The Impact of Layer Perturbation Potential Energy on the East Asian Summer Monsoon

LIDOU HUYAN, a,b JIANPING LI, c,d SEN ZHAO, e,f CHENG SUN, c DI DONG, a,b TING LIU, g AND YUE Fei ZHAO h

a State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China
b University of Chinese Academy of Sciences, Beijing, China
c State Key Laboratory of Earth Surface Processes and Resource Ecology, and College of Global Change and Earth System Science, Beijing Normal University, Beijing, China
d Laboratory for Regional Oceanography and Numerical Modeling, Qingdao National Laboratory for Marine Science and Technology, Qingdao, China
e School of Ocean and Earth Sciences and Technology, University of Hawai‘i at Manoa, Honolulu, Hawaii
f Key Laboratory of Meteorological Disaster of Ministry of Education, and College of Atmospheric Science, Nanjing University of Information Science and Technology, Nanjing, China
g State Key Laboratory of Satellite Ocean Environment Dynamics, Second Institute of Oceanography, Hangzhou, China
h National Meteorological Information Center, Beijing, China

(Manuscript received 9 October 2016, in final form 21 April 2017)

ABSTRACT

This paper analyzes the relationship between the 1000–850-hPa layer perturbation potential energy (LPPE) as the difference in local potential energy between the actual state and the reference state and the East Asian summer monsoon (EASM) using reanalysis and observational datasets. The EASM is closely related to the first-order moment term of LPPE (LPPE1) from the preceding March to the boreal summer over three key regions: the eastern Indian Ocean, the subtropical central Pacific, and midlatitude East Asia. The LPPE1 pattern \((\cdot, +, +)\), with negative values over the eastern Indian Ocean, positive values over the subtropical central Pacific, and positive values over East Asia, corresponds to negative LPPE1 anomalies over the south of the EASM region but positive LPPE1 anomalies over the north of the EASM region, which lead to an anomalous downward branch over the southern region but an upward branch over the northern region. The anomalous vertical motion affects the local meridional circulation over East Asia that leads to a southwesterly wind anomaly over East Asia (south of 30\(^{\circ}\)N) at 850 hPa and anomalous downward motion over 100\(^{\circ}\)E–120\(^{\circ}\)E (along 25\(^{\circ}\)–35\(^{\circ}\)N), resulting in a stronger EASM, more kinetic energy over the EASM region, and less boreal summer rainfall in the middle and lower reaches of the Yangtze River valley (24\(^{\circ}\)–36\(^{\circ}\)N, 90\(^{\circ}\)–125\(^{\circ}\)E). These LPPE1 anomalies in the eastern Indian Ocean and subtropical central Pacific appear to be connected to changes in local sea surface temperature through the release of latent heat.

1. Introduction

The climate of China is greatly affected by the East Asian monsoon (Tao and Chen 1987). In particular, the flood season, large-scale distribution of rainfall patterns, rain belt movement, and extreme precipitation over East Asia are associated with the East Asian summer monsoon (EASM). Furthermore, drought and flood events related to the EASM often cause heavy economic losses and casualties. Understanding the formation and variation of the EASM will assist our understanding of the effects of climate change, revealing patterns of seasonal precipitation variability, and generating new theories and methods for climate prediction. Therefore, the EASM variability and its related climate anomalies have received considerable attention (Li et al. 2011a, b, 2013).

Numerous studies have investigated the EASM and have focused on, for example, the characteristics and variability of the EASM (Li et al. 2004; Qian 2005; He et al. 2006), interaction between the EASM and El Niño–Southern Oscillation (ENSO; Lau and Nath 2000; Wang et al. 2008b), the simulation and forecasting of the East Asian monsoon (Wang et al. 2004), the impact on the formation of ENSO (Li and Mu 1998), the subtropical East Asian monsoon (He et al. 2004; Zhu et al.)

Corresponding author: Prof. Jianping Li, ljp@bnu.edu.cn

DOI: 10.1175/JCLI-D-16-0729.1

© 2017 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the AMS Copyright Policy (www.ametsoc.org/PUBSReuseLicenses).
the impact of underlying processes on the EASM (Liang and Wu 2003; Zhang and Li 2004; Zuo and Zhang 2016), the decadal variability of the EASM (Yang et al. 2005), and the kinetic energy (KE) of the Asian monsoon system (Guan 2000). Although there has been much work and significant achievements in the field of EASM research, our current forecast skill with respect to the EASM is still not high and does not meet the needs of production and life (e.g., Kang et al. 2002; Wang et al. 2005; Song and Zhou 2014). This implies that our understanding of the dynamics and physical processes associated with the EASM and its climatic impacts needs to be improved, and that further exploration in this field will be of both great theoretical significance and practical value.

The ongoing evolution of the atmospheric circulation is driven by the changing nature of the atmospheric energy balance, and the variability of the EASM is related to the changes in the KE of the EASM circulation (e.g., van Mieghem 1973; Li et al. 2016). Previous studies have shown that the EASM is connected to the underlying external forcing, such as sea surface temperature (SST) and snow cover (Guo and Wang 1986; Wu et al. 2004, 2005; Zhang and Li 2004; Nan and Li 2005; Li et al. 2011a, b; Zheng et al. 2014). However, the theory of atmospheric energetics confirms that external diabatic heating cannot be directly converted into KE, which indicates that the variations of the underlying external forcing are not the direct sources of the KE (e.g., Margules 1903; Lorenz 1955). Therefore, it is necessary to explore the sources of KE within the EASM system. In addition, there have been few studies of the variability of the EASM from the point of view of regional energetics. Research from this new perspective is hoped to bring a better understanding of the variability of the EASM and its link to the external forcing.

The framework of modern atmospheric energetics is based on Lorenz’s study of available potential energy (APE; Lorenz 1955). Further development of the theory of APE has aided our understanding of the efficiency of energy conversion between APE and KE in terms of the global mean (Johnson 1970; Bullock and Johnson 1971; Gu 1990; Gao and Li 2007). However, the application of APE is restricted because the concept of APE is based on globally averaged variables, which imposes limitations on the regional energy conversion; consequently, modifications are required when applying APE in the local region. Li and Gao (2006) introduced the concept of perturbation potential energy (PPE), which indicates the maximum amount of total potential energy that could be converted into KE at the local scale. Locally, the first-order moment term of PPE (PPE1) is an order of magnitude larger than the second-order moment term of PPE (PPE2; Li and Gao 2006; Li et al. 2016).

