Role of the Surface Boundary Conditions in Boreal Spring on the Interannual Variability of the Multistage Evolution of the East Asian Summer Monsoon

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ABSTRACT

The seasonal movement of the upper-level jet plays a key role in the evolution of the East Asian summer monsoon (EASM). However, it remains unresolved how interannual changes in surface boundary conditions can influence the upper-level flow over East Asia, thereby modulating the onset of the EASM. Here we capture the timing of multistage evolution over East Asia using the upper-level zonal wind in a two-phase linear regression model. In addition, we show the impact of two surface boundary conditions on the timing of the EASM onset related to the strength of the upper-level zonal wind: 1) eastern Eurasian snow cover and 2) western North Pacific (WNP) sea surface temperature (SST) tendency. The eddy heat fluxes induced by the enhanced eastern Eurasian snow cover develop an anomalous anticyclonic circulation to the northwest, which causes anomalous warm southwesterly flow toward the north. These can make a reversal of the meridional temperature gradient, which results in the early monsoon onset via changes in the upper-level jet. The upper-level jet also responds to the SST tendency in April over the WNP via thermal wind balance and the resultant changes in transient eddy-induced heat transport. Our findings suggest potential sources for seasonal predictability in the interannual EASM onset dates.

1. Introduction

The East Asian summer monsoon (EASM) is accompanied by a strong low-level southwesterly flow along with the western North Pacific subtropical high (WNPSH), which transports high levels of moisture, resulting in summer monsoon rainfall over East Asia (Ding and Chan 2005; Ha et al. 2012). Variations in the EASM can cause disastrous flooding and anomalous dry spells. Thus, understanding the interannual variability of this phenomenon will be beneficial for long-term water management plans of affected countries. The EASM includes mei-yu in China (20°–30°N, 110°–120°E), changma in Korea (30°–40°N, 125°–135°E), and baiu in Japan (25°–35°N, 135°–145°E), and it seems to be related to the seasonal movement of the upper-level jet (Fig. 1). Defining the onset date of the EASM has been controversial since it can be distinct in accordance with the spatial and temporal characteristics of rainfall, with the three major stationary rainy periods from

mid-May (mei-yu and baiu) to late June (changma) (Wang and Xu 1997; Wang et al. 2004; Li and Zhang 2009; Kajikawa et al. 2012). Due to the different regional constraints, this study assumes the definition of the monsoon onset over East Asia as the timing of the northward propagation of the rainband; hence, we will denote the northward shifts of the rainband to latitude 30°–40°N, namely, changma, as the onset timing of the multistate evolution over East Asia. This could reflect the northward movement of the WNPSH accompanying the low-level southwesterly wind along its northwestern edge (Seo et al. 2011).

Commonly, the onset date of monsoon has been characterized by threshold exceedances, either in rainfall (Chakraborty and Agrawal 2017; Moon and Ha 2017) or wind (Li and Zhang 2009; Liu et al. 2015). Li and Zhang (2009) considered the seasonal abrupt changes of wind direction over Asia as onset and withdrawal dates. Park et al. (2015a) determined the onset date of changma by using three threshold-based conditions; meridional gradient of the equivalent potential temperature, the 355-K isotherm of equivalent potential temperature at 850 hPa, and the 5880-gpm isopleth for three consecutive days. Contrary to other monsoon systems with a sharp increase in rainfall, the amount of rainfall over East Asia at onset date

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FIG. 1. The latitude–time cross sections of pentad precipitation (mm day^{-1} ; shading), 850-hPa winds (m s^{-1} ; vectors), and 200-hPa zonal wind (m s^{-1} , red contours) zonally averaged from the (a) mei-yu region (110° – 120°E), (b) changma region (125° – 135°E), and (c) baiu region (135° – 145°E) from April to August.

gradually increases and its variation is more pronounced because of premonsoon rainfall and wind anomalies by transient weather systems (Fig. 2). Thus, these methods are prone to misidentification and do not give robust observational estimates. In other words, threshold definitions for rainfall may not be appropriate for identifying the EASM onset. Cook and Buckley (2009) introduced an alternative and more robust approach to determine the large-scale South Asian summer monsoon onset without specifying a particular threshold. They used the changepoint detection algorithm by adapting a two-stage linear regression to the annual rainfall cycle. Not having defined the onset date by an arbitrary threshold is the advantage of this approach.

As a subtropical monsoon, it is still challenging to predict the timing of the EASM because it is influenced not only by tropical processes, including El Niño–Southern Oscillation (Wang et al. 2000; Wu et al. 2003; Yun et al. 2008; Chu et al. 2012; Oh and Ha 2015) and the Indian Ocean (Xie et al. 2009; Zheng et al. 2014; Kim et al. 2018), but also by midlatitude systems, such as the upper-level jet stream, Rossby waves (or Silk Road pattern), and

transient eddies (Enomoto et al. 2003; Sampe and Xie 2010; Song et al. 2013; Hong and Lu 2016; Sung et al. 2006; Oh et al. 2018). The midtropospheric warm advection caused by the upper-level jet stream can affect the timing and position of the East Asian rainband (Sampe and Xie 2010; Wang

FIG. 2. Climatological time series (1997–2014) of GPCP daily precipitation (mm day^{-1}) over the South Asian summer monsoon (SASM) region (10° – 30°N , 60° – 100°E), East Asian summer monsoon (EASM) region (25° – 40°N , 110° – 145°E), and western North Pacific summer monsoon (WNPSM) region (5° – 20°N , 115° – 150°E).

