The effect of regional warming on the East Asian summer monsoon

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Abstract

East Asian summer monsoon (EASM) precipitation has changed significantly due to regional warming. In this study, effect of regional warming on the EASM summer precipitation is investigated during 1979–2018 with a particular emphasis on the frontogenesis. The frontogenetic function shows that change in circulation and temperature distribution associated with regional warming is fairly consistent with precipitation change. Specifically, changes in temperature gradient and deformation, both in terms of its magnitude and direction with respect to thermal gradient, result in cross-frontal circulation change with respect to its position and magnitude; this, in turn, is closely linked with EASM precipitation change. Due to warming, frontal circulation has weakened except in the mature stage, when the front is over the Kuroshio Current. From a slightly southward shifted position, the northward migration of the front is more precipitous than in the past reaching its northernmost latitude earlier by ~2 weeks. On the other hand, the southward migration has become slightly sluggish. In association with the frontogenesis, large-scale precipitation is more strongly affected north of ~30° N. South of ~25° N, precipitation change is due largely to convective precipitation.

Keywords East Asian monsoon · Frontogenesis · Regional warming · Climate change

1 Introduction

East Asia summer monsoon (EASM) is a term that encompasses the overall summer climate system in East Asia. East Asian summer precipitation constitutes an important element of this monsoon system and exhibits significant seasonally repeating characteristics. East Asian summer precipitation is a popular research subject, since it is closely tied with economic issues, disaster prevention, agricultural policy, water resource development plan, and so forth. Furthermore, a large number of people is affected by this climate system because of the high population density in East Asia. Accordingly, variability of the EASM rainfall and its climatic change is a scientific issue of great importance.

This study focuses on analyzing large-scale dynamical effect of regional warming on the frontogenesis and

precipitation associated with the EASM. According to the geographical locations in East Asia, the arrival of the EASM rainband and the subsequent rainfall period is called by various names: ‘Meiyu’ in China, ‘Baiu’ in Japan, and ‘Changma’ in Korea. Effort to explain the dynamical properties of the EASM rainband has a long history. Recent studies include Sampe and Xie (2010), who asserted that the westerly jet plays a role of environmental forcing for the ascending motion in the Meiyu-Baiu rainband. They pointed out that positive horizontal thermal advection by the mid-tropospheric jet induces adiabatic ascending motion to maintain thermal energy balance. As the westerly departs from the eastern flank of the Tibetan Plateau, a wide region of warm advection is developed and upwelling is stimulated over East Asia.

The idea of thermal energy being carried by the westerly jet along the monsoon rainband is expanded into advection of moist static energy by Chen and Bordonì (2014). They also proposed that the existence of Tibetan Plateau is central to the formation and evolution of the Meiyu-Baiu system. Horinouchi (2014) examined the relationship between upper-level disturbance, interpreted as Rossby waves along the jetstream, and synoptic variability of precipitation and moisture transport in summer over East Asia and the

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Northwest Pacific. By conducting potential vorticity (PV)-inversion and Q-vector analysis, he elicited that secondary circulation forces upwelling relevant to precipitation band. Horinouchi and Hayashi (2017) analyzed relationship between upper-tropospheric dynamics and EASM precipitation. They showed that precipitation along the jetstream is induced predominantly by geostrophic flow regardless of the phase of upper-level disturbances.

A number of papers analyzed the relationship between the EASM rainband and summertime large-scale circulation around East Asia. Chen and Chang (1980), Ninomiya (1984), Akiyama (1990), Kodama (1992) are good references to understand fundamental dynamical nature of EASM. Yihui and Chan (2005) overviewed general properties of the EASM including its onset, evolution, contrast with the Indian Monsoon, and its variability on diverse time scales. They also asserted that internal variability of atmospheric circulation is as important for EASM variability as other well-known external forcing such as Eurasian snow cover and Pacific Ocean SST. In view of these earlier studies, atmospheric circulation change due to regional warming may significantly affect EASM rainfall.

Many of the earlier papers that explore the impact of global warming on monsoon precipitation used atmospheric modeling results (e.g. Kimoto 2005; Kitoh and Uchiyama 2006; Arai and Kimoto 2007; Kawase et al. 2009; Li et al. 2010; Zhu et al. 2012; Seo et al. 2013; Chen and Sun 2013; Hong and Ahn 2015; Zhou et al. 2017; Liu et al. 2018; Horinouchi et al. 2019; Jin and Stan 2019). It is difficult to summarize the past work in a short paragraph, since each study is conducted in a unique setting in terms of model, domain and period to address distinct aspects of climate impact on EASM precipitation. Nevertheless, a common result is that the amount of EASM precipitation has increased and the position of the monsoon rainband has shifted southward due to global warming. Through modeling efforts, one can also compare the evolution pattern and intensity of EASM precipitation among different global warming scenarios, and address future change in EASM precipitation.

In this study, however, no modeling result is used. A statistical analysis tool called the cyclostationary empirical orthogonal function (CSEOF) technique (Kim et al. 1996, Kim et al. 2015; Kim and North 1997) is used to extract the seasonal cycle of the EASM and the global warming signal from a reanalysis dataset. Based on the seasonal cycle and the warming signal, we will address how regional warming affects the seasonal cycle of the EASM. Using this decomposition, we will investigate how the precipitation pattern associated with the EASM has been changing with the changing climate and what we should expect in future if this trend continues. We will, in particular, focus on the impact of regional warming on the dynamical characteristics

of the monsoonal front and associated precipitation during the past 40 years.