According to Gao and Li (2012, 2013), PPE is closely related to the atmospheric circulation variability, and they also analyzed the characteristics of the surface perturbation potential energy. To investigate energy transformation features at different levels in the light of the baroclinic of atmosphere, Wang et al. (2012) introduced the concept of layer perturbation potential energy (LPPE) and also applied the theory to explore the variability of the South China Sea summer monsoon (Wang et al. 2013). The atmospheric PPE has been used to study variations in tropical Pacific atmospheric available energetics during an ENSO cycle (Dong et al. 2017). These previous results indicate that PPE is a key link between the local diabatic heating and the variability of KE for circulation. Therefore, the PPE theory can provide a new perspective for studying the energetics of the regional circulation of the EASM and a theoretical foundation from which to examine the sources and variability of the KE of the EASM (Li et al. 2016). We should also consider that there are complicated external forcings that exert effect on the variability of EASM, such as ENSO, the tropical Indian Ocean, the subtropical Pacific, and the Tibetan Plateau (e.g., Wu et al. 2004, 2005; Wang et al. 2008b; Ding et al. 2010; Li et al. 2011a, b; Zheng et al. 2014). The aim of this paper is to examine the bridging role of LPPE1 between the variability of the EASM and underlying external forcing over the key regions, and to explore the possible mechanisms that control this relationship.

The remainder of this manuscript is organized as follows. In section 2, we provide a brief review of the datasets and methods of analysis used in this work. Section 3 characterizes the relationship between LPPE1 and the EASM. The atmospheric circulation anomalies over East Asia associated with LPPE1 and possible physical mechanisms are discussed in section 4. The connection between LPPE1 and the local underlying external forcing is given in section 5. In the final section, we present a summary and discussion.

2. Data and methodology
a. Data

The atmospheric variables were obtained from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis. These data have a horizontal resolution of 2.5° × 2.5° and cover the period from 1948 to the present (Kalnay et al. 1996) and include temperature, winds, vertical velocity, surface pressure, and precipitable water content (PWC). The SST data were obtained from the

8

JOURNAL OF CLIMATE VOLUME 30

2.5

8
Extended Reconstructed SST, version 4 (ERSST.v4; Huang et al. 2015; B. Huang et al. 2016; Liu et al. 2015) that covers the period from 1854 to the present and has a horizontal resolution of 2° × 2°. Precipitation data were taken from the monthly surface precipitation dataset for China, version 2.0 (V2.0), compiled by the China Meteorological Administration (Zhao et al. 2014; Zhao and Zhu 2015), which is gridded at a resolution of 0.5° × 0.5° and covers the period from 1961 to 2013. The station rainfall data (available online at http://bcc.nccma.net/channel.php?channelId=106) observed by 160 stations across China were also used. For atmospheric reanalysis data and SST data, we focused our analysis on monthly data for the period 1948–2014. Here, we define the boreal summer as June–August (JJA).

b. Methodology

The EASM index (EASMI) adopted in this work is derived from the unified dynamic normalized seasonality (DNS) monsoon index (Li and Zeng 2002, 2003). It can better describe the variability of the EASM strength compared to the other EASM indices in aspect of capturing of the leading modes of the East Asian summer precipitation (Li and Zeng 2005; Wang et al. 2008a). Further description of the DNS index is given in appendix A. The EASMI is defined as the areal averaged boreal summer mean DNS index over the region 10°–40°N, 110°–140°E; that is,

\[ \text{EASMI} = \{\delta\}_{(10°–40°N, 110°–140°E)} \]

(1)

where \{\}P denotes the areal average of the δ values over the domain D (here D is 10°–40°N, 110°–140°E). See Li and Zeng (2000, 2002, 2003, 2005), Feng et al. (2010), Li et al. (2010), and Zheng et al. (2014) about more details on the physical definition of the DNS index and the EASMI.

To provide a theoretical foundation for this work, the basic concept of PPE and the governing equations for PPE1 and KE are reproduced here from the previous studies (Li and Gao 2006; Wang et al. 2012, 2015; Li et al. 2016). The mathematical formula for LPPE is expressed as follows:

\[
P'_{l1} = \frac{1}{\gamma_d} \int_{p_1}^{p_2} T' \, dp \quad \text{and} \quad \gamma_d = g/c_p, \]

(2)

where \( p_0 \) is the reference pressure (usually taken to be 1000 hPa); \( p_1 \) and \( p_2 \) are the lower and upper limits of the vertical integration over the pressure range, respectively; and \( p \) is pressure. Also, \( g/c_p \) is the dry adiabatic lapse rate, where \( g \) is the gravitational acceleration, and \( c_p \) is the specific heat at constant pressure; and \( \kappa = R/c_p \), where \( R \) is the gas constant for dry air. In addition, \( T \) is temperature, \( T' \) is the departure from this global average, \( \theta \) is potential temperature, and \( \theta \) is a global average on the isobaric surface. The mathematical expressions of the first- and second-order moment terms of LPPE are

\[
P'_{l1} = \frac{1}{\gamma_d} \int_{p_1}^{p_2} T' \, dp \quad \text{and} \quad \gamma_d = g/c_p. \]

(3)

\[
P'_{l2} = \frac{\kappa p_0}{2\gamma_d} \int_{p_1}^{p_2} \frac{T'^2}{p^{1+\kappa}} \left( -\frac{\partial \theta}{\partial p} \right) \, dp. \]

(4)

It is obvious that the values of \( P'_{l1} \) may be positive or negative, whereas the values of \( P'_{l2} \) are always positive. Note that, at the regional scale, \( P'_{l1} \) generally exceeds \( P'_{l2} \), and can represent the energy conversion efficiency at the certain height of atmospheric column locally. A comparison with the APE (Lorenz 1955) shows that the values of the global mean of \( P'_{l2} \) are equal to the values of APE, although their physical meanings are different. PPE and APE represent the available potential energy at the local and global scales, respectively. We focus on the first-order moment term of LPPE (LPPE1) in this work.

The governing equations of LPPE1 and KE are briefly introduced below:

\[
\frac{\partial}{\partial t} \text{LPPE1} = \mathcal{S}_L + \mathcal{N}_L + R_L - C_K + G_L \quad \text{and} \quad (5)
\]

\[
\frac{\partial}{\partial t} \text{KE} = B_K + \mathcal{R}_K + C_K + M_K, \quad (6)
\]

where the kinetic energy \( KE = g^{-1} \int_{p_1}^{p_2} e_k \, dp \). Also, \( C_K \) represents the conversion term between LPPE1 and KE, which represents the local energy conversion efficiency and closely linked to the physical process of the impact of LPPE1 on the local atmospheric circulation, and \( G_L \) represents the source (sink) term of LPPE1, which depends on diabatic heating and boundary terms from the atmospheric radiation forcing, latent heat, underlying heat forcing, and boundary forcing. A more complete description of PPE theory and the related governing equation is given in appendix B.

