maximum vorticity tendency) $\geq 1 \times 10^{-10} \text{ s}^{-2}$, and $P \geq 0 \text{ mm day}^{-1}$ over a box of $5^\circ \times 5^\circ$ around the genesis location.
- 2) Synoptically, the rainstorm genesis is coupled with a midtropospheric subsynoptic-scale trough.

c. Data

A rainstorm is identified using criteria 2–6 with surface and upper-air winds, but assessed by all criteria including surface (WMO/GHCN) station and satellite (e.g., TRMM)

rainfall measurements, global precipitation analyses (e.g., PERSIANN, GPCP), and IR satellite images or TBB. Based on these identification criteria for rainstorms, the data sources utilized for the first group of criteria are designated as dataset I (Table 1) and the second group of criteria as dataset II (Table 2). Because the analysis period is close to four decades, it is difficult (if not impossible) to process all data sources in a uniform manner. In addition, the identification and depiction of a rainstorm need the horizontal resolution of data to be $\leq O(10^2) \text{ km}$. Proper use of all available data sources for the above manual identification criteria of rainstorms over three decades is a time-consuming task.

3. Statistics of rainstorm genesis

Using the identification criteria for rainstorms and the backtracking procedure to locate their genesis positions as outlined in section 2, a total of 297 rainstorms and their genesis locations are identified during the period 1979–2016 (Table 3): 248 geneses occurred over the southwestern China–northern Vietnam region and the northern South China Sea, and 49 geneses over the Bay

TABLE 2. Dataset-II precipitation: Station rainfall, analyzed global precipitation, and rainfall proxy derived from satellite observations.

Dataset	Source	Spatial resolution	Temporal resolution	Data period	Source information
Station rainfall	WMO surface stations	Over the region of interest, particularly Hong Kong and SW Taiwan, Hanoi (WMO 48820)	Hourly/3-h rainfall	1979–present	
Analyzed global precipitation	TRMM (3B42v7)	0.25° (lon) × 0.25° (lat)	3 h	1998–2016	Huffman and Bolvin (2015)
	PERSIANN	0.25° (lon) × 0.25° (lat)	Daily	1983–present	Ashouri et al. (2015)
	GPCP v2.2	2.5° (lon) × 2.5° (lat)	Daily	1979–present	Huffman and Bolvin (2013)
Rainfall proxy (IR image)	OLR	2.5° (lon) × 2.5° (lat)	Daily	1974/2006–present	Liebmann and Smith (1996)
	TBB (bright temperature of blackbody)	1° (lon) × 1° (lat)	3 h	1980–present	
	GMS-1–5	5 km × 5 km	1 h	Jul 1977–May 2003	Blersch and Probert (1991); Meteorological Services Centre Japan (1997)
	GOES-9	4 km × 4 km	1 h	May 1995–Feb 2005	Clark (1983)
	MTSAT-1–2	4 km × 4 km	1 h	Feb 2005–present	Faller (2005)

TABLE 3. Genesis occurrences of rainstorms over three regions (the southwest China–northern Vietnam region, the northern South China Sea, and the Bay of Bengal–Myanmar region), merger of rainstorms with mei-yu front west of 120°E, rainstorm across Taiwan, and merger of rainstorms with mei-yu front east of 120°E. Ratio of rainstorm merger with the mei-yu front west of 120°E, rainstorm population across Taiwan with genesis occurrence, and merger of rainstorms with the mei-yu front east of 120°E.

Region of rainstorm genesis	Genesis occurrence of rainstorm over region N/total	Occurrence of rainstorm genesis on front	Rainstorm merger with mei-yu front west of 120°E (merger with front/genesis occurrence)	Cross Taiwan (rainstorm across Taiwan/genesis occurrence)	Merger with mei-yu front east of 120°E (merger with front/rainstorm population across Taiwan)
Total	297	0	91 (31%)	204 (69%)	67 (33%)
Southwest China–northern Vietnam (I)	128 (43%)	0	50 (39%)	91 (71%)	35 (38%)
Northern South China Sea (II)	120 (40%)	0	28 (22%)	82 (68%)	19 (23%)
BOB–Myanmar (III)	49 (17%)	0	13 (27%)	31 (63%)	13 (42%)

of Bengal (BOB)–Myanmar region. Two characteristics of rainstorm genesis in the latter region are different from those over the former region:

- The rainstorm genesis occurrence ahead of the midtropospheric BOB trough is overlaid by the upper-level Tibetan anticyclone.
- To reach East Asia, the BOB–Myanmar rainstorms are propagated eastward by monsoon westerlies over a distance ≥ 2000 km through different weather systems across South Asia and northern Indochina.

For these reasons, the impact of the BOB trough on the genesis of the BOB–Myanmar rainstorms and the effects of different weather systems on the eastward propagation of these rainstorms will be reported in a separated study. The present study focuses only on the 248 rainstorm geneses. Salient features for these genesis locations are characterized as follows.

a. Geographic location of genesis

Genesis locations and trajectories of identified rainstorms are shown in Figs. 6a and 6b. Interestingly, the timing of the genesis is generally divided into two groups: afternoon–evening (0600–1200 UTC, red dots) and midnight–morning (1800–0000 UTC, blue dots). Most of the former group appear over land, while most of the latter group occur over the sea. This land–sea contrast of genesis timing should be modulated by the large-scale environmental flow through its response to the diurnal variation in radiative heating. The population of genesis (N_F) (Fig. 6c) exhibits triple maxima along the longitudinal direction. A major maximum, N_F , appears over land, particularly along the eastern slope of the mountains in the north Indochina–southwest China region, and two secondary maxima, N_F , in the Tonkin Bay–northern South China Sea and the BOB–Myanmar regions. A population distribution of rainstorm genesis over these regions is digitized in the first column of Table 3. A majority of the East Asian rainstorm genesis occurs east of 100°E, but about 1/6 over the BOB–Myanmar region. The dynamic impact of this land–sea differential radiative heating on rainstorm genesis will be addressed in later sections.

b. Vertical locations of genesis and the spatial relationship between genesis and fronts

The geneses of all identified rainstorms, whether over land or sea, only occur in the midtroposphere (700–500 hPa) (Fig. 6d). As with the genesis of the African easterly wave (e.g., Burpee 1972), there should be some special dynamic constraints of the monsoon circulation in Southeast/East Asia on the occurrence of rainstorm genesis. Shown in Figs. 5a and 5b, a typical rainstorm

genesis (30 May 2008) in northern Vietnam is located farther west than the East Asian front. Is this null relationship between the genesis and the front true for all rainstorm geneses? To clarify this concern, distances between genesis locations over land (red dots) and sea (blue dots) in Figs. 6a and 6b, respectively, and tails of these possibly related fronts appearing on the JMA surface charts are shown in the scatter diagrams of Fig. 6f. Some identified rainstorms during certain stages of their life cycle may merge with fronts at downstream locations (solid/open triangles in Fig. 6e) east of their genesis locations. The relationships between rainstorm genesis from three regions and fronts, and the merger of these rainstorms and fronts are summarized in Table 3 in terms of the rainstorm population from these regions.

