The Interannual Variability of Summer Upper-Tropospheric Temperature over East Asia

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ABSTRACT

By using 55-yr NCEP–NCAR reanalysis data, two dominant interannual variability modes of summer upper-tropospheric (500–200 hPa) temperature over East Asia are identified. The first empirical orthogonal function (EOF1) mode in its positive sign features a monopole cooling anomaly, while the second mode (EOF2) features a meridional dipole mode, with the positive (negative) center located south (north) of 35°N. The EOF1 (EOF2) mode is associated with ENSO developing (decaying) summers. They are the result of dynamical teleconnections remotely induced by ENSO and local moist processes. During the El Niño developing summer, the Indian summer monsoon precipitation decreases and forces the Silk Road teleconnection pattern at 200 hPa, featuring an anomalous cyclone over the East Asian continent. Coupled with the anomalous northerly wind in eastern China at 850 hPa, rainfall over north (south) China is suppressed (enhanced). The anomalous cyclone in the upper troposphere, associated vertical motion, and precipitation contribute to the heat and vorticity balance and maintain the monopole cooling. In the El Niño decaying summer, driven by the combined effects of a local SST anomaly and remote warm SST anomaly forcing from the Indian Ocean, precipitation is reduced over the western Pacific Ocean. Less latent heat is released and forces the Pacific–Japan teleconnection pattern along the East Asian continent, inducing a tripolar rainfall anomaly over East Asia. The tripolar precipitation and vertical motion anomalies and the zonal extended cyclonic anomaly in the upper troposphere provide the heating and momentum flux balance and maintain the temperature anomaly pattern during the ENSO decaying summer.

1. Introduction

The East Asian summer monsoon (EASM) is an important component of the global monsoon system and plays an important role in global climate variability. The distinctive topography of East Asia produces unique features of the EASM (Ding 1994; Webster et al. 1998; Wang et al. 2001). The summer monsoon results from the differential heating between Eurasian landmass and the adjacent oceans (Webster et al. 1998). Multiscale variability of the EASM is greatly affected by the changes of land–sea thermal contrast. Changes of land–sea thermal contrast have been used to explain the interdecadal variability of the EASM (Ding et al. 2008; Li et al. 2010) and to measure its predictability (Zhou and Zou 2010). Understanding the mechanism of tropospheric temperature change is crucial to the prediction of land–sea thermal contrast and thereby the monsoon variability.

The interdecadal variability of tropospheric temperature over East Asia has been well documented. For example, during the period from 1950 to 2000, contrary to the warming trend elsewhere, a strong tropospheric cooling trend was prominent over East Asia during July and August that contributes to the tendency toward increased droughts in northern China and more floods along the Yangtze River valley (Yu et al. 2004; Yu and Zhou 2007). Further analysis reveals that the tropospheric cooling over East Asia is a regional manifestation of an interdecadal variability mode of tropospheric temperature change across the entire subtropical Northern Hemisphere, which exhibits a significant cooling center over East Asia and two warming centers over the North Atlantic and North Pacific Oceans (Zhou and Zhang 2009). The interdecadal upper-tropospheric cooling is also evident in spring (Yu and Zhou 2007), which is associated with an anomalous meridional cell with descending motions in the latitudes (26°–35°N) and low-level northerly winds over southeastern China (22°–30°N, 110°–125°E),
leading to deficient rainfall over South China (Xin et al. 2006).

In contrast to sufficient studies of interdecadal variability, less effort has been devoted to the interannual variability of tropospheric temperature over East Asia. There have been many studies on the interannual variations of tropical tropospheric temperature (TT) associated with ENSO. As a dynamical response to equatorial diabatic heating, the warm phase of ENSO is characterized by an overall warming of tropical TT superimposed upon a distinctive equatorially symmetric dumbbell-shaped pattern straddling the equator (Yulaeva and Wallace 1994). Recent studies have identified different modes associated with two flavors of El Niño in terms of the three-dimensional structure of tropical atmospheric temperature. The first is a deep-warm mode that features a coherent zonal mean warming throughout the troposphere from 30°N to 30°S with cooling aloft. The second is a shallow-warm mode that features strong wave signatures in the troposphere with warmth (coolness) over the central Pacific (western Pacific) (Trenberth and Smith 2006, 2009). Both the deep-warm mode and the shallow-warm mode can be reasonably reproduced by atmospheric general circulation models forced by historical sea surface temperature (SST), demonstrating that the two modes are driven by tropical SST variability (Zhou and Zhang 2011). There are also studies on the extratropical TT changes that highlight the eddy-driven meridional circulation changes (Seager et al. 2003). However, up to now no special attention has been paid to the East Asian domain. The present study aims to answer the following questions. 1) Are there any distinct interannual variability modes of summer upper-tropospheric temperature (UTT) over East Asia? 2) What are the thermal and dynamic processes that contribute to the interannual variations of summer UTT over East Asia? 3) Whether and how does ENSO affect the interannual variability modes of East Asian summer UTT?