The statistical methods used in this study were correlation, singular value decomposition (SVD; Wallace et al. 1992), and composite analysis. For a given index, high- and low-index cases were identified as the fluctuations of the index beyond one standard deviation in this study. The composite difference refers to the difference in the corresponding elements between the high- and low-index cases. The 9-yr low-pass and high-pass components of a variable were obtained using a Gaussian filter. The significance of the correlation between two
low-pass time series was accessed using the effective number of degrees of freedom (Bretherton et al. 1999).

3. Relationship between LPPE1 and the EASM

Figure 1 presents the correlation coefficients between the EASMI and 1000–850-hPa LPPE1. To find the significant PPE signals that exert stable effect on the variability of the EASM, we analyze the relationship between LPPE1 from the preceding spring (March–May) to the boreal summer and the EASMI. There are significant negative values in the tropical Indian Ocean, but significant positive values over the subtropical central Pacific and midlatitude East Asia (Figs. 1a–d). According to previous studies, the EASM is a complex atmospheric system that is subject to the combined influences of tropical, subtropical, and mid-to-high-latitude systems (e.g., Huang et al. 1999; Ding et al. 2010; Yun et al. 2010; Oh and Ha 2016). LPPE1 is significantly correlated with the EASMI over the three key regions: the eastern Indian Ocean (10°S–10°N, 72.5°–110°E), subtropical central Pacific (10°–20°N, 175°E–160°W), and midlatitude East Asia (30°–40°N, 115°–140°E), and these regions were selected to further explore the relationship between LPPE1 and the EASM. The time series of LPPE1 averaged over the eastern Indian Ocean, subtropical central Pacific, and East Asia regions are denoted as $I_{EIO}$, $I_{SCP}$, and $I_{EA}$, respectively. The correlation coefficients between the EASMI and $I_{EIO}$ in March, April, May, and the boreal summer were −0.39, −0.37, −0.38, and −0.37, respectively, which are all above the 99% confidence level based on the Student’s $t$ test; this indicates that LPPE1 over the eastern Indian Ocean region is most significantly connected to the variability of the EASM among the three key regions. To investigate the connection between LPPE1 over the key regions with the variability of the EASM, an LPPE1 index covering the LPPE1 signal over three key regions denoted as LPPEIIPA is used, the expression of which is

$$LPPEI_{IPA} = \frac{1}{4} (I_{SCP} + I_{EA}) - \frac{1}{2} I_{EIO}. \quad (7)$$

The time series of the EASMI and LPPEIIPA during the boreal summer are positively correlated (significant at the 99.9% confidence level; Fig. 2). We further analyzed the sliding correlations between the EASMI and LPPEIIPA during the boreal summer with a 41-yr moving window (not shown). The results shown that correlation coefficients between the EASMI and LPPEIIPA are stable and mostly can be significant at the 99.9% confidence level. The results confirm the stable and significant relationship between the variability of the EASM and LPPEIIPA.
There is a stable relationship between the interannual components of the EASMI and LPPEI\textsubscript{IPA} from March ensuing the boreal summer, and their decadal parts are also significantly correlated (Table 1), showing that the variability of LPPEI\textsubscript{IPA} and the EASM strength are connected on both the interannual and decadal time scales. These results indicate that it is reasonable to employ the PPE theory to analyze the atmospheric energetics associated with the EASM variability.

Figure 3 displays the correlation distributions between LPPEI\textsubscript{IPA} and boreal summer rainfall over eastern China using the gridded precipitation dataset. There are negative values (the stippled areas) in the middle and lower reaches of the Yangtze River valley (Figs. 3a–d), which are consistent with the relationship between the EASMI and boreal summer rainfall (Fig. 3e). To ensure the robustness of the results, we also used the China 160-station precipitation dataset. The significant negative correlations between LPPEI\textsubscript{IPA} from preceding March to the boreal summer and boreal summer precipitation in the middle and lower reaches of the Yangtze River valley (not shown) coincide with the results presented in Fig. 3. These results further confirm that the anomalies of LPPE1 over the eastern Indian Ocean, subtropical central Pacific, and East Asia regions are closely connected with the variability of the EASM.

To highlight the connection between LPPE1 over the key regions and boreal summer rainfall in the middle and lower reaches of the Yangtze River valley, we define a rainfall index that is the time series of boreal summer rainfall averaged over the significantly correlated area 28.25°–31.25°N, 106.75°–118.25°E denoted as R\textsubscript{IYR}. Figure 4 shows the reversed relationship between R\textsubscript{IYR} and LPPEI\textsubscript{IPA} during the boreal summer, with a correlation coefficient of −0.56 based on the gridded dataset; for the station dataset, the value is −0.52 (both significant at the 99.9% confidence level). This also demonstrates the consistency of the two rainfall datasets used here. In the following discussion, the R\textsubscript{IYR} time series is based on the boreal summer gridded precipitation dataset.

### 4. Connections between LPPEI over the key regions and the EASM

#### a. Boreal summer atmospheric circulation anomalies over East Asia associated with LPPEI\textsubscript{IPA}

We used composite analysis to examine the close connection between the EASM and LPPEI over the three key regions and explore the boreal summer atmospheric circulation anomalies associated with LPPEI\textsubscript{IPA}. To examine the stable influence of LPPEI\textsubscript{IPA} on the EASM, we also analyze the circulation anomalies corresponding to the preceding LPPEI\textsubscript{IPA}; here we mainly give the composite differences of atmospheric circulation patterns associated with LPPEI\textsubscript{IPA} in May and boreal summer. Figure 5 displays the composite differences of the boreal summer horizontal wind field at 850 hPa between the high- and low-LPPEI\textsubscript{IPA} years. The results show that there is an anomalous cyclonic circulation and southwesterly wind over East Asia (south of 30°N), and an anomalous westerly wind over 10°–20°N, implying stronger convection anomalies over the monsoon trough associated with the high-LPPEI\textsubscript{IPA} values in both May (Fig. 5a) and the boreal summer (Fig. 5b). Figure 6 presents the composite differences of the boreal summer vertical motion along 25°–35°N between the high- and low-LPPEI\textsubscript{IPA} years. There are downward motion anomalies within 100°–120°E, indicating that there is weaker convective activity over the mei-yu front, and anomalous downward motion in the middle and lower reaches of the Yangtze River valley in the high-LPPEI\textsubscript{IPA} case (Figs. 6a,b).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>LPPEI\textsubscript{IPA} in JJA</td>
<td>0.52 (0.64, 0.52)</td>
</tr>
<tr>
<td>LPPEI\textsubscript{IPA} in May</td>
<td>0.41 (0.61, 0.39)</td>
</tr>
<tr>
<td>LPPEI\textsubscript{IPA} in April</td>
<td>0.52 (0.73, 0.48)</td>
</tr>
<tr>
<td>LPPEI\textsubscript{IPA} in March</td>
<td>0.47 (0.71, 0.39)</td>
</tr>
</tbody>
</table>

