An Intraseasonal Mode of Atmospheric Variability Relevant to the U.S. Hydroclimate in Boreal Summer: Dynamic Origin and East Asia Connection

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ABSTRACT

Intraseasonal modes of atmospheric variability over the Northern Hemisphere (NH) midlatitudes in boreal summer are identified via an empirical orthogonal function (EOF) analysis of the daily 10–90-day bandpass-filtered 250-hPa streamfunction for the period of 1950–2016. The first two EOF modes are characterized, respectively, by (i) a single-signed streamfunction anomaly that extends across the NH and (ii) a regional dipole structure with centers over the Aleutian Islands and northeastern Pacific. The third EOF mode (EOF-3) is a quasi-stationary wave train over the Pacific–North American sector with an equivalent barotropic structure in the vertical. EOF-3 is associated with a northwest–southeast oriented anomalous precipitation dipole over the United States. A nonmodal instability analysis of the boreal summer climatological flow in terms of the 250-hPa streamfunction reveals that one of the top “optimal mode” disturbances mimicking the EOF-3 structure grows from an initial precursor disturbance over East Asia through extracting kinetic energy from background flow and attains its maximum amplitude in around nine days. An additional lag regression analysis illustrates that anomalous latent heating associated with cloud and precipitation formation over East Asia is responsible for generating the precursor disturbance for the EOF-3-like optimal mode. This result suggests the existence of an important connection between the hydrological cycles of East Asia and North America, which is dynamically intrinsic to the boreal summer upper-tropospheric flow. Knowledge of such a connection will help us better understand and model hydroclimate variability over these two continents.

1. Introduction

Extreme hydrological events include heavy precipitation as well as droughts, both of which have profound impacts on human life, health, and socioeconomic well-being (e.g., Easterling et al. 2000). In the continental United States, flood and drought are among the top three weather-related causes of death and property damage. A better understanding of extreme hydrological events and improved capability in predicting their occurrence can greatly reduce death tolls and financial losses caused by these extremes. Numerous studies have shown that hydrological extremes are related to global climatic teleconnections that have spatial scales near 10^4 km, such as El Niño–Southern Oscillation (ENSO) and Pacific decadal oscillation (PDO). For example, during El Niño winters, the altered heating pattern over the eastern tropical Pacific favors a pattern shift in the jet stream and associated storm tracks: A stronger-than-normal subtropical branch of the jet stream extends into the

United States and causes enhanced precipitation over the southern half of the country (Raubert et al. 2005). ENSO and PDO can also link together to affect U.S. hydroclimate (McCabe and Dettinger 1999). When ENSO and PDO are in phase (i.e., El Niño and warm PDO or La Niña and cold PDO), precipitation anomalies over the Great Plains are greatly intensified (Hu and Huang 2009).

Global climatic teleconnections and extreme hydrological events are usually connected by large-scale meteorological patterns that have spatial scales greater than mesoscale (10–100 km) but smaller than planetary scale (10^4 km). For example, a positive SST anomaly over the central/eastern tropical Pacific (e.g., due to El Niño) may result in changes in the characteristics of extratropical Rossby wave propagation (Chen 2002; Yang et al. 2002; Seager et al. 2010). These changes in large-scale meteorological patterns are ultimately responsible for modifying regional water and energy budgets (e.g., changing moisture transport patterns) and producing mesoscale weather event (e.g., thunderstorms) outbreaks that are directly responsible for the occurrence of individual hydrological extremes (Zhao et al. 2016, 2017a). During

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boreal summer, a circumglobal teleconnection (CGT) pattern having a zonal wavenumber-5 structure is a major source of climate variability and predictability in the mid-latitude regions including Europe, Asia, and North America (e.g., [Ding and Wang 2005](#); [Wang et al. 2012](#); [Teng et al. 2013](#)). In terms of the impact of the CGT pattern on the U.S. hydroclimate, [Ding and Wang \(2005\)](#) proposed that an enhanced Indian summer monsoon initially generates a west-central Asian high that then excites a Rossby wave train extending downstream to the North Pacific and North America along the extratropical waveguide.

In addition to the global-scale CGT pattern, previous studies have also specifically focused on the Pacific–North American sector. [Zhao et al. \(2017b\)](#) found that, beyond the impact of the PDO and ENSO, interannual variations in the U.S. summer dryness typically have a northwest–southeast (NW–SE) oriented dipole structure, a downstream manifestation of a quasi-stationary wave train originating from the North Pacific and extending through North America to the North Atlantic. [Schubert et al. \(2011\)](#) showed that the fourth rotated empirical orthogonal function (REOF) of the 250-hPa monthly mean meridional wind exhibits a Rossby-wave-like structure and played a major role in the development of two extreme summer hydrological events in the United States: the 1988 drought and 1993 flood. The results of a stationary wave model diagnosis suggested that transient vorticity fluxes are the dominant forcing of the REOF wave, consistent with the results of [Mo et al. \(1995\)](#) and [Liu et al. \(1998\)](#). [Zhu and Li \(2016\)](#) provided evidences suggesting that a NW–SE dipole rainfall pattern (similar to that in [Zhao et al. 2017b](#)) on interannual time scales is triggered by heating over subtropical East Asia via an upper-level Rossby wave train during summer through the application of a linear anomaly atmospheric general circulation model. They found that it takes about 30 days for an initial disturbance over subtropical East Asia to travel through the North Pacific to North America.