4. Favorable environment for rainstorm genesis

We focus on rainstorm genesis occurrence over two maximum regions (Fig. 6a): the northern Vietnam–southwestern China region and the northern South China Sea. From a climatological perspective, we first examine the large-scale environment favorable for the occurrence of rainstorms in these two regions. The Charney–Stern instability criterion is then tested to determine whether the large-scale circulation over these two regions satisfy this criterion. Finally, the synoptic condition for a typical case of rainstorm genesis in each preferred region will be depicted and inferences of the rainstorm genesis mechanism substantiated with the diagnostic analysis of the vorticity and heat budgets.

a. Large-scale monsoon circulation during the mei-yu season

After monsoon onset in Southeast/East Asia, the monsoon circulation during the active phase (16 May–15 June) (Fig. 7) is characterized by the following features. The upper-level strong westerlies migrate across the Tibetan Plateau and the tropical easterly jet develops around the southern rim of the Tibetan high. In contrast, juxtaposed with the western North Pacific subtropical high, the Asian continental thermal low embedded by the Tibetan Plateau is surrounded by the monsoon westerlies in South Asia and the monsoon southwesterlies in Southeast/East Asia.

The latitude–height cross section of the zonal wind component at 112.5°E (Fig. 7) shows the lower-tropospheric monsoon westerlies across the northern part of Southeast Asia merge with the midtropospheric westerlies of the short-wave trough along the eastern seaboard of East Asia. As depicted by the 700- and 500-hPa streamline charts in Fig. 7, a confluence forms by these westerly flows over East Asia. To 700 hPa, this

confluence forms a positive shear flow, extending from southwest China to the Yangtze River valley. The dynamic implication of this *shear flow* may be illustrated by the local Hadley circulation [depicted by $(v_D, -\omega)$] at two longitudes (104° and 114°E) (Fig. 8), where v_D is the meridional divergent wind. These are the locations where the maximum population of the rainstorm genesis (Fig. 6c) occurs in the middle troposphere. The upward branches of these local Hadley circulations are centered slightly north of 20°N across the shear line. The coincidence between the upward branch of the local Hadley circulation and the midtropospheric positive shear line provide a favorable environment for the genesis of rainstorms. This environmental condition may be coherently intensified by 1) the occurrence of cold surgelike northwesterlies east of the Tibetan Plateau in tandem with the eastward propagation of a midlatitude short-wave trough across the eastern seaboard of East Asia and 2) the monsoon southwestlies across Indochina. The confluence of these two intensified flows in the midtroposphere is a favorable environment for rainstorm genesis and will be illustrated by case analyses in the next section.

b. Dynamic instability

As demonstrated by Burpee (1972), the genesis of African easterly waves occurring along the southern side of the African easterly jet is attributed to Charney–Stern instability (Charney and Stern 1962): a reversal of the meridional gradient q_y of Ertel potential vorticity q occurs along the south flank of a baroclinic jet and $q_y < 0$ underneath this jet coupled with a maximum θ_y (meridional gradient of potential temperature θ) near the surface (Thorncroft and Hoskins 1994).

It was previously illustrated that the western end of the positive monsoon shear flow (between northern Vietnam and eastern China) provides an environment convectively and dynamically favorable for rainstorm genesis. From a dynamic perspective, it is important to understand why the possible mechanism of rainstorm genesis exists in the midtroposphere. Suggested by the contrast between locations of rainstorm genesis (Figs. 6a,c) and the environmental circulation (Fig. 7) where the rainstorm geneses are embedded, the Charney–Stern instability is a likely mechanism. The maximum rainstorm genesis occurrence (Fig. 6b) appears at two longitudes: 104°E over the northern Vietnam–southeastern China region and 114°E over the northern part of the South China Sea. The latitude–height cross sections of (q_y, u) are shown in Figs. 8a and 8c. South of the maximum westerlies a sign change of q_y occurs in the midtroposphere. Potential temperatures near the ground at both longitudes (Figs. 8b,d) exhibit a positive meridional

FIG. 6. Genesis statistics for rainstorms: (a) locations of genesis and trajectories for all identified rainstorms east of 100°E. (b) As in (a), but for all identified rainstorms west of 100°E. (c) Accumulated genesis occurrence for each degree of longitude. (d) Vertical locations of all identified geneses. (e) Merging locations (triangles) of rainstorms and East Asian fronts. (f) Distance between the genesis location and the tail of East Asian fronts for all identified rainstorms over the southwest China–northern Indochina region (0600–1200 UTC; red dots; land) and over the northern South China Sea (1800–0000 UTC; blue dots; sea). Red and blue dots in (a) represent genesis locations that occurred during late evening and early morning, respectively. The trajectories of rainstorms are indicated by color lines corresponding to their genesis in (a) and (b). The merging location of rainstorms and East Asian front in (e) is indicated by solid and open triangle in red (southwest China–northern Indochina rainstorm) and blue (northern SCS rainstorm).

gradient, $\theta_y > 0$, slightly north of the latitude where q_y changes its sign. These two conditions satisfy the Charney–Stern instability criterion. The upward branches of the local Hadley circulations at the two longitudes (Figs. 8b,d) indicate the uplift of the air mass is coincident with the dynamically unstable zone, which facilitates rainstorm genesis.

c. Case analysis

Inferred from Figs. 4 and 5, the rainstorm genesis may occur when the northwesterly flow east of the Tibetan