#### Table 1. Correlation coefficients between the LPPEI\textsubscript{IPA} and EASMI for the period 1948–2014. The values in parentheses represent the 9-yr low-pass and high-pass time series, respectively, which may represent the decadal and interannual components of a variable. The correlation coefficients significant at the 99% confidence level using the Student’s $t$ test are in boldface. The significance of the correlation between two low-pass time series was assessed using the effective number of degrees of freedom (Bretherton et al. 1999).
EASM and reduced boreal summer rainfall in the middle and lower reaches of the Yangtze River valley (Zhang and Tao 1998; Zhang et al. 2003).

To confirm the results above, we also analyzed the composite differences of the boreal summer horizontal wind field at 850 hPa and vertical motion between the high- and low-RIYR EASMI years. The anomalous anticyclonic circulation over the East Asia–western North Pacific (south of 30° N) region, easterly wind over 10°–20° N (Fig. 5c), and the upward branch over 100°–120° E (Fig. 6c) in the high-RIYR case support the inverse relationship between the variability of boreal summer rainfall in the middle and lower reaches of the Yangtze River valley and LPPEIPA. The circulation anomalies associated with the high-LPPEIPA case above match the anomalous cyclonic circulation and southwesterly wind over the East Asia–western North Pacific region (south of 30°N; Fig. 5d) and anomalous downward branch over 100°–120°E (Fig. 6d) in the intensified EASM case, and further indicate the significant positive relationship between the variability of the EASM and LPPEIPA. The background circulation patterns associated with LPPEIPA, RIYR, and EASMI support the robust connection between LPPE1 over these key regions and the EASM, which further suggests that the atmospheric circulation anomalies associated with the high-LPPEIPA case are favorable to the strengthened EASM.

b. Possible physical mechanisms

If we are to improve our understanding of the mechanisms that drive the impact of LPPE1 over the key regions on the atmospheric circulation anomalies associated with the variability of the EASM, we need to identify the conversion relationship between LPPE1 and KE over the local region. Figure 7 shows the leading
SVD mode of boreal summer LPPE1 (Figs. 7a,c) and KE (Figs. 7b,d), which explains 45% of the total covariance. The two time series are strongly correlated, with a coefficient of 0.93 (significant at the 99.9% confidence level). The LPPE1 field is characterized by a positive phase in the tropical Indian Ocean and South China Sea, and a negative phase in the subtropical central Pacific and midlatitude East Asia. The KE field has a negative phase in the tropical Indian Ocean, central Pacific, and East Asia. With the high correlation coefficient of the time series of LPPE1 and KE fields, when there are negative LPPE1 anomalies in the tropical Indian Ocean and positive LPPE1 anomalies in the subtropical central Pacific and midlatitude East Asia, then positive KE anomalies appear over the East Asia region. These results suggest the coupled relationship between LPPE1 over the key regions and KE over the EASM region (10°–40°N, 110°–140°E), and also show that there are positive KE anomalies over the EASM region when there is a negative–positive–positive LPPE1 pattern (−, +, +) over the eastern Indian Ocean, subtropical central Pacific, and East Asia regions (i.e., during the positive LPPE1IPA case).

In addition, the significant negative correlation over the tropical Indian Ocean and midlatitude East Asia in the LPPE1 field reflects the out-of-phase variations in LPPE1 anomalies over the tropical and middle latitude regions. The meridional out-of-phase pattern of LPPE1 may favor the meridional circulation adjustment in the local area. Figure 8 shows the composite differences of meridional circulation along 110°–140°E between the high- and low-LPPE1IPA cases in May (Fig. 8a) and the boreal summer (Fig. 8b). There is anomalous upward motion over the north of the EASM region (40°–50°N) and anomalous downward motion over the south of the EASM region (0°–10°N), which implies that there is anomalous convergence over 40°–50°N and divergence over 0°–10°N. The anomalous vertical movement results in anomalous southerlies at low levels, which then turn right, under the influence of the Coriolis force, and this can lead to strengthened southwesterly winds at low levels and positive KE anomalies over the EASM region. These KE anomalies within the EASM system are associated with a stronger EASM and anomalous southwesterlies match the horizontal wind anomalies over East Asia (south of 30°N) at 850 hPa in Figs. 5a,b. The anomalous downward branch around 30°N associated with the high-LPPE1IPA case is consistent with the downward anomaly over 100°–120°E in Figs. 6a,b. The above results indicate that the local meridional circulation adjustment can influence the local circulation anomalies over East Asia.
We also present the composite differences of the meridional circulation between the high- and low-RIYR EASMI years. This shows an anomalous upward motion over 0°–10°N and downward motion over 40°–50°N in the high-RIYR case (Fig. 8c), which is the opposite pattern to the high-LPPEIIPA case (Fig. 8a). The anomalous downward (0°–10°N) and upward (40°–50°N) branches associated with the high-LPPEIIPA case are consistent with vertical movement anomalies during the intensified EASM period (Fig. 8d). In addition, there is an anomalous downward branch with high LPPEIIPA, EASMI, and anomalous upward motion around 30°N with the high RIYR. These results further prove the close connection between LPPEI over the eastern Indian Ocean, subtropical central Pacific, and East Asia regions and the variability of the EASM.

To verify the physical mechanisms that explain the impact of the out-of-phase LPPE1 pattern over the tropical Indian Ocean and midlatitude East Asia on the local meridional circulation from the point of view of energetics, we investigated the composite anomalies of the boreal summer LPPE1 (regional zonal mean over 110°–140°E) based on the high- and low-LPPEIIPA years. When LPPEIIPA is in its high phase, there are negative LPPE1 anomalies over 0°–10°N and positive LPPE1 anomalies over 40°–50°N (Fig. 9a). According to PPE theory, the variability of LPPE1 can represent the changes of the perturbation component of air temperature. That is, when there are negative LPPE1 anomalies over 0°–10°N and positive LPPE1 anomalies over 40°–50°N, the perturbation air is colder (warmer) over 0°–10°N (40°–50°N). Therefore, there is anomalous upward branch associated with the warmer air and anomalous downward branch with the colder air, LPPE1 converts to KE, and the vertical motion anomalies and local energy conversion can then induce the anomalous meridional circulation as presented in Fig. 8a. The situation for the negative LPPEIIPA case is reversed (Fig. 9b), indicating that there is an anomalous upward branch accompanying the warmer perturbation air over 0°–10°N and an anomalous downward branch with the colder perturbation air over 40°–50°N, which can lead to the development of the local meridional circulation pattern that is the opposite with those in the high-LPPEIIPA case.