Barotropic instability analysis has been widely used to study the dynamics of large-scale meteorological patterns that are linked to midlatitude climate systems. Early studies examined growth rates, mechanisms, and criteria of barotropic instability associated with idealized basic flows (e.g., [Hoskins 1973](#); [Baines 1976](#); [Mak 1983](#)). [Mak and Cai \(1989\)](#) examined the modal (associated with one normal mode; either an exponentially intensifying or decaying pattern) and nonmodal (associated with growth comprising more than one normal mode) instability of an idealized barotropic jet streak. They found that (i) the unstable normal modes (i.e., intensifying normal modes) of the jet streak mainly appear in the downstream region of the jet and (ii) the maximum instantaneous growth rate of a nonmodal

disturbance could be several times greater than that of the most unstable normal mode. Nonmodal instability was also investigated in other studies related to explosive cyclogenesis, numerical weather prediction, and large-scale dynamics (e.g., [Farrell 1988, 1989a,b](#); [Molteni and Palmer 1993](#)). [Huang \(1999\)](#) studied the scale-dependent characteristics of the optimal (nonmodal) growth (which indicates that initial disturbances would accumulatively intensify the *most* over a specified time interval τ) for disturbances on an idealized zonally asymmetric barotropic flow. For an optimization time (i.e., time interval τ) of three days, a smaller-scale initial perturbation may trigger an optimal mode with larger amplification factor. Further energetics analysis indicated that the energy conversion in the optimal growth is dominated by the latitudinal shear straining term ($-u'v'\partial\bar{u}/\partial y$).

Since the 1980s, many studies have explored barotropic instability of the observed time-mean flows. [Simmons et al. \(1983\)](#) were the first to show that low-frequency oscillations (e.g., the Pacific–North American pattern) during boreal winter are related to the most rapidly growing normal mode of the observed climatological mean flows. They also found that the fastest growing mode occurs in the Pacific jet exit, where the deformation term $[(v'^2 - u'^2)\partial\bar{u}/\partial x]$ provides the largest barotropic energy conversion. Subsequent studies cast doubt on the role of modal instability (associated with only one normal mode) in producing wintertime low-frequency oscillations (e.g., [Anderson 1991](#); [Borges and Sardeshmukh 1995](#)). Because of potential shortcomings of the modal perspective, nonmodal instability analysis (NIA; related to multiple normal modes) was subsequently utilized. [Borges and Hartmann \(1992\)](#) demonstrated, in a divergent barotropic model with basic flows defined by the 1985/86 wintertime-mean flow, that the initial growth of the optimal mode greatly exceeds that of the most unstable normal mode, consistent with idealized flow analyses (e.g., [Farrell 1989a](#); [Mak and Cai 1989](#)). [Chang and Mak \(1993, 1995\)](#) studied intraseasonal disturbances and their relationship to the corresponding winter-mean flow through the application of a dynamical EOF analysis. Their results confirmed the greater role of nonmodal (optimal) growth for the winter intraseasonal disturbances compared to modal growth. Some improvements of the barotropic models employed, such as considering the barotropic relationship between potential vorticity and streamfunction, has also been explored (e.g., [Mitas and Robinson 2005](#)).

The goal of the current study is to identify the dynamic origin and growth of the large-scale disturbance associated with the secondary (non-ENSO or PDO related) pattern in U.S. dryness during summer, that is, the NW–SE oriented dipole pattern discovered by [Zhao et al. \(2017b\)](#).

Previous studies have thoroughly investigated the instability characteristics of extratropical circulation during the cool season (from both modal and nonmodal perspectives), but some unresolved issues remain regarding the summer circulation. For example, it is unclear where the precursor disturbance of the U.S. dryness-related large-scale meteorological pattern (especially on intra-seasonal time scales) arises and how long the disturbance takes to attain its maximum amplitude. Moreover, the external forcing responsible for the upstream precursor has not been thoroughly investigated. To address the above questions, in the current study we adopt a hierarchical analysis approach. The quasi-stationary wave train pattern associated with NW–SE oriented dipole will first be isolated and then contrasted with the results of a parallel NIA. The source location and external forcing of the wave train pattern will also be studied. The results of the current paper provide some essential guidelines for understanding the dynamic origin and forcing associated with the U.S. hydrological extremes in boreal summer.

2. Data and methodology

a. Data

The observed daily precipitation rate used in this study is from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) daily global and U.S. unified gauge-based analysis of precipitation at the resolution of 1.0° longitude by 1.0° latitude (Chen et al. 2008). This study also uses the daily mean geopotential height and zonal and meridional winds from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis at a resolution of 2.5° longitude by 2.5° latitude and with 17 pressure levels in the vertical ranging from 1000 to 10 hPa (Kalnay et al. 1996). Daily zonal and meridional winds are used to compute daily streamfunction on isobaric surfaces. The cloud ice and liquid water measures are taken from the NASA Modern-Era Retrospective Analysis for Research and Applications (MERRA) reanalysis with a resolution of 1.0° longitude by 1.0° latitude and 42 pressure levels in the vertical ranging from 1000 to 0.1 hPa (Rienecker et al. 2011). The analysis covers boreal summer (June–August) for the period 1950–2016. Because of the length limits of the data, the analyses of the daily global precipitation and cloud water are restricted to the periods 1979–2016 and 1981–2011, respectively.