Information on the dataset used in this study is presented in Sect. 2. The primary tool for analysis in this study is the CSEOF technique. Section 3 discusses the basic features of CSEOF analysis and the regression analysis in CSEOF space. Results of analysis are presented in Sect. 4. The evolution pattern of East Asia summer precipitation as well as its frontal structure and cross-frontal circulation are described and contrasted between the pre-warming period and the post-warming period. Change in the frontal structure, a key element of monsoonal precipitation, is investigated at four critical stages of the EASM in this section. Section 5 summarizes the change in precipitation and frontal structure of the East Asian summer monsoon, followed by concluding remarks in Sect. 6.

2 Data

Data used in this study are the European Center for Medium-Range Weather Forecast (ECMWF)-interim reanalysis (ERA) datasets (Dee et al. 2011) for 40 years (1979–2018). We used the records in summer from May 15 to September 11 at a 5-day resolution covering the domain 90° – 180° E and 0° – 60° N at a $1.5^{\circ} \times 1.5^{\circ}$ resolution. We analyzed the data at 23 vertical levels (1000–200 hPa). The variables used in this study include 2 m (surface) air temperature, precipitation, large-scale precipitation, and convective precipitation and pressure variables such as temperature, geopotential, zonal and meridional wind, and vertical (pressure) velocity.

3 Method of analysis

CSEOF analysis is conducted on all variables to extract the seasonal cycle and the global warming signal. In CSEOF analysis, space–time data are decomposed as (Kim et al. 1996, 2015; Kim and North 1997).

$$T(r, t) = \sum_n B_n(r, t) T_n(t), \quad (1)$$

where $B_n(r, t)$ are cyclostationary loading vectors (CSLVs) and $T_n(t)$ corresponding principle component (PC) time series. CSLVs are periodic in time,

$$B_n(r, t) = B_n(r, t + d), \quad (2)$$

where the period d is called the nested period and represents the period of the statistics of the underlying data. As in EOF analysis, CSLVs are mutually orthogonal and PCs are mutually uncorrelated, i.e.,

Fig. 1 The seasonal cycle of precipitation (shading; mm day^{-1}), lower-tropospheric (1000–850 hPa) geopotential height (contoured at 3 m interval) and wind (vectors) over East Asia during summer (May 15–September 11): **a** May 15–Jun 13, **b** Jun 14–Jul 13, **c** Jul 14–Aug 12, and **d** Aug 13–Sep 11. All the patterns are anomalies with respect to the summer mean over the 40 years. The positive contours are in red and the negative in blue. Vectors with the wind speed greater than 2 m s^{-1} are drawn. The red dashed line represents the cross-section used in Figs. 3, 4, 5, 8 and 9. **e** The corresponding PC time series

$$\frac{1}{Nd} \sum_{r=1}^N \sum_{t=1}^d B_n(r, t) B_n(r, t) = \delta_{nn}, \quad (3)$$

$$\frac{1}{M} \sum_{t=1}^M T_n(t) T_m(t) = \lambda_n \delta_{nm}, \quad (4)$$

where N is the number of spatial points, M is the number of temporal points, δ_{nm} is the Kronecker delta, and λ_n , called eigenvalue, represents the variance of $T_n(t)$. Thus, CSEOF analysis decomposes data into mutually orthogonal patterns of evolution, of which the PC time series are mutually uncorrelated.

In order to make two or more variables to be physically consistent with each other, regression analysis is

conducted in CSEOF space (Kim 2017). Let us consider another variable, $P(r, t)$, which can be decomposed as

$$P(r, t) = \sum_n C_n(r, t) P_n(t). \quad (5)$$

Then, the PC time series of the so-called “target variable”, $T_n(t)$, is written as a regression equation in terms of the PC time series of the “predictor variable”, $P_m(t)$:

$$T_n(t) = \sum_{m=1}^M \alpha_m^{(n)} P_m(t) + \epsilon^{(n)}(t), \quad n = 1, 2, \dots, \quad (6)$$

where M is the number of predictor time series used for multivariate regression, $\alpha_m^{(n)}$ are regression coefficients, and $\epsilon^{(n)}(t)$ is regression error time series. Then, the so-called “regressed” CSLVs are determined as

Fig. 2 The patterns of precipitation (shading; mm day⁻¹), lower-tropospheric (1000–850 hPa) geopotential height (contoured at 3 m interval) and wind (vectors) over East Asia during summer (May 15–September 11) associated with regional warming: **a** May 15–Jun 13, **b** Jun 14–Jul 13, **c** Jul 14–Aug 12, and **d** Aug 13–Sep 11. All the patterns are anomalies. The positive contours are in red and the negative in blue. Vectors with the wind speed greater than 0.3 m s⁻¹ are drawn. **e** The corresponding PC time series (red curve) and its linear trend (blue dashed line), with 3 warmest years (red dots) and 3 coolest years (blue dots)

$$C_n^{(\text{reg})}(r, t) = \sum_{m=1}^M \alpha_m^{(n)} C_m(r, t). \quad (7)$$

Then, the predictor variable can be rewritten as

$$P(r, t) = \sum_m C_m(r, t) P_m(t) = \sum_n C_n^{(\text{reg})}(r, t) T_n(t). \quad (8)$$

Thus, the target and the predictor variables are written together as

$$\{T(r, t), P(r, t)\} = \sum_n \{B_n(r, t), C_n^{(\text{reg})}(r, t)\} T_n(t). \quad (9)$$

Thus, the CSLVs of the target variable, $B_n(r, t)$, and those of the predictor variable, $C_n^{(\text{reg})}(r, t)$, are governed by the same

PC time series $T_n(t)$, and are considered to be physically consistent with each other.