et al. 2018). Sampe and Xie (2010) also showed the roles of the dynamical disturbances corresponding to transient eddy activity on the EASM system. The associated midlatitude teleconnection by upper-level jet displacement and its structure and energetics were described by Enomoto et al. (2003) and Kosaka et al. (2009), respectively. They revealed that the Silk Road pattern (e.g., Rossby waves through the subtropical jet) could induce an anticyclonic circulation over Japan with an equivalent barotropic structure, which is associated with the interannual variability of the EASM. Considering these facts, the upper-level zonal wind and East Asian rainband are closely related (Zhang et al. 2006), in which meridional shifts are also connected to climatological features including the local meridional temperature gradient, heating over the Tibetan Plateau, and the interaction between the mean flow and transient eddies (Park et al. 2015b; Chen and Bordonì 2016).

Each of these remote teleconnections has its own seasonal characteristics (Stuecker et al. 2015), or the winter-time peak of these remote signals is stored in the other ocean basins (Wu et al. 2009; Xie et al. 2009; Wang et al. 2013; Song and Zhou 2014a,b). The Arctic Oscillation (AO) could play a key role in the variation of the EASM (Gong and Ho 2003; Gong et al. 2011). There are possible mechanisms for how winter or spring AO affects the EASM or how the AO signals persist through the summer. Some explanations suggested that the AO influence the surface land conditions over the entire Eurasian continent (Cohen et al. 2007; Yeo et al. 2017), inducing potential changes in snow cover over the Eurasian continent. Such land surface anomalies may further influence the atmospheric conditions via heat fluxes and the snow–albedo or snow–hydrological feedbacks (Liu and Yanai 2002; Yim et al. 2010; Wu et al. 2014a,b; Zhang et al. 2017). Liu and Yanai (2002) found that excessive Eurasian snow cover corresponds to a decreased temperature as well as a low-level cyclonic circulation anomaly with atmospheric disturbances, which can weaken the EASM. Yim et al. (2010) showed the roles of the two distinct structures of Eurasian snow cover in boreal spring on the EASM rainfall and found the opposite variation of the snow cover is much more closely associated than the continent-wide variability related to the spring AO. The other explanations showed that the influence of the spring/winter AO could be stored by ocean thermal anomalies. Gong et al. (2011) showed that the roles of the air–sea interaction over the western North Pacific (WNP) on the spring AO-related summer circulation. Furthermore, warming over the WNP can influence transient eddy activity and destabilize the low-level atmosphere by the local thermal gradient (Heo et al. 2012; Park et al. 2017). Such sea surface temperature (SST) anomalies can increase low-level baroclinicity with the potential of promoting storm activity over the North Pacific. Park et al.

(2017) proposed that a midlatitude circulation pattern that is associated with the northward upper-level jet stream over East Asia could be further modified by the enhanced atmospheric convection over the warm WNP region. Using an idealized experiment, Ogawa et al. (2012) demonstrated that the response of the storm track to meridional SST anomalies is quite linear between 30° and 45°N latitude can lead to poleward displacements of the upper-level jet core location. In spite of these studies, key questions regarding the dynamical mechanism linking meridional SST anomalies, upper-level winds, transient eddies, and seasonal shifts of the EASM rainband remain unresolved. Given this lack of dynamical understanding, this study aims to determine the timing of the various multistage evolution over East Asia and to evaluate how the interannual changes in land and ocean surface condition can influence the upper-level zonal wind as well as the resultant monsoon frontal system over East Asia.

2. Data and method

a. Observation and reanalysis datasets

Here we used daily surface heat fluxes (shortwave, longwave, sensible, and latent heat fluxes), winds, air temperature, and geopotential height data from 1979 to 2016, reconstructed from the European Centre for Medium-Range Weather Forecast (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011). The horizontal resolution used here is $1.5^\circ \times 1.5^\circ$. Soil moisture data were obtained from the monthly mean $0.5^\circ \times 0.5^\circ$ Leaky Bucket Model, generated by the National Oceanic and Atmospheric Administration Climate Prediction Center (CPC) (Fan and van den Dool 2004). The pentad precipitation data were collected from the Global Precipitation Climatology Project (GPCP) data (Adler et al. 2003) with a $2.5^\circ \times 2.5^\circ$ horizontal resolution from 1979 to 2016, while the daily precipitation from the GPCP was used for the period 1997–2014. In addition, the monthly SST data for the period 1979–2016 were obtained from the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) of the British Atmospheric Data Centre (BADC) (Rayner et al. 2003). Also, monthly snow cover data for the same period were obtained from the Rutgers University Global Snow Laboratory (<http://climate.rutgers.edu/snowcover/>).

b. Characterizing the multistage evolution of the EASM: Two-phase linear regression