b. The statistical analysis

Intraseasonal fluctuations in daily streamfunction at 250 hPa are obtained using a 10–90-day Butterworth

bandpass filter, minimizing the loss of data at the beginning and end points (Russell 2006). An EOF analysis via a covariance matrix is applied to the bandpass-filtered summer daily streamfunction (over the region 20° – 70° N, 0° – 360°) to identify major patterns of intra-seasonal variability in the circumglobal circulation in the Northern Hemisphere (NH) midlatitudes during boreal summer. Note that in the high-latitude grid points, the convergence of meridians results in a smaller area compared to the low latitude. To reduce the impact of the high-latitude grid points, we have applied area weight for the data before the EOF analysis. The square root of the cosine of latitude is used to compute the area weight (e.g., Baldwin and Dunkerton 2001; Hendon et al. 2007). Following Hartmann (2016), we apply Eq. (24) of North et al. (1982) to infer if each EOF mode identified is statistically distinct. It showed that the standard error in the estimation of the eigenvalues is approximately

$$\Delta\eta = \eta\sqrt{2/N^*}, \quad (1)$$

where η is the eigenvalue and N^* is the number of degrees of freedom. The upper and lower boundary of each eigenvalue is $\eta + \Delta\eta$ and $\eta - \Delta\eta$, respectively. If an EOF mode is statistically distinct, its eigenvalue should be smaller (larger) than the lower (upper) boundary of its prior (subsequent) mode.

Composite daily anomalies (unfiltered data; deviations from the long-term daily climatological values) in streamfunction, geopotential height, and precipitation are created based upon the time series of the principal components (PCs) obtained in the EOF analysis. Large-magnitude days for each EOF are defined in the following manner: if the corresponding EOF PC value on a given day is greater (smaller) than the 95th (5th) percentile of all PC values for this mode, we classify that day as a PC+ (PC−) event [note that the PC value for PC+ (PC−) event is positive (negative)]. Then, if two or more consecutive days are denoted as PC+ or PC− events, only the day with the largest PC magnitude [i.e., the largest (smallest) PC value for PC+ (PC−) event] will be reserved. For convenience, we denote the day of PC+ or PC− event as day 0, 1-day lag/lead as day −1/day +1, 2-day lag/lead as day −2/day +2, and so on.

Linear lag regression analysis is conducted by regressing a daily unfiltered field (e.g., streamfunction) at each grid point onto the daily PC time series (the former lags the latter by several days). In other words, we are looking for the best estimates of parameters E and F for the equation field ($t - \text{lag}$) = $E + F \times \text{PC}(t)$. The slope (i.e., F) of the lag regression analysis is plotted. The Student's t test is applied to examine the significance of regressed anomalies.

c. The modal instability

Following Eq. (13.36) of Holton and Hakim (2012), the unforced nondivergent barotropic vorticity equation in spherical coordinates may be written as

$$\frac{\partial \nabla^2 \psi}{\partial t} = \frac{1}{r^2 \cos \varphi} \left(\frac{\partial \psi}{\partial \varphi} \frac{\partial \nabla^2 \psi}{\partial \lambda} - \frac{\partial \psi}{\partial \lambda} \frac{\partial \nabla^2 \psi}{\partial \varphi} \right) - \frac{2\Omega}{r^2} \frac{\partial \psi}{\partial \lambda}, \quad (2)$$

where

$$\nabla^2 \psi = \frac{1}{r^2 \cos^2 \varphi} \frac{\partial^2 \psi}{\partial \lambda^2} - \frac{\tan \varphi}{r^2} \frac{\partial \psi}{\partial \varphi} + \frac{1}{r^2} \frac{\partial^2 \psi}{\partial \varphi^2}. \quad (3)$$

In (2) and (3), ψ is streamfunction; r is Earth's radius; φ and λ are the latitude and longitude, respectively; and Ω is the angular speed of Earth's rotation. To obtain the linearized version of (2), the flow is separated into mean ($\bar{}$) and perturbation ($\prime$), and products of perturbation fields are neglected,

$$\begin{aligned} \frac{\partial \nabla^2 \psi'}{\partial t} = \frac{1}{r^2 \cos \varphi} \left(\frac{\partial \bar{\psi}}{\partial \varphi} \frac{\partial \nabla^2 \psi'}{\partial \lambda} + \frac{\partial \psi'}{\partial \varphi} \frac{\partial \nabla^2 \bar{\psi}}{\partial \lambda} \right. \\ \left. - \frac{\partial \bar{\psi}}{\partial \lambda} \frac{\partial \nabla^2 \psi'}{\partial \varphi} - \frac{\partial \psi'}{\partial \lambda} \frac{\partial \nabla^2 \bar{\psi}}{\partial \varphi} \right) - \frac{2\Omega}{r^2} \frac{\partial \psi'}{\partial \lambda}. \end{aligned} \quad (4)$$

We assume solutions of the form

$$\psi'(\lambda, \varphi, t) = \phi(\lambda, \varphi) e^{I\sigma t}, \quad (5)$$

where $\phi(\lambda, \varphi)$ is the perturbation streamfunction amplitude for each grid point over the domain of 20°–70°N, 0°–360° (consisting of 144 and 21 grid points in the longitudinal and latitudinal direction, respectively); I is the imaginary unit; and σ is the frequency of oscillation. Substituting (5) into (4), casting the obtained equation in second-order center finite-difference form (since the region in this analysis is not on a global grid, we simply apply finite difference for discretization), and applying boundary conditions (cyclical boundary in the longitudinal direction is used), we obtain the matrix equation

$$\sigma \mathbf{L} \boldsymbol{\phi} = \mathbf{R} \boldsymbol{\phi}. \quad (6)$$