This procedure can be repeated for other predictor variables. As a result of regression analysis in CSEOF space, entire variables can be written as

$$Data(r, t) = \sum_n \{B(r, t), C_n^{(\text{reg})}(r, t), D_n^{(\text{reg})}(r, t), \dots\} T_n(t), \quad (10)$$

then, the terms in curly braces are considered physically consistent and depict multi-faceted evolution patterns of a physical process for each n . Surface air temperature is used as the target variable for extracting the warming mode. Then, the first mode ($n = 1$) depicts the seasonal cycle of summer

Fig. 3 The evolution of the seasonal cycle of precipitation (shading in units of mm day^{-1}) together with **a, b** vertical velocity (red contours in units of $10^{-4} \text{ Pa s}^{-1}$), and **c, d** convergence of 500-hPa Q-vectors (red contour at the interval of $5 \times 10^{-13} \text{ m}^{-2} \text{ s}^{-1}$) and lower-tropospheric (1000–500 hPa) wind (blue contours at the interval of

$2 \times 10^{-7} \text{ s}^{-1}$) along the cross section (see Fig. 1) joining (139.5° E , 10.5° N) and (109.5° E , 55.5° N). All the patterns are anomalies. The left panels represent the pre-warming period and the right panels the post-warming period

precipitation, and the second mode ($n = 2$) represents the evolution associated with warming over the study domain.

4 Results

In this study, data from May 15 to September 11 (120 days) each year from 1979 to 2018 (40 years) are used to investigate summertime variability in East Asia. Therefore, $d = 24$ pentads (120 days), and the total record length is 960 points (24 pentads $\times$ 40 years). By conducting regression analysis for the first two CSEOF modes, the entire evolution patterns associated with the seasonal cycle and the warming mode are identified.

Figure 1 shows the seasonal cycle of precipitation over the East Asian domain at the four characteristic stages of evolution: (a) initiation, (b) maturity, (c) transition, and (d) termination (Lim et al. 2002). In the initiation stage, precipitation begins to increase in the low-latitude regions including the Indochina Peninsula and Southeast China. In the mature stage, active moisture transport is seen along the eastern coast of the Asian continent, marking the onset of long spell of precipitation in East Asian countries including China, Japan, and Korea. The transition period exhibits the southward migration of the jet axis and weakening of the westerly jet along the rainband. During this period, monsoonal circulation begins to be reversed due to the establishment of low pressure anomaly over the ocean. In the termination stage, monsoon circulation is completely reversed

Fig. 4 Anomalous patterns of (a, b) precipitation (shading; mm day⁻¹) and 500-hPa vertical velocity ($-\omega$, contour; 10^{-4} Pa s⁻¹), (c, d) precipitation (shading; left color bar), 500-hPa Q-vectors (vectors; 10^{-7} m⁻¹ s⁻¹), Q-vector convergence (contours; 10^{-12} m⁻² s⁻¹), and vertical velocity (color of vectors; 10^{-4} Pa s⁻¹; right color bar), (e, f) 500-hPa $\frac{1}{2}|\nabla_h T|D \cos(2\eta)$ (shading; 10^{-12} K m⁻¹ s⁻¹; left color bar), $D(\sin 2\eta, -\cos 2\eta)$ (vector; 10^{-6} s⁻¹), $|\nabla_h T|$ (color of vector; 10^{-6} K m⁻¹; right color bar), and 200-hPa wind speed (contours; m s⁻¹), (g, h) ageostrophic circulation (streamlines) with the magnitude of $\frac{1}{2}|\nabla_h T|D \cos(2\eta)$ in color, wind convergence (contours; 10^{-6} s⁻¹), and precipitation (red curve; mm day⁻¹) along the cross section (see dashed lines in a, b, e, and f) joining (139.5° E, 10.5° N) and (109.5° E, 55.5° N) in the initiation stage of the East Asian summer monsoon (May 15–19). Positive contours in a–d are in red and negative in blue. The left column represents the pre-warming period and the right column the post-warming period

with a southward moisture transport anomaly along the eastern side of the continent.

The corresponding PC time series shows that the amplitude of the seasonal cycle of precipitation varies by about $\pm 50\%$ during 1979–2018. While convective precipitation explains a larger fraction of total precipitation over the domain (see supplementary Figs. S2a and S2b), large-scale precipitation is a major source of variability for the monsoon precipitation band (see Figs. S2c and S2d).

Figure 2 shows the warming mode and the corresponding PC time series. As can be seen in the figure, significant change is seen in terms of low-level tropospheric circulation and precipitation due to warming. The linear fit to the PC time series indicates that the magnitude of warming has increased by ~ 2.3 during the past 40 years; while the PC time series shows both anthropogenic and natural causes of regional warming (Lean and Rind 2008; Liu et al. 2013; Zhang 2015), the linear trend may primarily reflect the man-made change. Multiplying this magnitude change with the CSEOF loading vector, we see that the effect of warming on summer precipitation in East Asia amounts to a significant fraction of the seasonal variation of precipitation.