Following Cook and Buckley (2009), we initially have defined the EASM onset in terms of the annual cycle of rainfall over East Asia. This definition captures the climatological monsoon onset at certain East Asian monsoon regions well, but it is not consistent with the

interannual variability in onset dates reported in previous studies (Ha et al. 2005; Park et al. 2015a). This is because rainfall events and wind anomalies caused by transient weather systems can lead to misidentifications in the premonsoon stage over East Asia. This emphasizes the need to develop more robust methods to identify EASM variability. Therefore, in this study, we consider zonal winds at 200 hPa, which are less affected by complex weather processes near the surface (Zhao et al. 2015). Lau et al. (2000) proposed that the EASM can be depicted by upper-level vorticity, with a northward advance of the WNPSH, of which process is connected to the tropical heating. We average the upper-level zonal wind (25° – 35° N, 110° – 150° E) and define it as an index, where the month-to-month variation from April to August is the largest (Fig. 3a), representing the large monthly fluctuation of the upper-level zonal wind. Figure 3b shows an example annual cycle of daily zonal wind at 200 hPa in the year 2007, and represents a time-integral index of the annual cycle averaged over 25° – 35° N, 110° – 150° E, starting from 1 January through to the end of the year. The summer monsoon period is broadly characterized by a near-zero slope in the time-integral (cumulative) distribution, indicating a weakening or reversal of westerlies in summer. To make use of these features, we fit a two-phase linear regression line to the cumulative upper-level zonal wind time series from April to August. The equation is shown below:

$$X_1 = \alpha_1 t + \beta_1, \quad 1 \leq t \leq c; \quad X_2 = \alpha_2 t + \beta_2, \quad c \leq t \leq n,$$

where α and β are the regression coefficients, t is time series under consideration, and n is the length of the regression fit time series, which is 153 from April to August in this study (Fig. 3b). The optimal change-point, c , is determined through an iterative fitting of the two-phase linear regression from time $t = 3$ to $t = n - 3$. X_1 and X_2 are the regression model. We set the onset date of the multistages as the point when the regression lines are close to a set of points (appendix A; Fig. A1). Climatologically, the onset date of the EASM is 13 June, when the upper-level zonal wind starts to decrease (Fig. 3b). It can be confirmed by the Gaussian distribution, which represents the probability for the given upper-level zonal wind. When the decreasing tendency of the upper-level zonal wind is distributed with a high probability, it occurs frequently (Fig. 3c). Around the monsoon onset dates (13 June), it can be shown that abrupt changes in the upper-level jet being dropped in most of the years exist. When we compare the monsoon onset date with the days representing each season (1 March, 1 June, 1 September, and 1 December), we can find, through an

analysis of the Gaussian distribution, that a sudden decrease (tendency: $-1.9 \text{ ms}^{-1} \text{ s}^{-1}$) in the upper-level zonal wind occurs on 13 June (Fig. 3c). We separately apply the two-phase linear regression model to the daily 200-hPa zonal wind in each year and determine 38-yr onset dates (Fig. 4). The standard deviation is 8.75 days, similar to the previous other EASM onset indices (Ha et al. 2005; Park et al. 2015a; Li et al. 2018a). It represents not only the timing of the northward movement of the rainband but also the start date for changma along 30° – 40° N. The variations in the EASM onset dates are represented by changing the length of the time series n from 149 to 157 to validate the EASM onset indices. The onset dates that we defined are not changed regardless of the length of the time series n . The onset date of the EASM in this study is shown in Table 1, and it can depict East Asian monsoon circulation and rainfall affected by conditions in both the tropics and midlatitudes (Park et al. 2015a; Li et al. 2018b). It can also reflect the northward movement of the EASM rainband. The climatological changma onset dates range from days 172 to 174 (21 June to 23 June) (Ha et al. 2005; Park et al. 2015a; Moon and Ha 2017), while the climatological mean onset in our study is 8–10 days earlier. The interannual variations in the onset dates are robust across data collected via the National Centers for Environmental Prediction (NCEP)–Department of Energy (DOE) Reanalysis 2 (Kanamitsu et al. 2002), Modern-Era Retrospective Analysis for Research and Applications (Rienecker et al. 2011), and the Japanese 55-year Reanalysis Project (Ebata et al. 2011) (Fig. 4).

c. Baroclinic energy conversion

To figure out the dynamic linkages among the upper-level zonal wind, transient eddies, and meridional thermal gradient with the EASM, we investigate the characteristics of the energetics. The baroclinic generation, from mean available potential energy (MAPE) to eddy available potential energy (EAPE) is roughly a multiplication of both the poleward eddy heat flux and meridional temperature gradient for baroclinic energy conversion (BCEC) I. Also, BCEC II describes from EAPE to eddy kinetic energy (EKE), which can be explained by upward heat flux (Cai et al. 2007; Lee et al. 2012):

$$C_1 = \left(\frac{p_o}{p} \right)^{c_v/c_p} \frac{R}{g},$$

$$C_2 = C_1 \left(\frac{p_0}{p} \right)^{R/C_p} \left/ \left(-\frac{d\theta}{dp} \right) \right.,$$

$$\text{BCECI} = -C_2 \left(\overline{u'T'} \frac{\partial \bar{T}}{\partial x} + \overline{v'T'} \frac{\partial \bar{T}}{\partial y} \right),$$