Plateau is intensified by an eastward-propagating mid-latitude synoptic short-wave trough located along the eastern seaboard of East Asia and the monsoon southwesterly flow is strengthened by the intensification of the monsoon circulation. To support this argument, the longitude–time diagrams of 700-hPa wind vectors at 40° and 25°N superimposed with rainstorm tracks (dark-blue dots) are shown in Figs. 9a and 9b, respectively. These figures show rainstorm genesis occurs when pulses of northwesterly wind (blue arrow) and southwesterly flows (red arrow) begin to intensify. Thus, a positive

FIG. 7. Monthly (16 May–15 Jun) averaged large-scale circulation depicted by streamlines at different levels for the period 1979–2016. Isochrons are colored yellow and orange with the scale shown in the lower-right-hand corner of each panel. The trough line is marked by thick red-dashed line. Blue H and red L denote anticyclone and trough/cyclone, respectively. The latitude–height cross section of westerlies (golden color) and streamfunction (contours) at 112.5°E are shown in the right-hand panel. The scale for u (112.5°E) is shown in the lower-right-hand corner of this panel. Departures of northwesterlies ΔV_{NW} in the northern red square and southeasterlies ΔV_{SE} in the southern red square from their corresponding climatological mean values are used to prepare Fig. 10.

shear flow is developed by the confluence of these two flows to establish an environment favorable for rainstorm genesis. Can the argument presented here be universally applicable to all 248 identified rainstorm geneses. Let ΔV_{NW} and ΔV_{SW} be departures from the mean Tibetan northwesterlies and the mean monsoon southwesterly flow at 700 hPa, respectively. All scatters (dots) showing wind departures are located over this region, where $\Delta V_{NW} > 0$ and $\Delta V_{SW} > 0$ in Fig. 10, imply the confluence of the intensified Tibetan northwesterly and monsoon southwesterlies in the mid-troposphere is universally applicable for the identified rainstorm genesis. Thus, positive shear flow developed by the confluence of these two flows establishes an environment favorable for rainstorm genesis. Synoptic conditions for a typical rainstorm genesis in northern

Vietnam and another in the northern South China Sea are analyzed in detail below to illustrate the dynamic/thermodynamic processes involved.

1) CASE 1: RAINSTORM GENESIS IN NORTHERN VIETNAM ON 30 MAY 2008

(i) Synoptics

On 30 May 2008, a well-developed deep 200-hPa short-wave trough over the eastern seaboard of northeast Asia was separated from the Southeast Asian anticyclone by the east Asian jet (Fig. 11a). Trained by a ridge, this trough couples with a low-level, high-low couplet (Figs. 11b,c). A cold surgelike northeasterly flow, coupled with the upper-level northwesterlies ahead of the ridge, straddles the high and low centers of

FIG. 8. Latitude–height cross sections of (a) $(\partial q / \partial y, u)$ (104°E), where q is potential vorticity; (b) $[\theta, (v_D, -\omega)]$ (104°E); and (c), (d) the corresponding cross sections at 114°E . The contour interval of $\partial q / \partial y$ and the scale of u are shown at the top right of (a) and (c) and at the bottom right of (c), respectively. Note, q is potential vorticity. Positive values of $\partial q / \partial y$ are shaded red. The color scale for θ is located at the bottom right of (d). All variables shown in this figure are averaged over the May–June period of 1979–2016.

this couplet. The confluence of the midtropospheric northwesterlies and the monsoon southwesterlies develop a cyclonic shear flow. The western end of this shear flow becomes a breeding ground for rainstorm genesis. This argument is confirmed by a low pressure region shown on the JMA surface chart marked by a red “L” (Fig. 11d), and by the IR image of MTSAT with an R2 symbol (Fig. 11e) in northern Vietnam. Additionally, the confluence depicted by the GFS analyses is validated by the wind vectors measured using radiosondes (Fig. 11f)—northwesterlies above 700 hPa at Hohhot, China (ZBHH), and westerlies in the lower troposphere at King Park (45004). Significant dewpoint depression exists at Hohhot over the entire troposphere within the cold surgelike flow, but almost disappears in the midtroposphere at King Park. This thermal contrast indicates the midtroposphere within the cyclonic shear flow is convectively unstable. Another synoptic feature is that the rainstorm genesis and the JMA-analyzed front are separated by a distance of 1000 km (Fig. 11d). The rainstorm genesis occurs west and independent of the front. Apparently, this rainstorm is not evolved from the MCS associated with the East Asian front. The synoptic conditions for this Vietnam rainstorm genesis are typical.

(ii) Heat and vorticity analyses

The thermodynamic equation (e.g., Holton and Hakim 2013) is

$$\frac{\partial T}{\partial t} + \mathbf{V} \cdot \nabla T - s_p \omega = \frac{1}{c_p} \dot{Q}, \quad \text{where} \quad s_p = -\frac{T}{\theta} \frac{\partial \theta}{\partial p}. \quad (1)$$

If $\partial T / \partial t$ (local rate of change of temperature) is smaller than other thermodynamic processes in Eq. (1), this equation may be approximated by

$$-s_p \omega \approx \frac{1}{c_p} \dot{Q} - \mathbf{V} \cdot \nabla T. \quad (2)$$

As revealed from Eq. (2), vertical motion ($-\omega$) can be initiated/maintained by diabatic heating $[(1/c_p)\dot{Q}]$ and thermal advection ($-\mathbf{V} \cdot \nabla T$). Note, diabatic heating $[(1/c_p)\dot{Q}]$ may be evaluated in terms of the *residual term* in Eq. (1) (e.g., Chen and Baker 1986). All thermodynamic processes involved with the R2 rainstorm genesis (Fig. 12) are highlighted by the following features:

- The diabatic heating $(1/c_p)\dot{Q} > 0$ (Fig. 12b) at the genesis location of rainstorm R2, and over its vicinity,

FIG. 9. The longitude–time diagram of 700-hPa wind vectors $\mathbf{V}$ (700 hPa) at (a) 40° and (b) 25°N. The longitudinal locations of all identified rainstorms (indicated by R1–R5) during the period mid-May–June 2008 are marked by dark-blue dots. Northwestery or northerly (southwesterly or southerly) wind vectors are represented by blue (red) vectors. The scale of wind vectors is shown on the top right of each panel.

while $\partial T/\partial t$ (Fig. 12e) is small/negative. The magnitude ratio $|\partial T/\partial t|/|\dot{Q}/c_p|$ around the R2 genesis location (Fig. 12d) is $<10\%$.