In (6), $\mathbf{L}$ and $\mathbf{R}$ are matrices ($N \times N$) and $\boldsymbol{\phi}$ is a vector ($N \times 1$), where N is the total grid points in the two-dimensional gridded field ($N = X \times Y$; $X = 144$, $Y = 21$). Following Mak (2002) and Mak (2011), we subsequently solve an eigenvalue–eigenvector problem with σ being an eigenvalue and $\boldsymbol{\phi}$ the corresponding eigenvector. A total of N pairs of eigenvalues and eigenvectors are generated. Each pair corresponds to a normal mode (i.e., $\boldsymbol{\phi} e^{I\sigma t}$).

d. The optimal (nonmodal) growth

We then make use of all of the resulting normal modes (N normal modes in total) obtained above as base functions to compute optimal modes over the optimization time τ , which means that initial disturbances would accumulatively intensify the *most* over the time interval τ in the presence of a basic flow (Mak 2011). The streamfunction perturbation at $t = \tau$ for all grid points of the domain (expressed as a vector) can be written as

$$\boldsymbol{\psi} = \sum_{n=1}^N \boldsymbol{\phi}_n e^{I\sigma_n \tau} a_n = \mathbf{P} \boldsymbol{\Lambda} \mathbf{a}. \quad (7)$$

In (7), for each value of n , $\boldsymbol{\phi}_n e^{I\sigma_n \tau}$ denotes a normal mode at $t = \tau$ (containing all grid points of the domain) and a_n is the projection coefficient associated with that normal mode; $\mathbf{P}$ is an $N \times N$ matrix, which is made up of $\{\boldsymbol{\phi}_n\}$; $\boldsymbol{\Lambda}$ is also an $N \times N$ matrix of which the diagonal elements are $\{e^{I\sigma_n \tau}\}$ and all others are zero; $\mathbf{a}$ is a vector with the elements of $\{a_n\}$. Following Mak (2011), the intensity of a disturbance at $t = \tau$ is

$$A(\tau) = \boldsymbol{\psi}^H \mathbf{D} \boldsymbol{\psi} = \mathbf{a}^H \boldsymbol{\Lambda}^H \mathbf{P}^H \mathbf{D} \mathbf{P} \boldsymbol{\Lambda} \mathbf{a} = \mathbf{a}^H \mathbf{B}(\tau) \mathbf{a}, \quad (8)$$

where $\boldsymbol{\psi}^H$ is the Hermitian of $\boldsymbol{\psi}$ and $\mathbf{D}$ is a unit matrix ($N \times N$). The intensity of an initial disturbance ($t = 0$) is

$$A(0) = \mathbf{a}^H \mathbf{P}^H \mathbf{D} \mathbf{P} \mathbf{a} = \mathbf{a}^H \mathbf{B}(0) \mathbf{a}. \quad (9)$$

Then the amplification factor of a disturbance is defined as

$$\gamma = \frac{\mathbf{a}^H \mathbf{B}(\tau) \mathbf{a}}{\mathbf{a}^H \mathbf{B}(0) \mathbf{a}}. \quad (10)$$

Equation (10) can be further reduced to an eigenvalue–eigenvector problem with γ being an eigenvalue and $\mathbf{a}$ the corresponding eigenvector,

$$\gamma \mathbf{B}(0) \mathbf{a} = \mathbf{B}(\tau) \mathbf{a}. \quad (11)$$

Similar to the modal growth, the optimal (nonmodal) growth will also generate N pairs of eigenvalues γ and eigenvectors $\mathbf{a}$ and, thus, N optimal modes for each optimization time τ . For a specified optimization time, the optimal mode with the largest amplification factor (i.e., the largest eigenvalue γ) is denoted as the first optimal mode. The remaining optimal modes are ordered in the sequence of the magnitude of amplification factor γ and are denoted as the second optimal mode, third optimal mode, and so on. Finally, using (7), the spatial structure of an optimal mode (corresponding to one pair of γ and $\mathbf{a}$) for the optimization time τ at a time t ($t \geq 0$) could be shown as

$$\psi'(\lambda, \varphi, t) = \text{Re} \left(\sum_{n=1}^N \phi_n e^{I\sigma_n t} a_n \right), \quad (12)$$

where Re denotes taking the real number. The corresponding initial disturbance ($t = 0$) for the above optimal mode is

$$\psi'(\lambda, \varphi, 0) = \text{Re} \left(\sum_{n=1}^N \phi_n a_n \right). \quad (13)$$