FIG. 3. (a) Month-to-month variation in the zonal wind at 200 hPa (m s^{-1} ; shading) from April to August and May–June climatology of the zonal wind at 200 hPa (m s^{-1} ; black contours) from ERA-Interim. The dashed green box (25° – 35°N , 110° – 150°E) is used as the region for the monsoon onset index in this study. (b) An example annual cycle of ERA-Interim daily upper-level wind averaged over 25° – 35°N , 110° – 150°E in the year 2007 (left y axis; m s^{-1} ; gray line), the 11-day running mean (left y axis; m s^{-1} ; black), the time-integral of the annual cycle (right y axis; m s^{-1} ; navy blue line), and the two-phase regression line of the time-integral time series (red dotted line). (c) Probability of the tendency of the zonal wind at 200 hPa from March to August, with the representative dates for each season (color lines), and the climatological onset date (black line). The vertical gray lines in (b) and (c) indicate the climatological onset date (13 Jun).

$$\text{BCEC II} = -C_1(\overline{\omega T'}),$$

where $p_0 = 1000 \text{ hPa}$, p is pressure, C_v (C_p) is the specific heat of dry air at the constant pressure (volume), R is the gas constant for dry air, and g is the acceleration of gravity. The overbars show a specific period (monthly time mean), and the 3–8-day bandpass filter is represented as the primes.

3. Characteristic of the multistage over East Asia and its transition timing

a. Three stages over East Asia

We set the onset dates as day 0 and examined the distinct step-wise characteristics depending on the northward movement of the EASM rainband. As aforementioned, the interannual variability of the onset date is similar to that of

FIG. 4. Time series of the onset dates over East Asia across the datasets: ERA-Interim, NCEP2, JRA55, and MERRA. The interannual variability (IAV; purple) of the onset dates is 2–8 years bandpass filtered from the original onset dates obtained from the ERA-Interim to focus on the IAV. The IAV accounts for 81% of the time series for the original onset dates. The x axis and left and right y axes denote years, onset dates, and normalized IAV of the onset dates, respectively.

the previous EASM onset indices (Ha et al. 2005; Park et al. 2015a), but ours is approximately 9 days earlier. This gap is caused by our definition representing the beginning of its transition defined by the changepoint rather than the first day of monsoon. Thus, the onset date in our study could explain the progression of the East Asian frontal system during the three stages. Figure 5 shows the anomalous precipitation and the associated circulation patterns at the three consecutive stages over East Asia. Each stage has a length of 9 days: the first stage (day -9 to -1), the second stage (day 0 to $+8$), and the third stage (day $+9$ to $+17$). The length of 9 days was averaged to show each stage without an overlap among the stages and to remove synoptic-scale fluctuations. A zonally elongated rainband extends from southern China to southern Japan, which results from southwesterly winds from the tropics; this rainband begins to move northward around the southern part of the Korean Peninsula, which is eventually characterized by a tripole pattern with positive rainfall anomalies on the Korean Peninsula and WNP (Fig. 5). The EASM is well established toward 30° – 40° N during the stages. The onset of mei-yu is consistent with a strong upper-level jet (Figs. 5a and 5d); then the jet stream is weakened or shifted to the north during the other stages (Figs. 5b and 5c).

Especially for the third stage, the upper-level jet is much more advanced to the north, with the enhanced low-level WNPSH which transports high levels of moisture (Figs. 5c and 5f).

b. The differences in the upper-level zonal wind between early- and late-onset years

To focus on the interannual variability, we have applied a 2–8-yr bandpass filter from the original datasets using a Lanczos filter (Fig. 3). Based on the interannual variability of the onset dates, the early- (late-) onset years are defined by deviations of less (greater) than 0.8 times the standard deviation. Figure 6a shows the latitude–time cross section of composite difference in pentad precipitation, averaged along 125° – 135° E, for early minus late-onset years. The early (late) onset indicates that northward propagation of rainfall occurs much earlier (later) than climatology; it also causes an increase in precipitation over northern (southern) East Asia (Fig. 6a). There are nine early- (nine late-) onset years: 1980, 1984, 1991, 1993, 1996, 1999, 2000, 2004, and 2013 (1982, 1983, 1988, 1992, 1995, 1998, 2002, 2014, and 2016). The mean onset dates of the early- and late-onset years are 153 (2 June) and 174 (23 June), respectively, which are 22 days apart, the difference of

TABLE 1. The onset dates of EASM in each year.

Year	Date	Year	Date	Year	Date	Year	Date	Year	Date
1979	10 Jun	1987	3 Jun	1995	26 Jun	2003	17 Jun	2011	9 Jun
1980	24 May	1988	12 Jun	1996	5 Jun	2004	4 Jun	2012	18 Jun
1981	11 Jun	1989	6 Jun	1997	15 Jun	2005	12 Jun	2013	1 Jun
1982	29 Jun	1990	9 Jun	1998	22 Jun	2006	14 Jun	2014	22 Jun
1983	1 Jul	1991	25 May	1999	31 May	2007	11 Jun	2015	18 Jun
1984	8 Jun	1992	22 Jun	2000	2 Jun	2008	14 Jun	2016	18 Jun
1985	20 Jun	1993	9 Jun	2001	10 Jun	2009	16 Jun		
1986	9 Jun	1994	14 Jun	2002	21 Jun	2010	18 Jun

FIG. 5. Spatial distributions of (a)–(c) mean 200-hPa winds (m s^{-1} ; contours) and its anomaly (m s^{-1} ; shading) and (d)–(f) anomalous 850-hPa winds (m s^{-1} ; vectors) and anomalous precipitation (mm day^{-1} ; shading) over the multistage of the EASM: (a),(d) stage 1 (day -9 to -1), (b),(e) stage 2 [onset date (day 0) to $+8$], and (c),(f) stage 3 (day $+9$ to 17). The anomalies are extracted from the daily precipitation climatology.