- Shown in Fig. 11c, rainfall occurs along the front, but becomes insignificant over the narrow trough (accommodating the rainstorm embryo of R2 with the symbol L) west of the front.
- The close resemblance between the distributions of $\dot{Q}/c_p$ and $-s_p\omega$ (Fig. 12f) strongly indicates that the vertical motion along the narrow trough and the front are a response of the air mass to the diabatic heating

generated from the heat of condensation from water vapor, supplemented by the sensible heat advection, $-\mathbf{V} \cdot \nabla T$ (Fig. 12a).

It is inferred from the approximated heat budget [Eq. (2)] that the spinup process for the R2 genesis is driven through the diabatic cooling ($-s_p\omega$) by the collaborative effect of $(1/c_p)\dot{Q}$ and $-\mathbf{V} \cdot \nabla T$ (Fig. 12c) along the positive shear zone and the front through the vortex stretching illustrated in terms of the vorticity equation.

FIG. 10. A scatter (black dot) diagram formed using northwesterly departures from the climatological values $\Delta V_{NW}(700\text{ hPa})$ averaged over the red square (centered at 40°N , 112.5°E) in Fig. 7, northeast of the Tibetan Plateau and monsoon southwesterly departures from the climatological value $\Delta V_{SW}(700\text{ hPa})$ averaged over the red square (centered at 20°N , 112.5°E) in Fig. 7 in southern China.

The complete vorticity equation (e.g., Holton and Hakim 2013) can be written as follows

$$\frac{\partial \zeta}{\partial t} = \underbrace{-\mathbf{V} \cdot \nabla(\zeta + f)}_{\text{VAD}} - \underbrace{(\zeta + f) \nabla \cdot \mathbf{V}}_{\text{VST}} - \underbrace{\omega \frac{\partial \zeta}{\partial p} - \frac{\partial \omega}{\partial x} \frac{\partial v}{\partial p} + \frac{\partial \omega}{\partial y} \frac{\partial u}{\partial p} + \frac{\partial F_y}{\partial x} - \frac{\partial F_x}{\partial y}}_{\text{VRE}}, \quad (3)$$

where ζ_t (tendency of relative vorticity), VAD (advection of absolute vorticity), VST (stretching of absolute vorticity), and VRE (residual term of the vorticity equation) for the genesis of a R2 rainstorm are shown in Fig. 13. VRE (Fig. 13d) is smaller in magnitude than individually with the other two dynamic processes: VAD and VST. The magnitude of ζ_t (Fig. 13a) is close to $-(\text{VAD} + \text{VST})$ (Fig. 13d), because VRE/ζ_t around the R2 genesis location is smaller than 10%. Thus, Eq. (3) can be simplified into Haynes and McIntyre's (1987) flux form for the vorticity equation:

$$\zeta_t \approx -\nabla \cdot [\mathbf{V}(\zeta + f)]. \quad (4)$$

Although the rainstorm propagation mechanism through VAD is not explicitly provided, ζ_t is simply a response of the local vorticity tendency to the convergence of total vorticity flux. Another insight for the

spinup of the R2 genesis can be revealed from the comparison between dynamic processes of VAD (Fig. 13b) and VST (Fig. 13c) over the area around the genesis location (indicated by the intersection of two thin red lines). Clearly, ζ_t is primarily contributed by VST through the vertical motion driven by diabatic heating illustrated by the heat budget of the R2 genesis shown in Fig. 12.

2) CASE 2: RAINSTORM GENESIS IN THE NORTHERN SOUTH CHINA SEA ON 4 MAY 1994

(i) Synoptics

A secondary maximum of rainstorm genesis occurrence frequency appears in the northern South China Sea (Fig. 6b). The synoptic condition of a typical rainstorm genesis for this region at 0600 UTC 4 May 1994 is shown in Fig. 14. An upper-level short-wave trough (thick red line) located along the eastern seaboard of northeast Asia was separated from the east–west-oriented South–Southeast Asian anticyclone by the subtropical westerly jet (Fig. 14a). The upper-level ridge–trough synoptic system was coupled with a low-level high–low couplet straddled by a cold surgelike northwesterly flow (Figs. 14b,c). Along the southeastern China coast in the low troposphere, the confluence of the midtropospheric northwesterly flow and the monsoon southwesterly flow form a narrow trough slightly west of Hainan Island (Figs. 14b,c). This narrow trough is verified with the radiosonde observations of wind vectors at Taiyuan, China (ZBYN), and King Park below 500 hPa (Fig. 14f). The northwesterly flow is dry (significant dewpoint depression), while the monsoon southwesterly flow is moist (dewpoint depression $\sim 0^\circ\text{C}$). A merger of these two air masses along this narrow trough made the flow convectively and dynamically unstable. Identified by PERSIANN precipitation (R in Fig. 14c) and the TBB image of GMS observations (R in Fig. 14e), a rainstorm genesis was located at approximately 400 km west of the tail of the East Asian front (shown on the JMA surface chart in Fig. 14d).

(ii) Heat and vorticity budget analysis

As the R2 rainstorm genesis in northern Vietnam, $(\partial T/\partial t)/(1/c_p)Q$ (Fig. 15d) is $\ll 1$ at the genesis location of the SCS R rainstorm and over its vicinity. The close resemblance in the spatial distribution of $[(1/c_p)Q(-\mathbf{V} \cdot \nabla T)]$ (Fig. 15c), and $-s_p\omega$ (Fig. 15f) reveals the vertical motion along the narrow trough and the front is a response of the lower-tropospheric flow to the diabatic heating generated by the heat condensation of the cloud droplet (Fig. 15b), supplemented by the weak sensible heat advection $-\mathbf{V} \cdot \nabla T$ (Fig. 15a). To satisfy mass conservation, the low-level convergence and