3. Results

a. The pattern of intraseasonal fluctuations

The intraseasonal modes in the upper-level streamfunction over the NH midlatitudes are identified through an EOF analysis (see [section 2b](#) for more details). The top three EOF modes are statistically distinct from one another (according to the application of the North test, that is, comparing the eigenvalue of one mode with the lower and upper boundary of its prior and subsequent mode, respectively) and account for, respectively, 13.3%, 4.9%, and 4.5% of the total variance ([Fig. 1](#)). The first EOF mode (EOF-1) exhibits a negative streamfunction anomaly pattern across the entire NH midlatitudes with strongest amplitudes over the northeastern Pacific–North American sector ([Fig. 2a](#)). [Figure 2c](#) displays the longitude–height cross section of the composite daily geopotential height anomalies (where the contours delineate statistical significance at the 0.01 level according to the Student's t test) at 30°N for the PC+ events (at day 0) for EOF-1. The center of negative streamfunction anomalies shown in [Fig. 2a](#) is strongest over the lower stratosphere (250–100 hPa). The single-signed streamfunction anomaly pattern bears some similarity to the CGT pattern (e.g., [Ding and Wang 2005](#); [Wang et al. 2012](#); [Teng et al. 2013](#)). The second EOF mode (EOF-2) is a dipole with an anomalously low (high) center over the Aleutian Islands (northeastern Pacific; [Fig. 2b](#)). The corresponding longitude–height cross section of the composite geopotential height anomalies at 50°N shows a westward tilt with height for the low pressure center over the Aleutian Islands, indicating a baroclinic structure in the vertical at this location ([Fig. 2d](#)). However, the vertical tilt of the high pressure center at around 40°N is much less pronounced than that of the low pressure ([Fig. 2e](#)).

[Figure 3a](#) shows the third EOF mode (EOF-3) of the intraseasonal variability in 250-hPa streamfunction over the NH during summer. It is characterized by a zonally oriented wave train extending from East Asia across the North Pacific and North America to the North Atlantic. This pattern bears a remarkable resemblance to the wave train pattern associated with the NW–SE dipole of

FIG. 1. The eigenvalue and upper and lower boundary of each eigenvalue for the first four EOF modes. The three lines for each mode denote the upper boundary, eigenvalue, and low boundary, respectively (from top to bottom).

U.S. dryness identified in [Zhao et al. \(2017b\)](#). [Schubert et al. \(2011\)](#) also captured a similar Rossby-wave-like pattern in their REOF analysis of summertime 250-hPa monthly mean meridional wind. [Figures 3b and 3c](#) show the composite 250-hPa streamfunction anomaly patterns for the PC+ and PC− EOF-3 events, respectively. A total of 79 (78) PC+ (PC−) event days are identified for the period 1950–2016. Verifying the EOF-3 structure, in both composites a wave train extends over the Pacific–North American sector, consisting of a longitudinal series of geographically fixed but oppositely signed ridge/trough features. The cross-sectional map at 45°N (the latitudinal center of the wave packet) shows that the EOF-3 wave train has an equivalent barotropic structure in the vertical, with little variation in the longitudinal location of geopotential height anomaly features with respect to height ([Figs. 3d,e](#)). Such equivalent barotropic structures are commonly found for large-scale meteorological patterns over extratropical regions (e.g., [Linkin and Nigam 2008](#); [Yang and Hoskins 2017](#)).

[Figures 4a and 4b](#) display the composite anomalies in the U.S. precipitation associated with the PC+ and PC− events for EOF-3. The composites show a NW–SE oriented anomaly dipole pattern with enhanced (diminished) and diminished (enhanced) precipitation, respectively, over the northwestern and southeastern United States during PC+ (PC−) events. [Figures 4c and 4d](#) show the U.S. summer mean precipitation anomalies for the two clusters (each cluster is a collection of U.S. summer mean precipitation anomalies in multiple years)

FIG. 2. The EOF modes of the 10–90-day filtered daily streamfunction at 250 hPa during summer for (a) EOF-1 and (b) EOF-2. (c) Composite geopotential height anomalies (shading; m) for the PC+ events of EOF-1 at day 0 from the surface to 100 hPa at 30°N. (d), (e) As in (c), but for the composite PC+ events of EOF-2 at 50° and 40°N, respectively. Contours are significant at the 0.01 level. The numbers shown in the top right corner of (a) and (b) and (c)–(e) are the variance of the EOF mode and the latitude, respectively.

identified in Zhao et al. (2017b). In their paper, cluster 1 occurred during 13 summer seasons, while cluster 2 was observed during 43 summers (their studying period is 1950–2005). It is clear that the composite precipitation anomalies associated with the PC+ and PC− events provide a strong qualitative resemblance to corresponding patterns of interannual variations in the summer U.S. precipitation, albeit with some difference in magnitude. To further examine the potential connection between the intraseasonal and interannual variability of U.S. summer precipitation, we also compute the empirical probability distribution function of the PCs of EOF-3 falling into the years of cluster 1 and cluster 2, respectively, during the period 1950–2005. The results again show that large-magnitude PC+ events occur more frequently during the years of cluster 1, while large-magnitude PC− events are more frequently seen during the years of cluster 2. These results confirm the linkage between intraseasonal and interannual variability of U.S. summer precipitation. Zhao et al. (2017b) suggested that the NW–SE dipole pattern in summer reflects the U.S. hydrological behavior during times of weak PDO or ENSO variability.

b. The optimal growth of the intraseasonal mode

Noting that the wave train pattern identified here represents an important mode of intraseasonal variability in

NH summer climate, we next investigate its dynamic origin and growth. Previous studies have demonstrated that, at least during winter, nonmodal growth plays a more important role than modal growth in generating intraseasonal disturbances (e.g., Chang and Mak 1993, 1995). Here we hypothesize that the emergence of EOF-3 is the result of one of the strongest nonmodal growths within a specific time interval τ in the NH summer climatological mean. We will test this hypothesis by studying the excitation of optimal modes through a nonmodal (optimal) instability analysis to understand how a weak precursor disturbance evolves into a disturbance resembling EOF-3.