In the following, evolution of variables during the monsoon period will be contrasted between the pre-warming and post-warming background conditions. The pre-warming and post-warming periods should be understood as the period of weak regional warming and that of strong regional warming, respectively. For this, seasonal evolution of precipitation in the pre-warming (post-warming) condition is defined by adding the regional warming mode (Fig. 2a–d) multiplied by the 1979 (2018) value of the linear trend in Fig. 2e. Thus, the regional warming mode represents the background change in which the seasonal cycle of precipitation evolves. Other variables in the pre-warming and post-warming conditions are defined in the same way. Other factors that can affect the East Asian monsoon are not taken into consideration in order to focus solely on the impact of regional warming.

The formation and evolution of monsoon rainband is an important scientific issue, since it essentially dictates the amount of rain at a specific region at a specific time of the year. Figure 3 shows precipitation along the path of monsoon evolution from (139.5° E, 10.5° N) to (109.5° E, 55.5° N) (red dashed line in Fig. 1). Figure 3 demonstrates that monsoon climatology has changed significantly during the past 40 years due to regional warming. Much heavier rainfall is seen in June at $\sim 25^\circ$ – 30° N (see Fig. 3a, b). On the other hand, rainfall has decreased significantly at $\sim 30^\circ$ – 35° N followed by a substantial precipitation reduction starting from mid-July. The Changma front affecting the Korean peninsula is not established firmly in early July. Further, the secondary (late) Changma has significantly dwindled in late August (see Fig. 3a, b).

The evolution of precipitation matches well with the 500-hPa vertical (negative omega) velocity, which in turn is explained by the lower-tropospheric convergence of horizontal wind and 500-hPa Q-vectors. According to the continuity equation (mass conservation principle) in pressure coordinates, horizontal convergence of wind is related with the variation of vertical velocity with pressure:

$$-\nabla_h \cdot \vec{v} = \frac{\partial \omega}{\partial p}. \quad (11)$$

Thus, positive wind convergence in the lower troposphere leads to an upward motion.

The Q-vectors are defined by

$$\vec{Q} = -\frac{R}{p} \left(\frac{\partial \vec{v}_g}{\partial x} \cdot \nabla T, \frac{\partial \vec{v}_g}{\partial y} \cdot \nabla T \right), \quad (12)$$

and depicts essentially variation of thermal advection ($-\vec{v}_g \cdot \nabla T$) due to the variation of geostrophic wind $\vec{v}_g$ in each direction. According to the quasi-geostrophic theory (Hoskins et al. 1978; Holton 2004), convergence of $\vec{Q}$ leads to an upward velocity. As can be seen in Fig. 3b, convergence of $\vec{Q}$ and that of lower-tropospheric wind explain reasonably the evolution of vertical velocity and precipitation.

In order to elaborate on the possible mechanisms of precipitation change due to regional warming, several thermodynamic quantities are compared between the pre-warming and the post-warming periods. Figure 4 shows the pattern of precipitation as well as those of other physical variables on May 15–May 19. At this early stage of the EASM, precipitation has generally increased over the continent and the western Pacific and decreased over southeast China and the southern Japan due to warming (see Fig. 4a, b). The patterns of vertical velocity are consistent with those of precipitation.

This is also reflected in the pattern of Q-vector convergence (Fig. 4c, d); slightly increased convergence and decreased divergence is seen over the western Pacific and

Fig. 5 Same as Fig. 4 except in the maturity stage of the East Asian monsoon (June 14–18)

over the continent north of $\sim 30^\circ$ N, respectively. While the Q-vector convergence is significant over the southern part of Japan in the post-warming period, downward motion is still dominant in this area (Fig. 4b, d).

From the thermodynamic energy equation, we obtain based on the quasi-geostrophic assumption (see “Appendix” for a full derivation)

$$\frac{D_s}{Dt} \left(\frac{R}{p} |\nabla_h T| \right) = \left[\left(\frac{R}{c_p p} \right) \nabla_h J_0 \cdot \frac{\nabla_h T}{|\nabla_h T|} + \frac{R}{2p} |\nabla_h T| D \cos(2\eta) \right]. \quad (13)$$

The left-hand side of (13) denotes the total change in horizontal temperature gradient (without sign) following the geostrophic motion. The right-hand side of (13) represents the contribution of diabatic/adiabatic heating J_0 (first term) and deformation D in the production of horizontal temperature gradient. Figure 4e and f show the effect of deformation in terms of the magnitude of D (length of vectors), angle 2η as a vector ($\sin 2\eta, -\cos 2\eta$), and the magnitude of horizontal temperature gradient $|\nabla_h T|$ (color of vectors) at 500 hPa in comparison with the 200-hPa wind speed (black contours). Thus, if $2\eta = 0$ ($2\eta = \pi$), the direction of vector is southward (northward), whereas vector points eastward (westward) for $2\eta = \pi/2$ ($2\eta = -\pi/2$). We will call $\frac{1}{2} |\nabla_h T| D \cos(2\eta)$ the frontogenetic function. The sign of the frontogenetic function is solely determined by $\cos(2\eta)$, which is expressed in terms of background shading in Fig. 4e, f.