FIG. 11. Synoptic conditions of the rainstorm genesis that occurred at 0600 UTC 30 May 2008: (a) 200-hPa streamline chart superimposed with isotachs (blue) and a trough line (thick red line). (b) As in (a), but at 700 hPa and superimposed with a front, a shearline (red-dashed line), a dark-blue arrow (northwesterly flow) with a blue H and red L to the arrow's SW and NE side, respectively, and a red arrow (southwesterly flow). (c) As in (b), but at 850 hPa with TRMM precipitation (blue) and the genesis location of rainstorm 2 (R2). (d) JMA surface analysis chart. (e) IR image of MTSAT superimposed with front and the R2 genesis location. (f) Radiosonde observations at Hohhot (ZBHH) and King Park (45004). Vertical profiles in (f) are dewpoint (blue) and temperature (red).

upper-level divergence of the air mass should be developed by the upward motion generated by diabatic heating associated with the rainstorm genesis. The ratio, $VRE/\zeta_i \ll 1$, is around the genesis location of the SCS rainstorm (Fig. 16f). Because the VAD process (Fig. 16b) is insignificant around this rainstorm genesis location, the major contribution to ζ_i comes from VST (Fig. 16c). Shown in Fig. 16d it matches well with ζ , the spinup process for the SCS rainstorm genesis apparently is

caused by the convergence of total vorticity flux driven by the vertical motion (Fig. 15f), as the thermodynamic and dynamic processes illustrated for the genesis of case 1.

d. Summary

The confluence of the cold surgelike northwesterly flow east of the Tibetan Plateau and the monsoon southwesterly flow develops a narrow trough. The

FIG. 12. The 700-hPa heat budget analysis for the rainstorm R2 genesis occurrence at 0600 UTC 30 May 2008: (a) the 700-hPa streamline charts superimposed with $-\mathbf{V} \cdot \nabla T$ (700 hPa), (b) $(1/c_p)\dot{Q}$, (c) [(a) + (b)], (d) [(e)/(b)], (e) $\partial T/\partial t$ (700 hPa), and (f) $-s_p\omega$ (700 hPa). The genesis location marked by red L and front are added at each panel, except for (d). Color scales for all variables are shown at the top right of each panel. Two thin red lines are added on each panel to indicate the identified location of rainstorm genesis.

western end of this narrow trough becomes a breeding ground for rainstorms during the active phase of the Southeast/East Asian summer monsoon. The spinup of air mass through vortex stretching/convergence of total vorticity flux at the western end of this trough makes rainstorm genesis possible by upward vertical motion. To ascertain these findings are representative for all identified rainstorm geneses, two three-dimensional scatter diagrams are shown in Fig. 17:

- (i) The heat budget analysis (Fig. 17a): Upward vertical motion ($-\omega > 0$) in the vicinity of rainstorm genesis is primarily maintained by diabatic heating, $\dot{Q}/c_p$.
- (ii) The vorticity budget analysis (Fig. 17b): In the vicinity of rainstorm genesis, the spinup of the rainstorm embryo ($\zeta_r > 0$) is induced/maintained by an upward vertical motion ($-\omega > 0$) through positive

vortex stretching [$-(\zeta + f)\nabla \cdot \mathbf{V} > 0$] or convergence of total vorticity flux $-\nabla \cdot [(\zeta + f)\mathbf{V}]$.

5. Diurnal variation of rainstorm genesis

The mechanism of rainstorm genesis was illustrated from two perspectives: 1) large-scale environmental flow favorable for rainstorm genesis and 2) dynamic instability of the environmental flow. As shown in Fig. 6a, most rainstorm geneses on land occur in the late evening and at sea in the early morning. It was inferred from this diurnal variation in the timing preference of rainstorm genesis is regulated by the diurnal variation of the large-scale monsoon circulation in response to the land-sea thermal contrast and the diurnal variation of dynamic instability of the environmental flow.

FIG. 13. The 700-hPa vorticity budget analysis at the genesis occurrence of rainstorm R2 (0600 UTC 30 May 2008): The 700-hPa streamline charts superimposed with (a) ζ , (700 hPa), (b) $-\mathbf{V} \cdot \nabla(\zeta + f)$ (700 hPa), (c) $-(\zeta + f) \nabla \cdot \mathbf{V}$ (700 hPa), (d) $[(b) + (c)]$, (e) $\text{VRE}\{a - [(b) + (c)]\}$, and (f) $[(e)/(a)]$ without the 700-hPa streamline chart. As in Fig. 12, two thin red lines are added on each panel to indicate the identified location of rainstorm.

a. Diurnal variation of the large-scale circulation in response to the land–sea thermal contrast

The direct atmospheric response to diabatic heating is the development of divergent circulation. Because vertical motion is part of divergent circulation, modulation of the divergent circulation through the land–sea thermal contrast should be reflected by vertical motion. As shown in section 4, rainstorm genesis is facilitated by vertical motion through vortex stretching/convergence of vorticity flux. Thus, the timing preference of the rainstorm genesis should be regulated by the diurnal variation of the land–sea thermal contrast through the divergent circulation. This inference can be substantiated by diagnostic analyses, including maintenance of divergent circulation by diabatic heating (Chen and Yen 1991 a,b) and the streamfunction tendency by vortex stretching (Sanders 1984; Chen and Chen 1997).

Divergent circulation can be well portrayed with velocity potential χ , while divergent winds can be measured with horizontal gradients of χ (i.e., $\mathbf{V}_D = \nabla\chi$). The maintenance of the divergent circulation by diabatic heating $\dot{Q}$ is illustrated by the χ -maintenance equation, a combination of the continuity and the thermodynamic equations:

$$\chi = \nabla^{-2} \left[\frac{\partial}{\partial p} \left(\frac{\partial T}{\partial t} + \mathbf{V} \cdot \nabla T \right) \right]_{\chi_H} + \nabla^{-2} \left\{ \frac{\partial}{\partial p} \left[\frac{1}{\sigma c_p} \left(\frac{T}{\theta} \dot{Q} \right) \right] \right\}_{\chi_{\dot{Q}}}, \quad (5)$$

where c_p , T , θ , p , and $\mathbf{V}$ are specific heat with constant pressure, temperature, potential temperature, pressure, and velocity vector, respectively. The diurnal component ($\sim$) of the divergent circulation is maintained by

FIG. 14. (a)–(f) As in Fig. 11, but for 4 May 1994 with PERSIANN precipitation in (c), and TBB, instead of IR, is used to generate images of convective clouds in (e). The genesis location of this rainstorm is indicated with a red-open arrow marked by R. One sounding station was replaced by Taipei.