Analogous to previous studies considering the winter climatological mean flow as a basic state (e.g., Simmons et al. 1983; Chang and Mak 1993, 1995), here we analyze the summer mean streamfunction $\bar{\psi}$ at 250 hPa. During NH summer (within 20°–70°N), the climatological upper-tropospheric streamfunction typically decreases from low to high latitudes (Fig. 5a). The modal (normal mode) and nonmodal (optimal mode) instability analysis will be performed for this given basic state (see sections 2c and 2d for details). To estimate the appropriate values for the optimization time τ (which is a very important parameter for NIA), we compute lag autocorrelations for the PC time series of the EOF-3. The results show that the correlation coefficient drops from 1 to 0

FIG. 3. (a) As in Fig. 2a, but for EOF-3. (b),(c) Composite streamfunction anomalies (shading; $\text{m}^2 \text{s}^{-1}$) at 250 hPa at day 0 for the PC+ and PC- events of EOF-3, respectively. (d),(e) As in (b),(c), but for composite geopotential height anomalies (shading; m) from the surface to 100 hPa at 45°N. Contours are significant at the 0.01 level. The numbers shown in the top right corner of (a) and (d),(e) are the variance of the EOF mode and the latitude, respectively.

at a lag of around seven days (Table 1). Therefore, we assess the optimal modes of the summer mean flow for optimization time τ ranging from 5 to 10 days (that are close to 7 days), which means that an initial disturbance in summer mean flow would likely intensify the most within 5–10 days. Such a parameterization of the optimization time for summer mean flow is generally consistent with those used in previous studies of winter flow (e.g., Borges and Hartmann 1992; Chang and Mak 1995; Mitas and Robinson 2005). Finally, N ($N = X \times Y$; $X = 144$, $Y = 21$) optimal modes are obtained for each optimization time τ (ranging from 5 to 10 days).

After obtaining all the optimal modes, we compare their structures to the EOF-3 pattern (both PC+ and PC- phase) to find those sharing the greatest similarity with EOF-3. For each optimization time τ , we assess the spatial correlation coefficient between the composite streamfunction anomalies at day 0 for the PC+ and PC- events (i.e., Figs. 3b,c) and all optimal modes at the time from $t = 0$ to $t = 10$ days [using (12) and setting t from 0 to 10]. The results reveal that the second optimal mode for the optimization time $\tau = 7$ ($\tau = 9$) days at $t = 9$ days is most similar to the composite structure of PC+ (PC-) events with a spatial pattern correlation of 0.67 (0.68). In addition, the spatial correlation coefficients for the second optimal mode (also at $t = 9$ days)

determined for $\tau = 5, 6, 8$, and 10 days are all close to 0.65, indicating a very small sensitivity to the precise choice of optimization time. In the following analysis, we will mainly focus on the second optimal mode for $\tau = 7$ and 9 days.

Figure 6a shows the time evolution map [$t = 0, 1, 3, 5, 7$, and 9 days for Eq. (12)] of the second optimal mode for the optimization time $\tau = 7$ days, meaning that this mode has the second largest amplification factor and intensifies the most at $t = 7$ days. Consistent with the result of the spatial correlation analysis, the structure of this mode at $t = 9$ days (i.e., Fig. 6a6) generally resembles the pattern of the composite streamfunction at day 0 for the PC+ events (i.e., Fig. 3b). The time evolution map reveals a quasi-stationary wave train structure; that is, downstream propagation of Rossby wave energy via a sequential zonal series of individual quasi-stationary troughs/ridges. Such a wave train pattern also resembles the Pacific–Japan teleconnection pattern captured in Li et al. (2014). Figure 6a1 shows that the initial disturbance (i.e., at $t = 0$ day) of this optimal mode first appears over East Asia (between 20° and 40°N) where the meridional gradient of the climatological zonal wind at 250 hPa has large positive values (i.e., $\partial \bar{u} / \partial y > 0$; Fig. 5b). We note the slight but distinct NW–SE tilt in the initial disturbance horizontal structure (i.e., $u'v' < 0$),

FIG. 4. (a),(b) As in Figs. 3b and 3c, but for composite daily precipitation anomalies (shading; mm day^{-1}). (c),(d) Composite of seasonal mean daily precipitation anomalies (shading; mm day^{-1}) for the two clusters during boreal summer identified in Zhao et al. (2017b). Contours are significant at the 0.1 level.

indicating that the latitudinal shear straining term ($-u'v'\partial\bar{u}/\partial y$) is positive and kinetic energy is being converted from the mean flow into the disturbance (Holton and Hakim 2012). This process dictates the growth of the initial disturbance, and the dominant role of the shear straining term in the optimal growth is consistent with the findings in Huang (1999). The optimal growth for an optimization time of $\tau = 9$ days leads to an oppositely signed structure (Fig. 6b). At the time $t = 9$ days, the optimal mode structure is spatially quite similar to the day 0 composite streamfunction pattern for the PC- events (Fig. 6b6 vs Fig. 3c). Compared to the second optimal mode for $\tau = 7$ days, the second optimal mode for $\tau = 9$ days exhibits a very similar quasi-stationary wave train pattern, but of opposite sign.