almost a month. When the EASM rainband moves earlier, the amount of rainfall over East Asia tends to be much larger along 30° – 40°N in June. It implies that the onset dates we defined are related to the start date of changma, showing a much larger amount of precipitation in June over the Korean Peninsula. Also, the potential sources for seasonal predictability of the EASM onset could be shown on a monthly time scale. To validate the monsoon onset indices, we found null distributions of a fixed seasonal cycle of the upper-level zonal wind plus random amplitude noise which is determined from the observations. When we calculated the probability density function with this simple null distribution, its standard deviation of this estimation error (3.75 days) is less than ours (8.75 days). This implies that the extreme onset dates such as early (2 June) and late monsoon (23 June) onset dates are significant because they exceed two standard deviations of the random distribution (appendix B; Fig. A2).

Here we have represented the annual cycles of the upper-level zonal wind for early- and late-onset years (Figs. 6b and 6c). There is a significant difference in the wind intensity from mid-April to June. In other words, when the northward

evolution of the jet stream occurs earlier, it could lead to the northward shift rainband and simultaneously correspond to a decreased rainfall over the south of 30°N . Here the question arises, what causes the intensity difference of the upper-level zonal wind to be much larger?

4. Possible mechanisms for the northward timing of the EASM onset

The upper-level zonal wind is closely connected to local meridional gradient of temperature, heating of the Tibetan Plateau, interaction between the mean flow and transient eddies, and the atmospheric circulation teleconnection patterns (Sampe and Xie 2010; Gong et al. 2011; Ogawa et al. 2012; Thompson and Birner 2012; Chiang et al. 2017). Since a significant difference in the upper-level zonal wind between the early- and late-onset years is demonstrated in April (Fig. 6c), we have checked the persistent and tendency 2-m air temperature over land and SST from March to May (Fig. 7). When monsoon onset occurs earlier, it features a negative surface temperature in March and April over the eastern Eurasian region with a warming

FIG. 6. The latitude–time cross section of composite difference (shading) in (a) pentad rainfall (mm day^{-1}) averaged over 125° – 135°E and (c) daily upper-level zonal wind averaged over 110° – 150°E with respect to early-minus late-onset years with each climatology (contours). (b) Composite annual cycles of zonal wind at 200 hPa (left y axis; m s^{-1}) at 25° – 35°N , 110° – 150°E for early-onset (red) and late-onset (blue) years, and their accumulative indices (right y axis; m s^{-1}). The dots in (a) and (c) indicate values significant at the 95% confidence levels with 16 degrees of freedom using the Student's t test. The dashed lines in (c) indicate the region (25° – 35°N) where we averaged in (b). Here the early- (late-) onset years are 1980, 1984, 1991, 1993, 1996, 1999, 2000, 2004, and 2013 (1982, 1983, 1988, 1992, 1995, 1998, 2002, 2014, and 2016).

tendency toward May over the WNP, which could act as precursors for the timing of the EASM onset.

a. Eastern Eurasian snow cover in March

The negative AO simply drives the low pressure in the midlatitudes, inducing greater movement of frigid polar air and extensive snow cover over the entire Eurasian continent by changes in atmospheric circulation (Fig. 8a). Besides, the variability of snow cover over the eastern Eurasian continent could lead to changes in the local temperature. The relationship between the snow cover and EASM onset dates is negatively significant (-0.50), which means that the greater snow cover in March is favorable to earlier monsoon evolution over East Asia. The associated soil moisture plays an important role in the linkage between the snow cover in March and the long memories of the land thermal condition in the following months. Zhang et al. (2017) mentioned that snowmelt in spring corresponds to the anomalous wet local soil condition in the following summer, which is accompanied by decreasing surface heat flux and near-surface temperatures. The eastern Eurasian snow cover is linked to the soil moisture from March to May (Fig. 8b). The anomalous wet soil moisture from March to

May implies a change in the local energy balance by the changes in snow cover. We have investigated the roles of snow cover on thermal condition via analyzing the surface heat fluxes, including changes in the snow–albedo and snow–hydrological effects (Chen et al. 2016). The surface heat fluxes could be affected by the presence and absence of clouds. The increase in cloudiness, especially low- and midlevel cloud cover, leads to a decrease in downward shortwave radiation reaching the surface and an increase in downward longwave radiation. In addition, the enhanced snow cover may lead to an increase in soil moisture while the snow is melted in boreal spring; the increased soil moisture results in cooler air temperature due to evaporation (Table 2). The variability of snow cover in March has lasted until April. In April, a strong cyclonic circulation under the negative AO persists over the eastern Eurasian continent, corresponding to greater snow cover and greater soil moisture (Table 2 and Fig. 9a). Climatologically, much of the snowmelt occurs in May. There are persistent cooling SSTs in May sustained by anomalous northerlies from the high-latitude and equatorward heat fluxes which lead to a westward acceleration of the low- and midlevel mean flow over East Asia (Hartmann 2007). The associated anticyclonic circulation over the continent brings warm air to the