At the initiation stage of the EASM, the direction of deformation is generally parallel to temperature contours ($2\eta \rightarrow 0$; southward arrows) in the northern part of the domain. On the other hand, the direction of deformation tends to be perpendicular to temperature contours in the southern part ($2\eta \rightarrow \pi$; northward arrows). This implies that isotherms become denser, and as a result, geostrophic wind speed increases in the northern part of the domain. In the southern part, the situation is opposite. This demarcation is a sign of frontogenesis, and develops slightly to the south of the 200-hPa jet speed maximum in agreement with earlier studies (Horinouchi and Hayashi 2017; Horinouchi et al. 2019).

According to the theory of frontogenesis (Hoskins 1982; Holton 2004), the increased geostrophic wind speed due to the increased thermal gradient is accompanied by cross-frontal ageostrophic wind in the direction perpendicular to the geostrophic wind. According to the mass conservation principle, this ageostrophic wind produces vertical motion, leading to a secondary cross-frontal circulation. The vertical section along the line connecting (139.5° E, 10.5° N) and (109.5° E, 55.5° N) clearly shows ageostrophic meridional circulation cell with a rising motion to the south and a sinking motion to the north

of the front (Fig. 4g, h). In the region with an upward motion, precipitation begins to increase, and precipitation deficit (with respect to the summer mean) is seen in the region of a downward motion.

Figure 5 shows precipitation in the mature stage of the EASM (June 14–18). A well-developed monsoon band and the convergence of Q vectors are clearly seen along the Kuroshio Current (Fig. 5c, d). The width of the monsoon precipitation band has increased and the region of Q-vector convergence has widened due to warming. Further, the arrival of the monsoon band in Japan has been slightly delayed (see e.g., Kitoh and Uchiyama 2006).

A comparison between Fig. 5e, f indicates that front is more strongly developed to the south of Korea and Japan under a warming condition; the frontogenetic function exhibits a higher value along $\sim 30^\circ$ N under a warming condition (Fig. 5e, f). This implies that cross-frontal circulation has become stronger mainly due to increased deformation D to the south of the maximum jet. As can be seen in Fig. 5g and h, monsoon front is more strongly developed (denser streamlines and darker colors), and its position is shifted by $\sim 2^\circ$ southward (see e.g., Kawase et al. 2009; Horinouchi et al. 2019). The upward branch of the ageostrophic circulation is wider, leading to a wider band of precipitation.

Figures 6 and 7 show the patterns of several key variables when the monsoon front is over the Kuroshio Current. Both surface convergence and vertical velocity are enhanced over the warm Kuroshio Current (Figs. 6a–c, 7a–c). Convergence of horizontal wind and vertical velocity are seen to increase remarkably in the troposphere due to warming (Figs. 6d, 7d). This is also reflected in the sharper ascension of the streamlines of the eastward motion along the Kuroshio Current (Figs. 6e, 7e).

It should be pointed out that precipitation is significantly enhanced only when the monsoon front is over the Kuroshio Current. This indicates that ocean-atmospheric interaction is a crucial element for the intensification of precipitation. It appears that the warming of the Kuroshio Current has increased convergence of horizontal wind and strengthened vertical velocity. This implies that the adiabatic/diabatic forcing term J_0 in the frontogenetic function (13) plays an important role in the strengthening of the monsoon front when the front is over the Kuroshio Current. As a result, precipitation is enhanced under the warming condition. As the monsoon front migrates further north, vertical velocity and surface wind convergence diminishes quickly.

Figure 8 shows precipitation in the transition phase of the EASM (Fig. 8a, b). Precipitation is dominant over the northern land areas, and significant precipitation deficit (with respect to the summer mean) is seen over the East China Sea.

Fig. 6 The patterns of **a–c** sea surface temperature (contour; °C), **a** total precipitation (shade; mm day^{-1}), **b** 1000-hPa wind convergence (shade; 10^{-6} s^{-1}), **c** 1000-hPa vertical velocity ($-\omega$; shade; $10^{-3} \text{ Pa s}^{-1}$), **d** vertical cross section along (110° E, 21° N) and (159.5° E,

35° N) (red dashed line in panel a) of vertical velocity (shade; $10^{-2} \text{ Pa s}^{-1}$) and wind convergence (contour; 10^{-6} s^{-1}), and **e** streamlines of zonal (m s^{-1}) and vertical velocity ($10^{-2} \text{ Pa s}^{-1}$) with the meridional component overlaid as color (m s^{-1}) on Jun 9–13

Fig. 7 Same as Fig. 6 except on Jun 14–18

The convergence pattern of Q vectors is marginally consistent with that of vertical velocity (Fig. 8a–d), suggesting that the vertical motion is induced by thermal advection as well as other mechanisms (such as convection from heat low).

In fact, a larger fraction of precipitation shown in Fig. 8a and b represents convective precipitation, the mechanism of which is only loosely connected with the frontogenesis (figure not shown).

Fig. 8 Same as Fig. 4 except in the transition stage of the East Asian monsoon (July 14–18)

In the pre-warming period, a narrow band of negative frontogenetic function runs northeastward in the middle of positive values to the west of $\sim 140^\circ$ E (Fig. 8e), which results in two cells of secondary circulation (Fig. 8g). A broad region of upward motion and precipitation surplus is seen north of $\sim 32^\circ$ N in the vertical section (Fig. 8g).