To further explore the coupled relationship between the boreal summer LPPE1 and KE over the EASM region, we calculated the areal averaged LPPE1 and KE...
over the EASM region based on the high- and low-LPPEIIPA years. The area-averaged KE anomalies were significantly positive (negative) in the high- (low-) LPPEIIPA case, which were consistent with the positive (negative) KE anomalies over the EASM region associated with the high (low) LPPEIIPA and stronger (weaker) EASM mentioned above. There were more (fewer) area-averaged LPPE1 anomalies in the high- (low-) LPPEIIPA case. The tendencies of the area-averaged LPPE1 and KE over the EASM region in the boreal summer (not shown) are positively correlated (significant at the 99% confidence level), which shows that when the variations of LPPE1 are positive over the EASM region, the local KE is more likely to increase and can indicate the positive relationship between the variability of the area-averaged LPPE1 and KE over the EASM region. Thereby, in the low-LPPEIIPA case, there are fewer area-averaged LPPE1 anomalies compared with those in the high-LPPEIIPA case, which indicates that there is less energy over the EASM region that could be transformed into the local KE. In addition, because of the consistent relationship between the variations of LPPE1 and KE over the EASM region, the less LPPE1 anomalies can lead to reduced KE anomalies during the low-LPPEIIPA case, which explains why there are more (fewer) KE anomalies over the EASM region during the high- (low-) LPPEIIPA case.

The results above suggest that the LPPE1 anomalies over the key regions can cause atmospheric circulation anomalies over East Asia via the meridional circulation adjustment during the boreal summer, which is induced by the out-of-phase variations in LPPE1 anomalies over the north and south of the EASM regions based on the PPE theory. The anomalous circulation patterns favor a stronger (weaker) EASM and less (more) boreal summer rainfall in the middle and lower reaches of the Yangtze River valley. The generation of LPPE1 anomalies over the key regions needs further investigation. LPPE1 is closely related to the underlying external forcing, the variability of which may be responsible for the changes of LPPE1.

5. LPPE1 and the underlying external forcing

According to the derivation of the governing equation for PPE1 (see details in appendix B), \( G_L = g^{-1} \int p_1 \nabla^2 \) is the source or sink term of LPPE1. Here, the perturbation diabatic heating \( Q \) is the deviation from the spherical average of the primitive diabatic heating \( Q \), which indicates the local variation of the diabatic heating. Besides, according to Li and Gao (2006), PPE1 represents the local energy conversion efficiency. Therefore, here we applied the perturbation components of the variables to investigate the contributions of underlying forcing over the key regions on the changes of LPPE1.

Figure 10 presents the heterogeneous correlation patterns of the leading SVD mode for the boreal summer LPPE1 (Figs. 10a,c) and perturbation SST (Figs. 10b,d) in...
May (Figs. 10a,b) and the boreal summer (Figs. 10c,d). The SST pattern of the leading mode is significantly correlated with the LPPE1 of the leading mode, with a correlation coefficient of 0.86 for SST in boreal summer, and for SST in May the value is 0.74 (both significant at the 99.9% confidence level). When there are warmer perturbation SSTs in the local region, positive LPPE1 anomalies appear in this area. The high correlation coefficients of the two time series indicate that LPPE1 increases (decreases) accompany warmer (colder) perturbation SST in the local region. The coupled relationship between LPPE1 and perturbation SST over the key regions indicates that the changes in the regional SST make a positive contribution to the LPPE1 anomalies, and this explains the role of local SST variability in the generation of LPPE1 over the key regions.

To provide a better understanding of the influence of the local SST on the variability of LPPE1 over the key regions, we focused on the local latent heat to analyze the connection between the underlying diabatic heating and

![Fig. 8. Composite differences of the boreal summer meridional circulation [vectors; meridional wind (m s$^{-1}$) and vertical velocity ($-10^{-2}$ Pa s$^{-1}$)] between the high and low LPPE1 in (a) May and (b) the boreal summer, and between the high and low (c) RIYR and (d) EASMI years. The light and dark shaded areas indicate anomalies significant at the 90% and 95% confidence levels, respectively, and the red (blue) shading corresponds to upward (downward) motion. The latitude–pressure cross sections are along 110°–140°E, and the $y$ axis is pressure (hPa).](image)

![Fig. 9. Meridional distributions of the composite of the boreal summer LPPE1 (regional zonal mean over 110°–140°E; 10^6 J m$^{-2}$) based on the (a) high- and (b) low-LPPE1 cases.](image)
the changes of LPPE1 over the key regions. As there are both ocean and land signals in the East Asia region, here we concentrate on the relationship between the variability of LPPE1 and local latent heat in the eastern Indian Ocean and subtropical central Pacific. Moreover, the surface latent heat flux may not represent the total latent heat release, so here we use the PWC to depict the latent heating released by the condensation process. The $I_{SCP}$ is highly related to the area-averaged perturbation PWC over the subtropical central Pacific at a confidence level greater than 99% (Fig. 11a). It can be seen that the $I_{EIO}$ and area-averaged perturbation PWC over the eastern Indian Ocean are also significantly correlated (significant at the 95% confidence level) (Fig. 11b). The significant positive relationships between the perturbation PWC and LPPE1 in the eastern Indian Ocean and subtropical central Pacific indicate that the increases (decreases) of latent heat release in the eastern Indian Ocean and subtropical central Pacific lead to heating (cooling) of the air column, resulting in more (fewer) LPPE1 anomalies in the eastern Indian Ocean and subtropical central Pacific. The positive correlations also match the positive contribution of perturbation SST to the variability of LPPE1, which highlights the connection between the changes in local SST and latent heat release. These results suggest that the changes of local SST affect LPPE1 in the eastern Indian Ocean and subtropical central Pacific by heating (cooling) the atmosphere through local latent heat release.