$$\hat{\chi} = \hat{\chi}_H + \hat{\chi}_Q. \quad (6)$$

The relationship between vorticity ζ and streamfunction ψ is $\zeta = \nabla^2 \psi$ or $\psi = \nabla^{-2} \zeta$. The streamfunction budget equation (i.e., an inverse Laplace transformation of the vorticity equation) may be written as

$$\frac{\partial \psi}{\partial t} = \nabla^{-2} \left[-\mathbf{V} \cdot \nabla (\zeta + f) \right] + \nabla^{-2} \left[(\zeta + f) \frac{\partial \omega}{\partial p} \right], \quad (7)$$

where $\partial \omega / \partial p = -\nabla \cdot \mathbf{V}$ (the continuity equation). According to the Helmholtz theorem, the horizontal

atmospheric flow can be split into divergent and rotational flows, $\mathbf{V} = \mathbf{V}_D + \mathbf{V}_R$. Because $|\mathbf{V}_D|/|\mathbf{V}_R| \ll 1$, the horizontal atmospheric flow is primarily depicted by the rotational component. Maintenance of the diurnal component of the rotational flow may be expressed as

$$\hat{\psi}_t = \hat{\psi}_A + \hat{\psi}_\chi. \quad (8)$$

Presumably, ζ/f (f = planetary vorticity) $\ll 1$ is true for the diurnal component of global circulation around 20°N . For the global $\hat{\psi}$ field, the streamfunction tendency caused by the vorticity advection $\hat{\psi}_A$ may not be significant, and Eq. (8) may be approximated as

FIG. 15. As in Fig. 12, but the rainstorm genesis occurred at 0600 UTC 4 May 1994.

$$\hat{\psi}_t \simeq \nabla^{-2}(-f \nabla \cdot \hat{\mathbf{V}}). \quad (9)$$

Assuming $f = f_0$ (constant), the diurnal mode of the streamfunction budget may be expressed as

$$\hat{\psi}_t \simeq -f_0 \hat{\chi}. \quad (10)$$

As shown in Eqs. (9) and (10), the diurnal mode of the atmospheric rotational circulation is largely driven by its divergent circulation.

The timing preference of rainstorm genesis between land and sea (Fig. 6a) will be illustrated with Eqs. (6) and (8). Over the analysis domain, the variance of $\hat{\chi}$ (850 hPa) contributed by $\hat{\chi}_H$ (850 hPa) and $\hat{\chi}_Q$ (850 hPa) are 8% and 92%, respectively. An empirical orthogonal function (EOF) analysis was then performed with $\hat{\chi}_Q$ (850 hPa) to illustrate the maintenance of the divergent circulation by diabatic heating. The first two principal eigenmodes, including eigenvectors and time series of eigencoef-ficients, are shown in the left-hand column of Fig. 18. The

land-sea thermal contrast is well portrayed by the first eigenvector E1 [$\hat{\chi}_Q$ (850 hPa)], which explains 63% of the variance of $\hat{\chi}_Q$ (850 hPa). The time series of this eigenmode's eigencoef-ficient C1 [$\hat{\chi}_Q$ (850 hPa)] reaches its maximum at 0300 UTC and minimum at 1500 UTC. Surprisingly, north-south differential heating also plays an important role, as depicted by the second eigenvector E2 [$\hat{\chi}_Q$ (850 hPa)], which explains 35.5% of the variance of $\hat{\chi}_Q$ (850 hPa), with its maximum eigenvalue C2 [$\hat{\chi}_Q$ (850 hPa)] occurring at 0900 UTC and its minimum eigenvalue occurring at 2100 UTC.

The EOF analysis of $\hat{\chi}$ (850 hPa) is shown in the right-hand column of Fig. 18. Temporal variations of the eigencoef-ficient time series of these two (850 hPa) eigenmodes (Fig. 18d) and the spatial structure for the two eigenvectors of $\hat{\chi}$ (850 hPa) (Figs. 18e,f) resemble the corresponding $\hat{\chi}_Q$ eigenmodes (Figs. 18b,c). Apparently, the diurnal variation of the regional divergent circulation is a response of the Asian monsoon system to both the land-sea thermal contrast and the

FIG. 16. As in Fig. 13, but the rainstorm genesis occurred at 0600 UTC 4 May 1994.

north–south differential heating. A daily clockwise rotation of $\hat{\chi}$ (850 hPa) around the Asian continent and the western North Pacific is formed by a combination of its two eigenmodes (not shown).

The clockwise rotation of $\hat{\chi}$ (850 hPa) over the Asian continent leads to the formation of the east–west circulation ($\hat{u}_D$, $-\hat{\omega}$) (20°N) between land and ocean shown in Fig. 19. At 1200 UTC (late evening, the left-hand column of Fig. 19), the upward branch of the east–west circulation over Indochina is coupled with low-level convergence and upper-level divergence, while the downward branch of the east–west circulation in the South China Sea is linked with low-level divergence and upper-level convergence. The structure of this east–west circulation facilitates the occurrence of rainstorm genesis over the northern Vietnam–southwestern China region and hinders it over the northern South China Sea. At 0000 UTC (early morning, the right-hand column of Fig. 19), the land–sea contrast of vertical motion, the associated

divergent circulation, and the rainstorm genesis occurrence frequency are reversed.