The frontal circulation has weakened significantly and its position has shifted northward by $\sim 5^\circ$ as a result of warming (Fig. 8f). This weakening is mainly due to weak temperature gradient both over the continent and the ocean. Also, the direction of deformation is not favorable for producing substantial temperature gradient. Due to the weak frontal circulation (Fig. 8h), precipitation has dwindled significantly to the north of the front (Fig. 8b).

Figure 9 shows the pattern of precipitation during the termination stage of the EASM (Fig. 9a, b). In the pre-warming period, precipitation is strongly developed over the Korean peninsula; this secondary maximum of precipitation is called the second (or late) Changma in Korea. While strong convergence of Q vectors is seen to the west of the Korean peninsula, upward velocity is not so strong in the post-warming period (Fig. 9b, d). While the frontogenetic function is fairly strong to the west of the Korean peninsula (Fig. 9f), ageostrophic meridional circulation is disorganized and is limited to the upper troposphere (Fig. 9h). As can be seen, the location of front is shifted slightly northward in the post-warming period.

5 Summary and discussion

In the present study, we investigated the physical and dynamical mechanisms of monsoon precipitation change in East Asia associated with regional warming. In particular, we focused on the effect of warming in terms of the frontogenetic nature of the East Asian summer monsoon. It should first be noted that the precipitation patterns in Figs. 4, 5, 8 and 9 agree reasonably with the composite patterns of precipitation based on three coolest summers and three warmest summers in the record (see Fig. S4).

Figure 10 shows the evolution patterns of precipitation along the three meridional sections in the pre-warming and post-warming periods. Along 125° E, precipitation has generally decreased throughout the summer due to warming except when the precipitation band is over the Kuroshio Current (Fig. 10a). To the north of $\sim 38^\circ$ N, both large-scale precipitation and convective precipitation contributes to this reduction. Between $\sim 26^\circ$ and 38° N, weakening of large-scale precipitation is primarily responsible for the reduction of precipitation. To the south of $\sim 26^\circ$ N, on the other hand, convective precipitation is largely responsible for the reduced precipitation. The increased precipitation over the

Kuroshio Current is caused mainly by the enhancement of large-scale precipitation.

Along 135° E, precipitation is enhanced over the Kuroshio Current in early June (Fig. 10b). Otherwise, precipitation is generally reduced until near the end of summer. Again, large-scale precipitation is responsible for the precipitation reduction north of $\sim 38^\circ$ N, whereas convective precipitation is generally responsible for the reduced precipitation south of $\sim 28^\circ$ E.

Along 145° E, increased precipitation is seen between 30° and 40° N until the middle of June and then decreased precipitation until the rest of the summer (Fig. 10c). This pattern is induced largely by the change in large-scale precipitation. In the lower latitude, on the other hand, precipitation has increased since July 16, and this change is primarily due to convective precipitation.

A notable change is seen in the frontogenesis of the East Asian monsoon as a result of warming. As summarized in Fig. 11, 200-hPa wind speed maximum migrates northward during the initiation and mature stages, and then southward during the transition and termination periods. The northward migration of the 200-hPa jet speed maximum has become faster and the southward migration starts ~ 2 weeks earlier under the warming condition, particularly in the eastern part of the domain. This change in the meridional migration speed is due to the variation of 200-hPa geopotential as a result of regional warming (see Fig. S8 and the discussion therein). The monsoon front, which is developed slightly to the south of the maximum wind speed, also migrates with the maximum wind speed. This implies that the transition period develops ~ 2 weeks earlier under the warming condition.

The 200-hPa wind speed has increased in the initiation and the early mature stages, but slightly weakens since the late mature stage of the EASM. The increased wind speed is due to stronger thermal gradient in the initiation stage in particular, and is favorable for frontogenesis. The cross-frontal circulation, however, is not strongly established until the monsoon band reaches the Kuroshio Current. Under the warming condition, stronger wind convergence and vertical velocity are observed in the lower troposphere over the Kuroshio Current, resulting in stronger cross-frontal circulation and increased precipitation (primarily large-scale precipitation).

When the maximum wind is at its highest latitude ($\sim 35^\circ$ – 40° N during the latter part of the mature stage and the early part of the transition stage), cross-frontal circulation is fairly weak in the post-warming period; while deformation is not weak, temperature gradient is significantly reduced at the location of the front. As a result, precipitation is significantly reduced. It should also be noted that the front is shifted slightly northward upon regional warming.

Fig. 9 Same as Fig. 4 except in the termination stage of the East Asian monsoon (August 23–27)

Fig. 10 Evolution of total precipitation (contours; mm day^{-1}), anomalous precipitation from the summer mean (shade; mm day^{-1}) and the contributions of large-scale precipitation and convective precipitation to precipitation anomalies (vectors) throughout the summer (May 15–September 11): **a** 125°E , **b** 135°E , and **c** 145°E . The blue contours represent precipitation $\leq 5 \text{ mm day}^{-1}$, and the red contours $> 5 \text{ mm day}^{-1}$. The contribution of large-scale precipitation and convective precipitation is measured in terms the angle of the vector,

with the zero angle measured from the x axis denotes equal contribution, positive angle denotes greater contribution by large-scale precipitation (45° contribution by large-scale precipitation only), and negative angle greater contribution by convective precipitation (-45° by convective precipitation only). The length of the vectors is proportional to the magnitude of the anomalous precipitation. The red triangles denote the four stages depicted in Figs. 4, 5, 8 and 9

As the EASM comes close to its termination, the front and jet move back southward, and the magnitude of deformation and temperature gradient increase again. Nevertheless, the amount of precipitation does not increase

compared to the maturity stage of the monsoon. This difference is attributed to the reduction of moisture supply by the low-level flow from the low-latitude region.