c. The external forcing for the initial disturbances

The results presented in the previous section illustrate that East Asia (within 20° – 40°N) is the dynamic origin for the third mode of intraseasonal variability in the NH boreal summer circulation. In this section, we go a step further to explore what might trigger the development of the initial disturbances over East Asia for the optimal modes discussed above (i.e., Figs. 6a1, 6b1). Lag regression analysis is applied to estimate the linear relationship between the EOF-3 PC time series and multiple fields (e.g., streamfunction and precipitation) that lag the PC time series by several days. Regressing

summer daily streamfunction anomalies at 250 hPa against the PC time series (the streamfunction lags the PC time series by 8, 6, 4, and 2 days), we obtain the associated regional evolution of the upper-tropospheric streamfunction anomalies from day -8 to day -2 for the PC+ events (where the contours delineate statistical significance at the 0.01 level according to the Student's t test; Fig. 7b). The spatial correlations between the second optimal mode for $\tau = 7$ days (Fig. 7a) and regressed streamfunction anomalies (Fig. 7b) are 0.25, 0.54, 0.73, and 0.83 from day -8 to day -2 . These maps obtained by lag regression analysis generally

FIG. 5. (a) Climatology of the streamfunction ($\text{m}^2 \text{s}^{-1}$) at 250 hPa for summer. (b) As in (a), but for zonal wind (contour; m s^{-1}) at 250 hPa and its meridional gradient (shading; s^{-1}).

TABLE 1. Lag autocorrelations for the PC time series of the EOF-3 with a lag from 0 to 8 days.

Lag (days)	0	1	2	3	4	5	6	7	8
Autocorrelation	1.00	0.95	0.82	0.65	0.44	0.25	0.07	−0.05	−0.14

mimic those of the second optimal mode for $\tau = 7$ days, especially after day -6 .

Figure 7b1 exhibits a meridional dipole structure with negative and positive streamfunction anomaly features over, respectively, the northern and southern part of East Asia at day -8 . This dipole structure bears some similarity to that at the time $t = 1$ day for the optimal mode (Fig. 7b1 vs Fig. 7a1), despite the circulation band in the latter pattern being much narrower than the former with a spatial shift in the center. The greater difference for the initial disturbance (around day -8) is likely the result of a high degree of idealization used in our analysis, including the use of a simplified unforced nondivergent barotropic model ignoring baroclinic processes that intrinsically affects the scale of the identified optimal disturbances. A better comparison of the initial disturbance between those found through NIA and lag regression analysis is expected if we switch to a more sophisticated model. We further investigate the vertical structure of the geopotential height anomaly at day -8 for the reference latitude of 30°N . Consistent with the horizontal distributions shown in Fig. 7b1, the

longitude–height cross-sectional map at 30°N also displays a ridge (trough) to the west (east; Fig. 8a).

The formation of ridge/trough patterns in the atmosphere could be associated with anomalous local diabatic heating processes (e.g., latent heat release; Rauber et al. 2005; Lu and Deng 2015). As we know, the latent heat release is directly related to water phase changes (from water vapor to liquid/ice water). Precipitation and changes in cloud water are two factors that lead to latent heat release. Figure 8b shows the vertical structure of the mass fraction anomalies in cloud liquid and ice water at day -8 for the latitude of 30°N obtained through the lag regression analysis. It is evident that positive mass fraction anomalies around $90^\circ\text{--}110^\circ\text{E}$ are most likely associated with enhanced latent heat release, which would warm up the layer and positively reinforce the upper-level ridge. Conversely, the negative mass fraction anomalies near $110^\circ\text{--}120^\circ\text{E}$ are very likely responsible for producing negative geopotential height anomalies over this region, albeit with varying degrees of consistency between the two fields around $120^\circ\text{--}130^\circ\text{E}$.

FIG. 6. (a) The time evolution (for $t = 0, 1, 3, 5, 7$, and 9 days) of the second optimal mode for the optimization time $\tau = 7$ days. (b) As in (a), but for the optimization time $\tau = 9$ days.

FIG. 7. (a) As in Fig. 6a, but for $t = 1, 3, 5$, and 7 days over East Asia. (b) Regression of daily streamfunction anomalies (shading; $\text{m}^2 \text{s}^{-1}$) at 250 hPa against the PC time series of EOF-3 (the streamfunction lags the PC time series by $8, 6, 4$, and 2 days). Contours are significant at the 0.01 level. The dashed black line in (b1) denotes the reference latitude of 30°N . The number shown in the parentheses is the spatial correlation coefficient between optimal mode and regressed anomalies.

Figure 8c shows the lag regression map for the vertically integrated (normalized by the thickness of the column) mass fraction anomalies of cloud liquid and ice water at day -8 . For the reference latitude of 30°N , the vertically integrated mass fraction anomalies are generally consistent with the results in Fig. 8b. Two bands with positive and negative mass fraction anomalies (significant at the 0.05 level), respectively, over the Yangtze River and Yellow River, can be observed in

Fig. 8c. The positive anomalies are responsible for the upper-level ridge shown in Fig. 8a, while the negative anomalies are associated with suppressed latent heat release and upper-level trough formation (see the negative band around 40°N in Fig. 7b1; the vertical structure of this trough is not shown). Parallel analyses of precipitation anomalies over land are generally consistent with those of cloud liquid and ice water (Fig. 8c vs Fig. 8d).