Next, we shall explore *dynamically* how this diurnal variation in vertical motion modulates the rainstorm genesis activity between land and ocean. Using Eq. (9) or (10), we argued the diurnal variation of the regional streamfunction follows the diurnal variation of velocity potential across the Indochina–southwestern China region and the northern part of the South China Sea. The diurnal component of the streamfunction budget at 0900 UTC (1600 LST) and 2100 UTC (0400 LST) are shown in Fig. 20. The variances of $\hat{\psi}_i$ (850 hPa) (computed by the central finite-difference approach) explained by $\hat{\psi}_A$ (850 hPa) and $\hat{\psi}_\chi$ (850 hPa) are 8% and 92%, respectively. Thus, $\hat{\psi}_i$ (850 hPa) is close to $\hat{\psi}_\chi$ (850 hPa) in spatial structure and magnitude. The negative $\hat{\psi}_i$ (850 hPa)/ $\hat{\psi}_\chi$ (850 hPa) center over Indochina and southern China at 0900 UTC facilitates the evening rainstorm genesis. During this time, values for $\hat{\psi}_i$ (850 hPa)/ $\hat{\psi}_\chi$ (850 hPa) over the South China Sea are close to zero, unfavorable for evening

FIG. 17. Three-dimensional scatter diagrams. (a) The heat budget at 700 hPa: upward vertical motion ($-\omega \geq 0$) around the genesis location of rainstorms is generated/maintained by the heat advection $-\mathbf{V} \cdot \nabla T$ and diabatic heating $[(1/c_p)\dot{Q}]$. (b) The vorticity budget at 700 hPa: convergence of total vorticity flux $-\nabla \cdot [(\zeta + f)\mathbf{V}]$ induced by upward motion ($-\omega > 0$) in the vicinity of rainstorm genesis results in cyclonic spinup ($\zeta_t > 0$) of the environmental flow which facilitates genesis. The two-dimensional scatter diagrams are formed by two variables; the smaller dots of different colors (green, blue, and black) are projections of the three-dimensional scatters (large red balls) on the two-dimensional walls along one of the three directions on the Cartesian coordinate.

rainstorm genesis. In contrast, the reverse conditions occur at 2100 UTC. The streamfunction budget for the diurnal mode over the concerned region demonstrates the impact of the vortex-stretching process on the genesis activity of rainstorms through the diurnal mode of the regional atmospheric circulation.

b. Diurnal variation of dynamic instability

The clockwise rotation of the continent-scale divergent circulation (Figs. 18, 19) in response to the east–west continent–ocean thermal contrast and the north–south thermal differential heating can drive

the clockwise rotation of the continental monsoon circulation primarily through vortex stretching (Fig. 20). Can the diurnal variation of the continental divergent and rotational circulation lead to the diurnal variation of the Charney–Stern instability criterion and result in the preferred timing of the rainstorm genesis over northern Indochina in the evening and over the South China Sea in the morning?

This question may be answered by the latitude–height cross section of (q_y, u) at 104°E (land) and 114°E (sea) for two 6-h windows (0900–1500 and 2100–0300 UTC) centered at 1200 UTC (1900 LST) and 0000 UTC (0700 LST), respectively, shown in the left- and right-hand columns of Fig. 21. A sign change of q_y appears between 20° and 22°N, south of maximum westerlies in both time windows (Figs. 21a,b,e,f), while the corresponding $\bar{\theta}_y$ shows a maximum meridional gradient in the lower troposphere (Figs. 21d,h) underneath $q_y < 0$ north of the latitude where q_y changes its sign (Figs. 21c,g). The Charney–Stern instability criterion at both longitudes (104° and 114°E) is met by the environmental flow.

Diurnal change of q_y and θ_y at the two longitudes between the two time windows are shown in Figs. 21c and 21d and Figs. 21g and 21h, respectively. Two salient features emerge from these two sets of figures:

- 1) The sign change in Δq_y (104°E) close to 20°N (Fig. 21c) is the same as q_y (104°E) (Fig. 21a), but the sign change in Δq_y (114°E) (Fig. 21g) is opposite q_y (114°E) (Fig. 21e).
- 2) The maximum (positive) value of $\Delta \theta_y$ (104°E) (Fig. 21d) appears underneath Δq_y (104°E) < 0 north of 20°N, while the minimum (negative) value of $\Delta \theta_y$ (114°E) (Fig. 21h) exists underneath Δq_y (114°E) > 0 north of 20°N.

These features indicate that rainstorm genesis is *more* likely to occur during the 0900–1500 UTC (evening) window at 104°E (land) and during the 2100–0300 UTC (early morning) time window at 114°E (sea). In other words, when the land is warmer than the air in the evening, the Charney–Stern instability criterion is met easier by the environmental flow over the land. The reverse situation is applied to the sea. The impact of this land–sea thermal contrast on the timing preference of rainstorm genesis is reflected by the Charney–Stern instability. The response of the basic flow to the land–sea thermal contrast should undergo a diurnal variation in both the (u, θ) fields; well depicted by the daily-mean component of the basic flow $(\bar{u}, \bar{\theta})$ added with their diurnal components $(\hat{u}, \hat{\theta})$. Note, the generations of the $(\bar{u}, \hat{\theta})$ component and the comparisons between

FIG. 18. Empirical orthogonal function (EOF) analysis of $\hat{\chi}_Q$ (850 hPa) and $\hat{\chi}$ (850 hPa): eigenvalue time series of the first two eigenmodes (a) $\hat{\chi}_Q$ (850 hPa) and (d) $\hat{\chi}$ (850 hPa) are shown. The first two eigenvectors of both (b), (c) $\hat{\chi}_Q$ (850 hPa) and (e), (f) $\hat{\chi}$ (850 hPa) are shown. Total variance explained by the two eigenmodes are represented by the percentage parenthesized behind C1 and C2, on the top of (a) and (d).

$(\bar{u}, \bar{\theta})$ and $(\bar{u} + \hat{u}, \bar{\theta} + \hat{\theta})$ are presented in the online supplemental material. The $\Delta(\partial q / \partial y)$ measured with $(\bar{u} + \hat{u}, \bar{\theta} + \hat{\theta})$ should resemble that measured with $(\bar{u} + \hat{u}, \bar{\theta} + \hat{\theta})$ at both longitudes 104° and 114°E shown in Figs. 21c and 21g, and $\Delta\theta_y$ shown in Figs. 21d and 21h. The generations of $(\partial q / \partial y)$ $(\bar{u}, \bar{\theta})$ and $(\partial q / \partial y)$ $(\bar{u} + \hat{u}, \bar{\theta} + \hat{\theta})$, and the comparisons between are also presented in the online supplement.