Fig. 11 Evolution of 500-hPa frontogenetic function ($10^{-12} \text{ K m}^{-1} \text{ s}^{-1}$; shading; left color bar) together with the magnitude (length of vectors) and direction of deformation with respect to the direction of thermal gradient (2η ; angle measured counterclockwise from south),

the magnitude of thermal gradient (10^{-6} s^{-1} ; color of vectors; right color bar), and 200-hPa wind speed (contours) throughout the summer (May 15–September 11): **a** 125° E, **b** 135° E, and **c** 145° E. The red triangles denote the four stages depicted in Figs. 4, 5, 8 and 9

6 Concluding remarks

As demonstrated in this study, dynamic frontogenesis theory applies reasonably to summer precipitation in East Asia. Since the thermal capacity of land is much lower than that of the ocean, surface air temperature of

mid-latitude land is higher than that of mid-latitude ocean while the difference is small in low-latitude. The difference in meridional temperature gradient between land and ocean causes zonally asymmetric jetstream. In summer, therefore, speed of jetstream increases as it passes the eastern shore of East Asia. This acceleration is accompanied

by considerable deformation, which causes frontogenesis and frontolysis around the jet, thus generating secondary circulation across the frontal structure. Monsoon rainband is formed between the northern frontogenetic region and southern frontolytic region of the troposphere.

In conclusion, the East Asian monsoonal precipitation change due to regional warming is closely linked with the change in the location and magnitude of frontogenesis and the resulting cross-frontal ageostrophic circulation. In particular, change in the latitudinal location of the monsoon rainband is captured faithfully in terms of the frontogenetic function. Due to regional warming, the monsoon front has shifted slightly southward in early summer (maturity stage), but slightly northward in mid summer (transition stage). Thus, the seasonal migration of the monsoon front is more precipitous than in the past. Further, the cross-frontal circulation has generally weakened except in the maturity stage. These, in general, result in decreased amount of large-scale precipitation along the front. This trend is expected to continue as warming deepens in East Asia. While the frontal theory is reasonable in explaining the EASM precipitation pattern and its change due to global warming, such factors as rainband-warm current interaction and strong convection over land in late summer also contribute to the variation of precipitation.

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Appendix

The thermodynamic energy equation is given by

$$\frac{D_g T}{Dt} = \frac{J}{c_p} + S_p \omega = \frac{J_0}{c_p}, \quad S_p = -T \frac{\partial \ln \theta}{\partial p}. \quad (14)$$

Then, we have

$$\frac{D_g}{Dt} \left(\frac{\partial T}{\partial x} \right) = \frac{\partial}{\partial x} \left(\frac{D_g T}{Dt} \right) - \frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} - \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y}. \quad (15)$$

Multiplying (15) by $\partial T / \partial x$, we obtain

$$\frac{D_g}{Dt} \left(\frac{\partial T}{\partial x} \right)^2 = 2 \left[\frac{1}{c_p} \frac{\partial J_0}{\partial x} \frac{\partial T}{\partial x} - \left(\frac{\partial T}{\partial x} \right) \left(\frac{\partial u_g}{\partial x} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial x} \frac{\partial T}{\partial y} \right) \right]. \quad (16)$$

Likewise,

$$\frac{D_g}{Dt} \left(\frac{\partial T}{\partial y} \right)^2 = 2 \left[\frac{1}{c_p} \frac{\partial J_0}{\partial y} \frac{\partial T}{\partial y} - \left(\frac{\partial T}{\partial y} \right) \left(\frac{\partial u_g}{\partial y} \frac{\partial T}{\partial x} + \frac{\partial v_g}{\partial y} \frac{\partial T}{\partial y} \right) \right]. \quad (17)$$

Therefore, we can write

$$\begin{aligned} \frac{D_g}{Dt} |\nabla_h T|^2 &= 2 \left[\frac{1}{c_p} \nabla_h J_0 \cdot \nabla_h T - \nabla_h T \cdot (\nabla_h \vec{v}_g \cdot \nabla_h T) \right] \\ &= - \left[(T_x^2 + T_y^2) \left(\frac{\partial u_g}{\partial x} + \frac{\partial v_g}{\partial y} \right) + (T_x^2 - T_y^2) \left(\frac{\partial u_g}{\partial x} - \frac{\partial v_g}{\partial y} \right) \right. \\ &\quad \left. + 2T_x T_y \left(\frac{\partial v_g}{\partial x} + \frac{\partial u_g}{\partial y} \right) \right] + 2\nabla_h \left(\frac{J_0}{c_p} \right) \cdot \nabla_h T. \end{aligned} \quad (18)$$