According to Eqs. (5) and (6), we can see that the underlying diabatic heating is contained in the source term $G_L$ as in the right-hand side of Eq. (5). The conversion term $C_K$ in both equations reflects the conversion between LPPE1 and KE in the local scale. This indicates that the contribution of external heating on the local KE of circulation has to be accomplished through the changes of LPPE1, and LPPE1 is the part that can be directly transformed into the local KE around the 

![Fig. 10. Heterogeneous correlation patterns of the leading SVD mode for the (left) LPPE1 and (right) perturbation SST in (a),(b) May and (c),(d) boreal summer. The contour interval is 0.2. The light and dark shaded areas indicate significant values at the 90% and 95% confidence levels, respectively. The percentages in red are the explained variance of the leading mode.](image)

![Fig. 11. Normalized time series of the perturbation PWC over the (a) subtropical central Pacific and (b) eastern Indian Ocean in May (red lines) and the boreal summer (blue lines), and $I_{SCP}$ in (a) and $I_{EIO}$ in (b) in the boreal summer (black lines) for the period 1948–2014. The correlation coefficients between the $I_{SCP}$ and PWC over the subtropical central Pacific are 0.33 and 0.73 in May and boreal summer, respectively, and the correlation coefficients between the $I_{EIO}$ and PWC over the eastern Indian Ocean are 0.30 and 0.42 in boreal summer and May, respectively.](image)
EASM system from the underlying external heating over the key regions. Then the variations of atmospheric circulation anomalies over East Asia are correspondences to the variations of the driving energy around the EASM system, and thus influence the strength of the EASM. Based on the above results, the close relationship between SST in the Indian Ocean and the EASM (Nan and Li 2005; Li et al. 2011a; G. Huang et al. 2016) can be explained through the PPE theory as follows: colder SST in the Indian Ocean basin can induce decreases of LPPE1, which means that there are negative LPPE1 anomalies in the eastern Indian Ocean region (i.e., the positive LPPE1IPA phase) and this favors a stronger EASM and reduced boreal summer rainfall in the middle and lower reaches of the Yangtze River valley. This suggests that LPPE1 in the eastern Indian Ocean may play a bridging role for the effects of the local SST on the variability of the EASM.

6. Summary and discussion

In this article, we have demonstrated that the 1000–850-hPa LPPE1 over three key regions (the eastern Indian Ocean, subtropical central Pacific, and East Asia) has a robust connection with the EASM variability from March to the boreal summer. We found that the positive (negative) LPPE1IPA case—that is, the −, +, + (−, −, −) LPPE1 pattern over the key regions—can cause negative (positive) LPPE1 anomalies over the south of the EASM region (0°–10°N) and positive (negative) LPPE1 anomalies over the north of the EASM region (40°–50°N). Based on the local energy conversion and the PPE theory, we conclude that the out-of-phase pattern of LPPE1 anomalies can lead to an anomalous downward (upward) branch over 0°–10°N, and an anomalous upward (downward) branch over 40°–50°N. The anomalous vertical movement induces the local meridional circulation adjustment over East Asia. The local meridional circulation anomalies then cause an anomalous cyclonic (anticyclonic) circulation, a southwesterly (northeasterly) wind over East Asia (south of 30°N) at 850 hPa, and an anomalous downward (upward) branch over 100°–120°E (along 25°–35°N) in the boreal summer, that can strengthen (weaken) the EASM intensity, enhancing (reducing) KE over the EASM region, and resulting in less (more) boreal summer rainfall in the middle and lower reaches of the Yangtze River valley. We conclude that a warmer (colder) local SST can induce more (less) LPPE1 in the eastern Indian Ocean and subtropical central Pacific by heating (cooling) the atmosphere via the increases (decreases) of local latent heat release. We mainly discussed the close connection between LPPE1 over the three key regions and the EASM variability in view of energetics. The results present that the related physical processes are as follows: the changes of LPPE1 over the key regions cause the variability of local KE within the EASM system based on the local energy conversion. The meridional circulation adjustment over East Asia responds to the local energy conversion and anomalous energy patterns. The local meridional circulation anomalies then induce anomalous atmospheric circulation and vertical motion over East Asia, leading to the strengthened or weakened EASM.

Our results suggest that the LPPE1 signal over the key areas could explain the physical processes and possible mechanisms related to the underlying external forcing transforming into KE of the EASM system, but could also provide significant information that would help to better predict the variability of the EASM. The physical mechanisms and processes that explain the influence of the changes of the underlying external forcing on LPPE1 over the key regions require further exploration.

We should also consider the interaction between the ocean and atmosphere, and the feedback between the heat fluxes and the changes in SST. It can be seen that there are significantly negative correlations over the tropical Atlantic Ocean (Figs. 1a–d). We added LPPE1 over the Atlantic Ocean into the LPPE1IPA and obtained a new LPPE1 index. The correlation coefficient between the EASMI and this new index has been slightly increased in comparison to that with the LPPE1IPA. The composite differences of atmospheric circulation between high and low cases of the new index (not shown) have not been significantly strengthened compared with the anomalous circulation patterns associated with the LPPE1IPA case (Figs. 5a,b and 6a,b). This indicates that the explained variation of the EASM has not been highly improved with LPPE1 over the Atlantic Ocean included, which also validates the significant contributions of the three key regions on the EASM variability we mostly discussed. Additionally, the possible mechanisms that give rise to the anomalous LPPE1 pattern over the eastern Indian Ocean, subtropical Pacific, and East Asia regions remain unsolved. We also found that LPPE1 over the key regions and the EASM are both correlated on the interannual and decadal time scales. We mostly investigated the relationship between the EASM and LPPE1 over the key regions on the interannual time scale, and more details of the physical processes on the decadal time scale need to be investigated in future work.

In this paper we mainly focused on summer rainfall over eastern China; more research is needed to further understand the intraseasonal modes of the EASM rainfall (e.g., Oh and Ha 2015, 2016; Yim et al. 2015;
Moreover, we mainly analyzed the boreal summer mean component of the EASM variability from the energetic view. It is also suggested that the subseasonality is one of the most predominant aspects of the interannual and decadal variability of the East Asian climate (e.g., Ha et al. 2009; Yun et al. 2010). Understanding the energy conversion within the EASM system on the subseasonal time scale would aid to better reveal the mechanisms on the EASM variability responding to the external conditions. All of these important issues will require further research.

Acknowledgments. This work was jointly supported by the National Natural Science Foundation of China (NSFC) Project 41530424 and SOA International Cooperation Program on Global Change and Air–Sea Interactions (Grant GASI-IPOVAI-03). We sincerely thank four anonymous reviewers for their constructive comments and suggestions that helped to improve our manuscript.