6. Conclusions

During the active summer monsoon phase (mid-May–late June) over the northern part of Southeast Asia and the southern part of East Asia, monsoon rain

is primarily produced by rainstorms developed from midtropospheric cyclones (MTCs). It has been regarded by the East Asian meteorological community that rainstorms during the monsoon phase originate/develop from mei-yu fronts. In fact, reviewing the JMA surface analysis charts, we find rainstorm genes occur west of the mei-yu fronts. Even five rainstorms identified by the LSESREX and the SoWMEX/TiMREX have their genes independent of the mei-yu fronts. To substantiate our argument, an identification scheme for rainstorm genesis with a two-step procedure was introduced: 1) rainstorm identification and 2) backtracking of the identified rainstorm to its genesis location. For the period 1979–2016,

FIG. 19. Composite diurnal mode of divergent circulation ($\bar{V}_D, \bar{\chi}$) in the (a),(d) upper level (400 hPa) and (c), (f) lower level (850 hPa) at 1200 UTC (1900 LST) and 0000 UTC (0700 LST), respectively. (b),(e) The corresponding east-west circulation ($\bar{u}_D, -\bar{\omega}$) at 20°N for both GMSs are shown.

a total of 297 rainstorms were identified and traced to their genesis locations: 248 originated over the southwestern China–northern Vietnam region and the northern South China Sea, and the other 49 rainstorm geneses occurred in the BOB–Myanmar region. Our efforts in this study focus only on the former group of rainstorms. Salient geographic features of rainstorm genesis locations in this group included the following:

- 1) A maximum occurrence of genesis appears in the evening over the northern Vietnam–southwestern China region, while a minor maximum exists in the morning over the northern South China Sea.
- 2) Geneses occur in the midtroposphere (700–500-hPa layer).
- 3) All 248 geneses are not directly linked to mei-yu fronts, in contrast with the traditional perception that rainstorms/MCSs evolve from mei-yu fronts.

These rainstorm genesis features are attributed to a special structure for the large-scale environmental flow, particularly in the midtroposphere, favorable for rainstorm genesis in Southeast/East Asia.

During May–June, the development of the Tibetan anticyclone moves the westerly jet northward across the plateau. The confluence of the midtropospheric anticyclonic flow around the eastern plateau and the lower-tropospheric monsoon southwesterlies develops a narrow trough in parallel with the monsoon southwesterlies with its western end located in southwestern China. The Charney–Stern instability criterion is met over this trough in the midtroposphere. When this environmental flow is intensified by the cold surgelike Tibetan northwesterly flow, coupled with a short-wave trough along the eastern seaboard of northeast Asia, rainstorm genesis occurs. For a great majority of these rainstorms, their geneses are triggered by the mechanism independent of East Asian fronts.

FIG. 20. Diurnal mode of the streamfunction budget at 0900 UTC for (a) $(\hat{\psi}_t, \hat{\psi})$ (850 hPa), (b) $(\hat{\psi}_A, \hat{\psi})$ (850 hPa), (c) $(\hat{\psi}_\chi, \hat{\psi})$ (850 hPa). (d)–(f) As in (a)–(c), but 2100 UTC. The contour interval of $(\hat{\psi}_t, \hat{\psi}_A, \hat{\psi}_\chi)$ (850 hPa) is $20 \text{ m}^2 \text{ s}^{-2}$, while the color scale of $\hat{\psi}$ (850 hPa) is shown at the bottom right of (f).

The diurnal variation of the regional atmospheric circulation in Asia and its surrounding oceans in response to the land–sea contrast and north–south differential heating is a *clockwise* rotation of the regional divergent circulation, as observed by [Chen \(2006\)](#). Accompanying the diurnal clockwise rotation of this divergent circulation, a diurnal oscillation of the east–west circulation appears between the land (northern Indochina) and the sea (South China Sea). The diurnal oscillation of upward vertical motions between these two regions results in the diurnal oscillation of the vorticity tendency, primarily through vortex stretching. In turn, this dynamic process spins up the atmospheric flow over land in late evening and over the sea in early morning, and results in the corresponding diurnal variation of rainstorm genesis between land and sea, as observed by this study.

New findings of the monsoon rainstorm genesis activity during May and June in this study offer suggestions for several future research tasks:

- 1) The development of a rainstorm from its genesis is accompanied by a downward penetration of vorticity anomaly. Analyzing the vertical structure of mesoscale convective vortices (MCVs) observed by the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX; [Davis et al. 2004](#)), [Davis and Trier \(2007\)](#) also found the downward development of MCV. Simulating some BAMEX MCVs with the Advanced Research Weather Research and Forecasting Model ([Skamarock et al. 2005](#)), [Davis and Galarneau \(2009\)](#) realized the vertical development of MCV vorticity anomaly includes both the downward penetration and the generation from the planetary boundary layer. The rainstorm and MCV may differ in some basic characteristics, but the BAMEX effort suggests some direction for our mechanism search for the down penetration of rainstorms.
- 2) Revealed from the longitude–time diagrams of rainfall and vorticity, the eastward propagation of the five rainstorms during May and June 2008 are

FIG. 21. Diurnal change of $(\partial q / \partial y, u)$ and θ at (left) 104°E and (right) 114°E : (a),(e) for 0900–1500 UTC; (b),(f) for 2100–0300 UTC; (c) difference between (a) and (b); (g) difference between (e) and (f); (d) $\Delta\theta [= \theta(0900 - 1500 \text{ UTC}) - \theta(2100 - 0300 \text{ UTC})]$ at 104°E ; and (h) as in (d), but at 114°E . The contour interval of all concerned variables in (a)–(c) and (e)–(g) are shown on the top right of (e), while that for (d) and (h) is shown on the top right of (h).

modulated by an intraseasonal mode. Juxtaposed with the South China Sea monsoon trough to the west and the western subtropical Pacific anticyclone to the east, the rainfall produced by these rainstorms should be affected by the modulation of these two synoptic systems by this intraseasonal mode. A future effort should be made to examine the impact of this modulation on the rainfall production of rainstorms over East Asia.

- 3) The geneses of BOB–Myanmar rainstorms occur ahead of the midtropospheric BOB trough. First, this synoptic system is different from that for the southwestern China–northern Vietnam rainstorm. Second,

the distance between the BOB–Myanmar region and East Asia is $\geq 2000 \text{ km}$. For the BOB–Myanmar rainstorms to propagate through such a long distance, it is likely these rainstorms should be propagated eastward by different weather systems across South and Southeast Asia. The genesis and eastward-propagation mechanisms for these rainstorms are likely different from the southwestern China–northern Vietnam rainstorms and need future examination.

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