Thus,

$$\frac{D_g}{Dt} |\nabla_h T| = \frac{1}{|\nabla_h T|} \left[\nabla_h \left(\frac{J_0}{c_p} \right) \cdot \nabla_h T - \frac{A}{2} \right], \quad (19)$$

where the term A is given by

$$\begin{aligned} A &= (T_x^2 + T_y^2) \left(\frac{\partial u_g}{\partial x} + \frac{\partial v_g}{\partial y} \right) \\ &\quad + (T_x^2 - T_y^2) \left(\frac{\partial u_g}{\partial x} - \frac{\partial v_g}{\partial y} \right) + 2T_x T_y \left(\frac{\partial v_g}{\partial x} + \frac{\partial u_g}{\partial y} \right). \end{aligned} \quad (20)$$

By using the stretching deformation and shear deformation defined by

$$D_{st} = \left(\frac{\partial u_g}{\partial x} - \frac{\partial v_g}{\partial y} \right), \quad D_{sh} = \left(\frac{\partial v_g}{\partial x} + \frac{\partial u_g}{\partial y} \right), \quad (21)$$

we can rewrite (20) as

$$A = \left[(T_x^2 - T_y^2) D_{st} + 2T_x T_y D_{sh} \right], \quad (22)$$

where divergence of geostrophic velocity is zero. Using matrix operations, (22) can be expressed as

$$\begin{aligned} A &= \begin{bmatrix} T_x & T_y \end{bmatrix} \begin{bmatrix} D_{st} & D_{sh} \\ D_{sh} & -D_{st} \end{bmatrix} \begin{bmatrix} T_x \\ T_y \end{bmatrix} \\ &= \begin{bmatrix} T_x & T_y \end{bmatrix} \begin{bmatrix} \cos \delta & \sin \delta \\ -\sin \delta & \cos \delta \end{bmatrix}^T \begin{bmatrix} D & 0 \\ 0 & -D \end{bmatrix} \begin{bmatrix} \cos \delta & \sin \delta \\ -\sin \delta & \cos \delta \end{bmatrix} \begin{bmatrix} T_x \\ T_y \end{bmatrix}, \end{aligned} \quad (23)$$

where

$$D = (D_{st}^2 + D_{sh}^2)^{1/2}, \quad (24)$$

and the angle of dilation (principal) axis with respect to the horizontal axis (x) is given by

$$\delta = \tan^{-1} (D_{sh} / (D + D_{st})). \quad (25)$$

It can be shown that

$$\begin{aligned} \tan 2\delta &= \frac{\sin 2\delta}{\cos 2\delta} = \frac{2 \sin \delta \cos \delta}{\cos^2 \delta - \sin^2 \delta} = \frac{2 \tan \delta}{1 - \tan^2 \delta} \\ &= \frac{2 \frac{D_{sh}}{D+D_{st}}}{1 - \left(\frac{D_{sh}}{D+D_{st}}\right)^2} = \frac{2D_{sh}(D+D_{st})}{(D+D_{st})^2 - D_{sh}^2} = \frac{2D_{sh}(D+D_{st})}{2D_{st}(D+D_{st})} = \frac{D_{sh}}{D_{st}}. \end{aligned} \quad (26)$$

Thus, $\tan 2\delta$ is the ratio of the shear deformation to stretching deformation. Note that $\delta \rightarrow 0$ if $D_{st} \gg D_{sh}$, and $\delta \rightarrow \pi/4$ if $D_{sh} \gg D_{st}$.

We can simplify (23) as

$$\begin{aligned} A &= D \left[(T_x \cos \delta + T_y \sin \delta)^2 - (T_x \sin \delta - T_y \cos \delta)^2 \right] \\ &= D \left[(T_y^2 - T_x^2) (\sin^2 \delta - \cos^2 \delta) + 4T_x T_y \sin \delta \cos \delta \right] \\ &= -D |\nabla_h T|^2 [\cos 2\theta \cos 2\delta + \sin 2\theta \sin 2\delta] \\ &= -D |\nabla_h T|^2 \cos (2\theta - 2\delta) = -D |\nabla_h T|^2 \cos (2\eta), \end{aligned} \quad (27)$$

where

$$\eta = \theta - \delta, \quad \text{and} \quad \theta = \tan^{-1} (-T_x / T_y). \quad (28)$$

Using (27), we finally obtain

$$\frac{D_g}{Dt} |\nabla_h T| = \left[\nabla_h \left(\frac{J_0}{c_p} \right) \cdot \frac{\nabla_h T}{|\nabla_h T|} + \frac{1}{2} |\nabla_h T| D \cos (2\eta) \right]. \quad (29)$$

As can be seen in (29), the magnitude of the geostrophic advection of thermal gradient depends on the gradient of $J_0 (= J + S_p \omega)$ in the direction of $\nabla_h T$, and on the deformation term D multiplied by $\cos(2\eta)$. If we multiply (29) by R/p , then we obtain a more familiar form

$$\frac{D_g}{Dt} \left(\frac{R}{p} |\nabla_h T| \right) = \left[\left(\frac{R}{c_p p} \right) \nabla_h J_0 \cdot \frac{\nabla_h T}{|\nabla_h T|} + \frac{R}{2p} |\nabla_h T| D \cos (2\eta) \right], \quad (30)$$

where the second term in the square brackets are identified as the norm of Q vector projected on $\nabla_h T$. The right-hand side of (30), then, represents the source/sink of thermal gradient. (29) indicates that increase in $|\nabla_h T|$ following the geostrophic motion is maximized if $\eta = 0$, i.e., when the deformation angle δ is identical with the angle θ . For more information about frontogenesis, see Keyser et al. (1988) and Cohen and Schultz (2005).

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