FIG. 7. The regressed field of the 2-m air temperature and sea surface temperature (T2m and SST, respectively; K) in (a) March, (b) April, and (c) April T2m and SST tendencies (May minus March) onto the interannual variability of EASM onset dates. Data are from the ERA-Interim and HadISST datasets. The dots indicate values significant at the 95% confidence levels with 36 degrees of freedom using the Student's t test. The designated boxes (40° – 55° N, 110° – 130° E) in (a) and (15° – 25° N, 135° – 170° E) in (c) are used as potential sources for seasonal predictability in the EASM onset.

north. Thus, the strong warm advection in the low- and midtroposphere by the southerlies in May exists, which results in the reversed meridional thermal gradient over the eastern Eurasian region (Figs. 9b and 10a). This continental warming has been accelerated by much more downward turbulence heat flux over the anticyclonic circulation. Thus, the near-surface thermal conditions associated with the increases in snow cover could substantially change the local

FIG. 8. (a) Interannual variability of the normalized March snow cover over eastern Eurasia (dashed navy blue line; 40° – 55° N, 110° – 130° E) and March Arctic Oscillation (black line), and (b) lead-lag relationships between the eastern Eurasia snow cover in March and Arctic Oscillation, zonal wind at 200 hPa (30° – 35° N, 110° – 150° E), soil water (40° – 55° N, 110° – 130° E), air temperature at 850 hPa (40° – 55° N, 110° – 130° E), and evaporation (40° – 55° N, 110° – 130° E). The gray dashed lines indicate values significant at the 95% and 99% confidence levels with 36 degrees of freedom using the Student's t test.

atmospheric temperature at 850-hPa (Fig. 8b). Due to the thermal redisplacement, the jet stream weakens in the entrance zone in June and an anomalous anticyclonic circulation at the low-level over southern Japan is followed. It is revealed that the low-level southerly flow along with the anticyclonic circulation over Japan forms the warm advection, resulting in the adiabatic ascent and the associated East Asian rainfall (Figs. 9c and 10b) (Wang et al. 2018). In other words, the surface thermal condition in the midlatitude becomes warming in the following month while the enhanced snow cover in March is melted. Thus, the upper-level zonal wind could be weakened or be meridionally shifted due to the thermal redisplacement. It is followed by an earlier transition of the EASM rainband via secondary circulation induced by the changes in the jet stream.

b. April SST tendency over WNP

The changes in atmospheric circulation in early summer around East Asia can be associated with warming over the WNP, and they are consistent with the changes in eddy activities. The second factor determining the timing of the EASM onset is May minus March

TABLE 2. Regression coefficients of various variables with respect to snow cover over the eastern Eurasian continent (40° – 55° N, 110° – 130° E). The single and double asterisks (*) and **) denote the significance at the 90% and 95% confidence levels, respectively, according to the Student's t test. Variables include snow cover (SC; %), surface air temperature (SAT; $^{\circ}$ C), soil moisture (Soilw; cm), low cloud cover (LCC), medium cloud cover (MCC), surface shortwave radiation (SW; W m^{-2}), surface longwave radiation (LW; W m^{-2}), sensible heat flux (SH; W m^{-2}), and latent heat flux (LH; W m^{-2}). A positive sign indicates a downward flux.

	SC	SAT	Soilw	LCC	MCC	SW	LW	SH	LH
March	0.79**	-0.68**	0.26*	0.46**	0.32**	-0.28*	0.39**	-0.29**	0.07
April	0.45**	-0.45**	0.31**	0.11	0.32**	-0.12	0.23	-0.17	0.17

SST tendency over the WNP (15° – 25° N, 135° – 170° E) (i.e., April SST tendency). The warming signal and the relevant dipole circulation pattern are constantly maintained in May but disappeared in June (Figs. 11c and 11f).

FIG. 9. The regressed field of geopotential height at 500 hPa (m; contours), winds at 850 hPa (m s^{-1} ; vectors), and air temperature at 2-m (hatched lines) in (a) April, (b) May, and (c) June onto the eastern Eurasian snow cover index (40° – 55° N, 110° – 130° E) in March. The positive (negative) shading represents 500-hPa poleward (equatorward) eddy heat flux $v'T'$ in (a) and (b) and precipitation (mm day^{-1}) in (c). The red (blue) hatched lines indicate warm (cold) values significant at the 95% confidence levels with 36 degrees of freedom using the Student's t test.

The SST tendency over the WNP may provide a favorable environment for eddy activity through baroclinic dynamics caused by changing the meridional thermal gradient across the Kuroshio Extension via joining the Kuroshio. Thus, we should understand how the responses of the SST tendency over the WNP affect the eddy activity in the midlatitudes via baroclinic energetics

FIG. 10. The latitude–pressure cross section of regressed fields of horizontal temperature advection (K day^{-1} ; shading), meridional wind and vertical velocity (vectors), and zonal wind (m s^{-1} ; contours) averaged along 110° – 130° E in (a) May and (b) June for the interannual variability of the March eastern Eurasian snow cover (40° – 55° N, 110° – 130° E). The black dots indicate values significant for each shaded variable at the 95% confidence levels with 36 degrees of freedom using the Student's t test.

FIG. 11. The regressed field of 500-hPa energy conversion from (a),(d) eddy available potential energy (EAPE) to eddy kinetic energy (EKE), (b),(e) meridional eddy heat flux, and (c),(f) meridional thermal gradient in (top) May and (bottom) June onto the interannual April WPSST tendency (15° – 25° N, 135° – 170° E). The contours indicate geopotential height at 500 hPa (m). The black dots indicate values significant for each shaded variable at the 95% confidence levels based on the Student's *t* test. The green (orange) hatched lines in (d) indicate positive (negative) precipitation and the red (blue) hatch lines in (c) and (f) represents warm (cool) 2-m air temperature significant at the 95% confidence levels with 36 degrees of freedom based on the Student's *t* test.