FIG. 8. (a) Regression of geopotential height anomalies (shading; m) from the surface to 100 hPa at 30°N against the PC time series of EOF-3 (the geopotential height lags the PC time series by 8 days). (b) As in (a), but for mass fraction anomalies (shading; kg kg^{-1}) of cloud liquid and ice water. (c) As in Fig. 7b1, but for mass fraction anomalies (shading; kg kg^{-1}) of cloud liquid and ice water, which is vertically integrated and normalized from the surface to 250 hPa. (d) As in (c), but for daily precipitation anomalies (shading; mm day^{-1}). Contours are significant at the 0.05 level. The gray shading is for topography. The dashed black lines denote the reference latitude of 30°N. The two solid blue lines (from the north to south) denote the Yellow River and Yangtze River, respectively.

In addition to local thermodynamic forcing (i.e., latent heating), dynamic processes (e.g., vorticity advection) can also cause a redistribution of mass in the atmosphere and create ridge/trough patterns (Chen 2010). For example, when a subtropical low-vorticity air mass is advected poleward, an atmospheric blocking occurs following the Rossby wave breaking at the exit of the storm track (Davini and D'Andrea 2016). Nonlinear eddy meridional vorticity transport may trigger intraseasonal oscillation perturbation in boreal summer over the midlatitude North Pacific (Wang et al. 2013). Following these studies, we examine the potential impact of remote dynamic forcing upon the formation of the upper-level ridge/trough over East Asia. A similar lag regression analysis is applied to multiple dynamical fields including absolute vorticity advection over East and Southeast Asia. No significant anomalous dynamical sources for the initial disturbance could be found within the surrounding region, confirming a dominant role for local thermodynamic forcing (i.e., diabatic heating processes) tied to monsoonal precipitation anomalies over East Asia in initiating the upper-level ridge/trough feature leading to the optimal growth of a disturbance resembling the EOF-3 mode of intraseasonal variability. This conclusion is generally consistent with previous

work of Zhu and Li (2016) conducted on interannual time scales.

4. Summary and concluding remarks

This study identifies the origin of one of the major intraseasonal modes of NH upper-tropospheric variability in boreal summer as described by variations in upper-tropospheric streamfunction. An EOF analysis is performed for daily 10–90-day bandpass-filtered 250-hPa streamfunction for the period of 1950–2016. EOF-1 is found to exhibit a negative streamfunction anomaly pattern across the entirety of NH midlatitudes, while EOF-2 is a dipole with anomalous low (high) values over the Aleutian Islands (northeastern Pacific). EOF-1 attains the largest magnitudes in the lower stratosphere, while EOF-2 exhibits a baroclinic vertical structure in the troposphere, especially for the center over the Aleutian Islands. EOF-3 consists of a quasi-stationary wave train pattern over the Pacific–North American sector that has an equivalent barotropic structure in the vertical. This pattern is linked to a NW–SE oriented dipole of precipitation anomalies over the United States in boreal summer. Both the wave train pattern and associated precipitation anomaly pattern closely resemble the

results of an earlier analysis of the interannual variability in the U.S. summer dryness (Zhao et al. 2017b).

The dynamic origin of the EOF-3 quasi-stationary wave train is explored in an unforced nondivergent barotropic model. A nonmodal (optimal growth) instability analysis of the climatological summer mean flow indicates that both positive and negative phases of EOF-3 may be triggered by ideally configured small-scale precursor disturbances over East Asia where the meridional gradient of zonal wind is very large. The initial disturbances include a NW–SE horizontal tilt that facilitates a kinetic energy transfer from the basic flow into the perturbation. Over a period of nine days, the precursor disturbances grow in magnitude as energy propagates downstream via North Pacific toward North America, forming the quasi-stationary wave train. A lag regression analysis based on the EOF-3 PCs reveals a dipole structure in upper-level streamfunction over East Asia that compares well to the precursor disturbance identified through the parallel instability analysis, indicating that upper-tropospheric variability over East Asia is likely the key leading to the development of the EOF-3 intraseasonal mode and associated downstream precipitation anomaly pattern over the United States in boreal summer. Lag regression analyses further reveal that phase changes of water linked to the formation of clouds and precipitation are responsible for local latent heating anomalies that likely trigger the development of the East Asian precursor disturbance for EOF-3. And these cloud and precipitation anomalies are directly associated with meridional shifts in the East Asian monsoonal rainfall in boreal summer.

The current work provides an effective method for understanding the dynamical pathway between nonlocal external forcing (i.e., diabatic heating related to East Asian rainfall anomalies) and excitation of modes of local hydroclimate variability (i.e., the dipole in the U.S. summer precipitation). The key difference compared to Zhu and Li (2016) is that we use NIA of the boreal summer background flow to demonstrate that the excitation of a cross-Pacific wave train with the initial disturbance coming from East Asia is intrinsically favored by the structure of the summer background flow. In other words, the emergence of such a wave train and the associated intraseasonal variability is not by chance, just like cyclones inevitably emerge from a baroclinically unstable background flow. This analysis framework also allows us to identify other important modes of variability preferentially occurring in boreal summer (i.e., through computing other “optimal modes”) and is therefore a more general approach of seeking the dynamical origin of large-scale disturbances of a specific class compared to the “initial value” approach taken by

Zhu and Li (2016) to examine the atmospheric response to a prescribed diabatic heating anomaly. Moreover, our analysis focuses on intraseasonal time scales, exhibiting a faster growth (around nine days) of the wave train from the initial disturbance over East Asia, and the structure of initial disturbance is obtained strictly from the summer background flow. Ongoing and future work includes a consideration of baroclinic and nonlinear processes involved in forming the EOF-3 mode and a more detailed examination of the implication of this result for the predictability of U.S. hydroclimate in boreal summer.

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