APPENDIX A

Dynamic Normalized Seasonality Index

The unified DNS monsoon index defined by Li and Zeng (2002, 2003) is based on the intensity of wind field seasonality and can be applied to describe both the seasonal cycle and interannual variability of monsoons over different domains. The DNS index is given by

$$\delta_{m,n} = \frac{\| \bar{V}_1 - \bar{V}_{m,n} \|}{\| \bar{V} \|} - 2,$$

where $\bar{V}_1$ is the January climatology wind vector, $\bar{V}$ is the average of the January and July climatological wind vectors, and $\bar{V}_{m,n}$ denotes the monthly wind vector in the $m$th month of the $n$th year; note that 2 is subtracted on the right-hand side of Eq. (A1) because the critical value of significant $\| \bar{V}_1 - \bar{V}_{m,n} \|/\| \bar{V} \|$ is 2 (Li and Zeng 2000).

APPENDIX B


a. Derivation of perturbation potential energy

The total potential energy (TPE; Margules 1903) of the air column per unit area using the isentropic coordinate system $(\lambda, \phi, \theta, t)$ can be written as

$$P = \frac{1}{(1 + \kappa) \gamma_d P_0} \int_{\theta_1}^{\theta_5} \bar{p}^{1+\kappa} d\theta + \frac{1}{(1 + \kappa) \gamma_d P_0} \theta_s \bar{p}_s^{1+\kappa} + z_s \bar{p}_s,$$

where $z_s = z_s(\lambda, \phi)$ is the surface topography; $\lambda$ and $\phi$ are the longitude and latitude, respectively; $\bar{p}$ is pressure; $\bar{p}_s = p_s(\lambda, \phi)$ is surface pressure; $g$ is the gravitational acceleration; $c_p = c_v + R$ is the specific heat at constant pressure, where $R$ is the gas constant for dry air, and $c_v$ is the specific heat at constant volume; and $\gamma_d = g/c_p$ is the dry adiabatic lapse rate. Also, $T$ is temperature; $\theta = T(\rho_0/p)^{\kappa}$ is potential temperature and $\kappa = R/c_v$; $\theta_s$ and $\theta_t$ are potential temperatures at $z_s$ and the top of tropopause, respectively; and $P_0$ is the reference pressure (usually taken as 1000 hPa). The concept of APE (Lorenz 1955) is based on the mass conservation and on the redistribution of the atmosphere into the reference state through isentropic processes.

The difference of the TPE between the actual state and the reference state through adiabatic redistribution is

$$P' = P - \bar{P} = P_a' + P_s',$$

where the tilde indicates physical variables in the reference state, $P_a$ is the portion related to the air column, and $P_s$ is the portion related to the surface topography; $P_a' = P_a - \bar{P}_a$, $P_s' = P_s - \bar{P}_s$, and $P'$ is made of two parts: the first part is related to the air column and is called the atmospheric perturbation potential energy (shortened to PPE), and the second part is related to topography and is called the surface perturbation potential energy (SPPE). Gao et al. (2006) showed that the potential temperatures at the boundary are invariant ($\bar{\theta}_t = \theta_t$, $\bar{\theta}_s = \theta_s$) through isentropic processes under the condition of conservation of mass. We obtain the expressions for the PPE and SPPE:

$$P_a' = \frac{1}{(1 + \kappa) \gamma_d P_0} \int_{\theta_1}^{\theta_5} (p^{1+\kappa} - \bar{p}^{1+\kappa}) d\theta$$

and

$$P_s' = \frac{1}{(1 + \kappa) \gamma_d P_0} \theta_s (p_s^{1+\kappa} - \bar{p}_s^{1+\kappa}) + z_s (p_s - \bar{p}_s).$$

Because the atmosphere is integrated as a whole, its adiabatic redistribution can only be accomplished on the global rather than the local scale. Therefore, the above-mentioned reference state is defined as the spherical averaged on isentropic surfaces $\bar{p} = \bar{p}(\theta)$.
We derive the formula in isobaric coordinates to facilitate practical computation. Only the atmospheric PPE is considered here. It is convenient to partition the total pressure field into two parts: standard pressure, which is the spherically averaged pressure on isentropic surfaces, and pressure perturbations from the standard values, so \( p = P + p' \). Then Eq. (B3) can be written as

\[
P_a' = \sum_{i=1}^{\infty} P_{ai}' = \frac{1}{\gamma_d P_{\alpha_0}} \int_{\theta_i}^{\theta_{0+}} \frac{p'}{P} \right[ \frac{p'}{P} + \frac{\kappa}{2} \left( \frac{p'}{P} \right)^2 + \frac{\kappa(\kappa - 1)}{3!} \left( \frac{p'}{P} \right)^3 + \ldots \right] d\theta,
\]

where \( P_{ai}' \) (for \( i = 1, 2, \ldots \)) depends on the power of perturbation pressure \( p'' \), and is called the ordinal-order moment term of PPE. According to Lorenz (1955), the perturbation pressure \( p' \) on an isentropic surface and the perturbation potential temperature \( \theta' \) on an isobaric surface can be approximated \( p'(\theta) \approx -\theta' \frac{dP}{\theta} \).

The expression of ordinal-order moment term of PPE in isobaric coordinate system can be written as

\[
P_{ai}' = \frac{P_{\alpha_0}^{(i-1)}}{\gamma_d P_{\alpha_0}^{(i+1)}} \sum_{j=0}^{i-1} \int_{\theta_j}^{\theta_{i-1}} \left( \frac{p'}{P} \right)^{i-j} \frac{d\theta}{\theta}.
\]

Making use of the relation between the perturbation potential temperature and perturbation temperature, \( \theta' = \theta'' \) at \( \theta' = T'' \) and \( \theta = T' \); then Eq. (B6) can be given by

\[
P_{ai}' = \frac{P_{\alpha_0}^{(i-1)}}{\gamma_d P_{\alpha_0}^{(i+1)}} \sum_{j=0}^{i-1} \left( \frac{p'}{P} \right)^{i-j} \frac{d\theta}{\theta}.
\]

The PPE1 and PPE2 are expressed as

\[
P_{a1}' = \frac{1}{\gamma_d} \int_{\theta_0}^{\theta'} T' dp \quad \text{and} \quad P_{a2}' = \frac{\kappa P_{\alpha_0}^{(i-1)}}{2\gamma_d} \int_{\theta_j}^{\theta_{i-1}} \left( \frac{p'}{P} \right)^{i-j} \frac{d\theta}{\theta}.
\]

Actually, high-order moment terms of PPE are much smaller compared with PPE1 and PPE2 and can be omitted.

b. Derivation of governing equations for perturbation potential energy and kinetic energy

Wang et al. (2012) introduced the governing equations for PPE1 and KE.