(Fig. 11). To diagnose the atmospheric circulation, we have investigated the eddy properties and local energetics. In general, the positive baroclinic energy conversion is an important energy source for the development and enhancement of transient eddy activity, and it is roughly proportional to the poleward eddy heat flux multiplied by the meridional temperature gradient (Cai et al. 2007). In May, a northeastward-moving WNP SST can provide the additional heat and moisture into the storm track (160° E– 180°) regions, influencing the local baroclinic growth. It can maintain the intense maximum Eddy growth rate between 850 and 700 hPa (figure not shown), which represents the potential energy from the mean circulation to the eddies (Fig. 11). The enhanced eddy kinetic energy, computed from the 3–8-day bandpass-filtered daily fields, can modulate the cyclone vorticity to the north and anticyclone vorticity to the south, which represents the dynamic forcing of the transient eddies (Lau 1988; Gong et al. 2011). In June, the anomalous anticyclonic circulation over Japan is simultaneously related to the poleward (equatorward) heat fluxes to the north (south), which can lead to the eastward (westward) acceleration of the mean flow (Fig. 11e). This anomalous anticyclonic circulation gives rise to the northward progression of the EASM rainband, resulting in an earlier monsoon transition. It is reasonably concluded that the April SST tendency over the WNP plays an important role in the timing of East Asian rainband transition by interacting with the eddy activity.

Since both surface boundary forcings play an important role in the variation of the monsoon circulation in June, we expect that the synergy is amplified when they occur during the same phase. Now we investigate the combined effect of the eastern Eurasian snow cover and April SST tendency over the WNP on the timing of the EASM onset. We calculate the composite difference between the positive (high snow cover and high WNP SST tendency) and negative (low snow cover and low WNP SST tendency) in-phase years

FIG. 12. The latitude–time cross section of composite difference in the zonal wind at 200 hPa averaged over 110° – 150° E (m s^{-1} ; contours), winds at 850 hPa (m s^{-1} ; vectors) and precipitation (mm day^{-1} ; shading) averaged over 125° – 135° E between positive and negative in-phase years for combined effect of the March eastern Eurasian snow cover (40° – 55° N, 110° – 130° E) and April SST tendency over WNP (15° – 25° N, 135° – 170° E). The black patterns indicate values significant for precipitation at the 90% confidence levels based on the Student's *t* test.

based on ± 0.33 standard deviations. Seven years correspond to a positive in-phase relationship (1980, 1981, 1984, 1991, 2001, 2007, and 2011) and another seven years denote a negative in-phase relationship between the two factors (1982, 1990, 1992, 1995, 1998, 2002, and 2014). The upper-level zonal wind along 25° – 35° N has been changed from April–May to June, indicating that it has been weakened or shifted to the north in June (Fig. 12). The two combined effects greatly contribute to the timing of rainband evolution, upper-level jet, and the relevant low-level winds supplying an abundant amount of heat and moisture from the warm WNP toward East Asia. The two boundary conditions affect the timing of the EASM onset, although the two indices are independent (correlation coefficient: 0.10).

5. Summary and discussion

A comprehensive understanding of the seasonal and interannual variation in the upper-level zonal wind is important to develop the improved predictions of the EASM (Lau et al. 2000; Liu and Yanai 2002; Zhao et al. 2015). This study determined the timing of the northward evolution rainband over East Asia and analyzed the physical mechanisms caused by the local surface boundary forcings. Following Cook and Buckley (2009), who used daily precipitation over India to determine the onset dates using the two-phase linear regression, we adjusted the changepoint detection of the accumulated zonal wind at 200 hPa, which is less affected by the complex topography near the surface or the weather processes. The definition of the multistages over East Asia should well capture the seasonal transitions in rainfall and winds at all levels, not requiring a specific threshold of rainfall or circulation. The climatological onset date for the period 1979–2016 is on 13 June. The mean early (late) onset date is 2 June (23 June) when the onset dates are defined by a deviation of less (greater) than 0.8 times the standard deviation, which causes almost a month's difference. Thus, the mechanisms for determining the timing of the EASM could be shown on a monthly time scale. The intensity of the upper-level zonal wind in April–May and June was shown to be a significant difference between the early- and late-onset years. For example, the strength of the westerly wind at 200 hPa for the early-onset years has been rapidly weakened or dropped due to the northward shifted jet location to 40° N much earlier than that for the late-onset years, with the changes in a local meridional thermal gradient.