Using the isobaric spherical coordinate system \((\lambda, \phi, \theta, t)\), the fundamental equations for the atmosphere (Peixoto and Oort 1992; see their chapter 14) can be written as

\[
\frac{du}{dt} - \frac{1}{a} u \nu \tan \phi = - \frac{1}{a \cos \phi} \frac{\partial \phi}{\partial \lambda} + f v + \gamma_a,
\]

\[
\frac{dv}{dt} + \frac{1}{a} v \nu \tan \phi = - \frac{1}{a} \frac{\partial \phi}{\partial \phi} - f u + \gamma_a,
\]

where \( d/dt = \partial/\partial t + (a \cos \phi)^{-1} u \partial/\partial \lambda + a^{-1} \nu \partial/\partial \phi + \omega \partial/\partial \theta \), \( a \) is the mean radius of Earth, \( \mathbf{V}_h = u \mathbf{i} + v \mathbf{j} \) is the horizontal velocity vector, \( \omega = dp/dt \) is the vertical velocity, \( D_{h} \mathbf{B} = (a \cos \phi)^{-1} [\partial B_i/\partial \lambda + \partial (B_i \cos \phi)/\partial \phi] + \partial B_i/\partial \phi \) is the divergence operator, \( \phi \) is the gravitational potential, \( \gamma_1 = \gamma_i + \gamma_\phi \mathbf{j} \) is the horizontal friction force, \( \alpha = \rho \) is specific volume, \( R \) is the gas constant of dry air, \( c_p \) is the specific heat at constant pressure, \( T \) is temperature, \( Q \) is diabatic heating, \( f = 2 \Omega \sin \phi \) is the acceleration by the Coriolis force, \( \Omega \) is the angular velocity of Earth’s rotation, and \( \mathbf{V}_h = (a \cos \phi)^{-1} \partial/\partial \lambda \mathbf{i} + a^{-1} \partial/\partial \phi \mathbf{j} \) is the horizontal gradient operator.

We use \((\lambda) = (4 \pi a) \int_{- \pi/2}^{\pi/2} \cos \phi \, d\phi \) to define the spherical average. The primitive variable \( T \) is \((\lambda) + (\phi) \), where the prime designates the deviation from the spherical average. Assuming that the physical variable consists of spherical-averaged and perturbation components: \( F(\lambda, \phi, p, t) = \bar{T}(p, t) + F'(\lambda, \phi, p, t) \), we then have

\[
\frac{1}{a \cos \phi} \frac{\partial \bar{T}}{\partial \phi} = 0, \quad \frac{1}{a} \frac{\partial \bar{T}}{\partial \lambda} = 0, \quad \frac{1}{a \cos \phi} \frac{\partial \bar{F}}{\partial \phi} = 0,
\]

\[
\frac{1}{a} \frac{\partial \bar{F}}{\partial \lambda} = F'(\phi, \lambda, t) - \bar{T}(p, t), \quad \frac{1}{a \cos \phi} \frac{\partial \bar{F}}{\partial \phi} = -\frac{1}{a} \bar{T} \tan \phi, \quad \frac{1}{a} \frac{\partial \bar{F}}{\partial \phi} = 0.
\]

(B15)

Applying the separation \( F(\lambda, \phi, p, t) = \bar{T}(p, t) + F'(\lambda, \phi, p, t) \) to the Eq. (B14), using identity Eq. (B15), we obtain the perturbation part of the thermodynamic equation:
\[
\frac{dc_p T'}{dt} = -\frac{\partial \phi'}{\partial p} + \left( \frac{\partial \phi'}{\partial p} - \frac{\partial \sigma}{\partial p} \right) + Q' + \left( \frac{\partial c_p T}{\partial p} - \frac{\partial c_p T}{\partial p} \right).
\]

The KE equation derived from the Eqs. (B10) and (B11) can then be written as

\[
\frac{\partial e_k}{\partial t} + D_z(V_h e_k) + \frac{\partial \omega e_k}{\partial p} = -D_z(V_h \phi') - \frac{\partial \omega \phi'}{\partial p} + \omega \frac{\partial \phi'}{\partial p} + V_h \gamma_h, \tag{B17}
\]

where \(e_k = 0.5(u^2 + v^2)\) and \(A'_i = c_p T'\).

Integrating Eqs. (B16) and (B17) vertically over the pressure, the governing equations of PPE1 and KE can be given by

\[
\frac{\partial}{\partial t} \text{PPE1} = \frac{3}{4} L + \mathcal{N}_L + R_L - C_K + G_L \quad \text{and} \tag{B18}
\]

\[
\frac{\partial}{\partial t} \text{KE} = B_K + \mathcal{R}_K + C_K + M_K, \tag{B19}
\]

where \(\text{PPE1} = g^{-1} \int_0^p c_p T' dp\) \cite{Li2006}, \(\text{KE} = g^{1} \int_0^p e_k dp\) is kinetic energy, and \(C_K = g^{1} \int_0^p \omega(\partial \phi' / \partial p) dp\) is the conversion term between PPE1 and KE, which depends on vertical velocity and atmospheric stability. When \(C_K\) is positive with warm (cold) air ascending (descending), then PPE1 converts to KE; when \(C_K\) is negative with warm (cold) air descending (ascending), KE converts to PPE1. Also, \(G_L = -g^{-1} \int_0^p Q' dp\) is the source (sink) term of PPE1; when it gives net heating, PPE1 will increase, whereas with net cooling, PPE1 will decrease. Uniform heating increases the total energy but not the PPE1. The horizontal divergence term is \(\mathcal{N}_L = -g^{-1} D_z \int_0^p V_h A'_i dp\), and \(\mathcal{N}_t = -g^{-1}(g\alpha_1 + c_p T') dp\) is related to surface vertical motion. Furthermore, \(R_L = -g^{-1} \int_0^p [\omega' \partial(\phi' T') / \partial p + \omega \partial(\phi' T') / \partial p - \omega \partial \phi' / \partial p] dp\) is related to the vertical redistribution of temperature; \(M_K = g^{-1} \int_0^p V_h \gamma h dp\) is related to viscous dissipation, with \(V_h\) the horizontal velocity vector and \(\gamma h\) the horizontal friction force; \(B_K = g^{-1} D_z \int_0^p V_h (\phi' + e_k) dp\) is related to transport and work by horizontal boundary pressure; and \(\mathcal{R}_K = -g^{-1} \omega(e_k + \phi') dp\) is related to topographic effects.

REFERENCES