Figure 13 summarizes the mechanisms of the two factors on the timing of the EASM onset. In April, there is the enhanced snow cover over the eastern Eurasia region under the negative AO, which corresponds to a cool

FIG. 13. Schematic diagram to illustrate how the two local surface boundary forcings advance the EASM onset.

near-surface thermal condition at the high latitudes. In May, the equatorward heat fluxes caused by the northerly anomalies lead to the westward acceleration of the mean flow over East Asia and the associated anticyclonic circulation to the northwest exists. The warm southwesterly flow along the western flank of the anticyclone contributes to the rapid temperature increase over eastern Eurasia. It gives rise to the weakened jet stream in the entrance zone via the thermal redisplacement and an anomalous anticyclone circulation at low levels over Japan in June. The low-level southerly flow along with the relevant anticyclonic circulation forms the warm advection and this causes the adiabatic ascent and more rainfall. Together with the eastern Eurasian snow cover effect, the warm WNP leads to the intensified poleward heat fluxes (160° E– 180°) and the associated cyclonic circulation to the north. It induces the northerly flow at high latitudes, which accelerates the equatorward heat fluxes in May over East Asia when the two boundary forcings are positively in-phased. Especially, in June, the equatorward heat fluxes over Japan induce the westward acceleration of the mean flow, which forms the anomalous anticyclonic circulation to the north and then shifts the East Asian rainband to the north. The interaction between extratropical eddies and the associated mean flow could determine the timing of the northward movement of the EASM rainband. We should remember that the March AO could be a potential

contributor to the timing of EASM transition by inducing the changes in SST distribution over the extratropics and surface cooling over northern East Asia (Gong et al. 2011). The March AO and snow cover are significantly correlated (-0.52) at the interannual time scale.

Oh and Ha (2016) have found physically meaningful and statistically robust predictors of dominant boreal summer intraseasonal modes over East Asia, based on the persistent and tendency signals of the surface boundary conditions using a physical-empirical model. The first mode, called the pre-mei-yu and baiu mode, have three predictors such as SST cooling tendency over the Kuroshio, persistent 2-m air temperature in the preceding winter over northern East Asia, and SST over the eastern Indian Ocean, which are similar to our study but with opposite signs. The rain-band in the pre-mei-yu and baiu mode does not propagate to the north, and hence its frequency of occurrence is opposite to the EASM onset timing. This means that the number of the pre-mei-yu and baiu mode would be increased if the timing of onset date over East Asia is delayed.

What about the trends of onset dates depending on various forcings? We have checked the observed and CMIP5-simulated annual cycles of upper-level zonal wind at 250 hPa under historical, historicalGHG, and historicalNat experiment, which includes all-forcing, only well-mixed greenhouse gases, and only natural variations, respectively, along with anthropogenic and aerosol forcing. The anthropogenic forcing is defined by the difference between historical and historicalNat runs, and the aerosol forcing is determined by the difference between anthropogenic and historicalGHG runs (Song et al. 2014). When we take a look at linear trends of onset dates using the CMIP5 models for the period 1979–2005 under different forcing experiments, both observed and historical simulated variations of the onset date are similar, with no significant trend. The interesting point is that only the anthropogenic aerosol forcing produces a significant trend, showing a delaying trend of the onset dates [$+18.9$ days $(27 \text{ yr})^{-1}$]. Song et al. (2014) mentioned that the EASM circulation patterns are weakened due to the continental cooling over East Asia under the aerosol forcing. In particular, the upper-level zonal wind is shifted southward. Chu et al. (2018) emphasized that the initial surface radiative cooling over the north of 30°N is responsible for the southward shift in the East Asian rain belt, indicating the late monsoon onset. This partly explains the reason why the onset dates are delayed over time in the anthropogenic aerosol forcing. Also, in the future, tropical monsoon shows clear delay under global warming (Biasutti and Sobel 2009; Song et al. 2018), but it is still in doubt whether the EASM, as a subtropical monsoon, behaves similarly or differently

FIG. A1. The schematic diagram for the definition of monsoon onset in this study. The annual cycle of accumulative daily upper-level wind averaged over $25^\circ\text{--}35^\circ\text{N}$, $110^\circ\text{--}150^\circ\text{E}$ (navy blue line). In the equations, X_1 and X_2 are the regression models, and c is a changepoint when the original cumulative index and the two-phase linear regression line have a minimum mean square error, which is determined through several iterations.

from the tropical monsoon because of considerable uncertainty about the dynamical forcing. The future change in the subtropical monsoon and its fundamental reason would be investigated in further study.

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APPENDIX A

Changepoint Detection for Onset Dates

We constructed the cumulative time series of the upper-level zonal wind, and then we used a two-phase linear regression of the cumulative time series to detect a changepoint c over a range from 1 April to 31 August:

$$X_1 = \alpha_1 t + \beta_1, \quad 1 \leq t \leq c; \quad X_2 = \alpha_2 t + \beta_2, \quad c \leq t \leq n,$$

where α and β are the regression coefficients, t is time series under consideration, and n is the length of the regression fit time series from April to August in this study. On the left-hand sides X_1 and X_2 are the regression models. The changepoint c is determined through the iterative fitting of the two-phase linear regression when the regression lines are close to a set of points (Fig. A1).

FIG. B1. (a) The annual cycles including 100 random samples (gray line) and their cumulative index (navy blue lines) and (b) probability density function with the random distribution. Here the x axis is Julian days. The blue dashed lines in (b) indicate two standard deviations of the random distribution.

APPENDIX B

A Statistical Validation for Onset Dates

We made 100 random samples (the range of noise is from minimum to maximum in observation), which were accumulated (Fig. B1a). Then, we calculated 100 random onset points and made the Gaussian distribution (Fig. B1b). Their average day is day 163.7 and standard deviation is 3.75 days. According to the interannual variability of the onset date we defined, the extreme onset dates such as early (31 May) and late monsoon

(24 June) onset dates exceed two standard deviations of the random distribution. It supports our results.

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