The remainder of the paper is organized as follows: Section 2 describes the data and analyses method. In section 3, principal interannual variability modes of the summer UTT over East Asia are presented. The thermal and dynamic processes associated with the first two leading modes are examined in section 4. The relationship between ENSO and summer UTT changes over East Asia and the potential mechanisms are discussed in section 5. Summary and discussion are provided in section 6.

2. Data and method description

a. Data description

The datasets used in the present study include

1) National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data from 1951 to 2005 (Kalnay et al. 1996),
2) the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) dataset from 1958 to 2002 (Uppala et al. 2005),
3) Japanese 25-yr Reanalysis (JRA-25) conducted by the Japan Meteorological Agency from 1979 to 2005 (Onogi et al. 2007),
4) monthly global SST for the period of 1950–2005 taken from the Hadley Centre Sea Ice and Sea Surface Temperature dataset version1 (HadISST1), provided by the Met Office Hadley Centre for Climate Prediction and Research (Rayner et al. 2003), and
5) the precipitation reconstruction data compiled by the Climate Prediction Center at the National Centers for Environmental Prediction (PREC) from 1948 to 2005 (Chen et al. 2002).

In addition, the monthly mean Niño-3.4 index, defined as SST anomalies averaged within the region 5°S–5°N, 120°–170°W is also used.

In addition to monthly mean fields, 6-hourly horizontal and vertical velocity and temperature data from the reanalysis datasets are also used to calculate the diabatic heating and transient eddy heating.

b. Methodology

To assess the relative roles of the thermodynamic processes, we diagnose the temperature heat budget and the vorticity equation of the East Asian summer upper troposphere. Following Yanai and Li (1994), the temperature equation is written as

$$\frac{\partial T}{\partial t} = -\nabla \cdot \mathbf{u} - \frac{\partial T}{\partial y} - \frac{(p/p_0)RCP}{\partial \theta} \frac{\partial \theta}{\partial \theta} - \frac{(p/p_0)RCP}{\partial \theta} \left[ \mathbf{v} \cdot \nabla \frac{\partial \theta}{\partial \theta} + \frac{\partial (\omega^2 \theta)}{\partial \theta} \right] + Q_1. \quad (1)$$

The vorticity equation can be written as

$$\frac{\partial \zeta}{\partial t} = - \nabla \cdot \mathbf{v} - \beta \mathbf{v} - (f + \xi) \cdot \nabla \mathbf{v} + \left[ \left( \frac{u^2}{\partial \xi} \frac{\partial \xi^2}{\partial y} + v^2 \frac{\partial \xi}{\partial y} \right) + \xi^2 \right] \cdot \nabla \mathbf{v} + \text{res}, \quad (2)$$

where $u$, $v$, $\omega$, $T$, $\xi$, and $\theta$ are three-dimensional (3D) velocities, temperature, relative vorticity, and potential temperature; $R$ and $C_p$ are the gas constant and the specific heat at constant pressure of dry air; $p_0 = 1000$ hPa; $f$ is the Coriolis parameter; $\beta = df/dy$ is a constant; and $Q_1$ is the diabatic heating. The overbar denotes the monthly average and the double prime represents the departure of 6-h reanalysis data from the monthly average.
The rhs terms in Eq. (1) are usually termed as zonal temperature advection, meridional temperature advection, adiabatic heating, transient eddy heating, and diabatic heating, respectively (Yanai and Li 1994; Seager et al. 2003; Tamura et al. 2010). The adiabatic heating includes two physical processes: one is the vertical advection associated with vertical motion and another is that the air expands and cools as it rises or contracts and grows warmer as it descends. The rhs terms in Eq. (2) are the horizontal vorticity tendency derived from horizontal vorticity advection, beta term, divergence, transient eddy vorticity flux, and residual term (res). The res term contains dissipation, external forcing, vertical advection, and the twisting terms.

The 3D diabatic heating $Q_1$ is diagnosed as a residual in the thermodynamic equation (e.g., Hoskins et al. 1989; Nigam 1994):

$$Q_1 = \frac{\partial T}{\partial t} + \nabla \cdot \nabla T + \left( \frac{p}{p_0} \right) R C_P \frac{\partial \theta}{\partial p}$$

$$+ \left( \frac{p}{p_0} \right) R C_P \left[ \nabla \cdot \nabla \theta + \frac{\partial (\omega \cdot \theta \phi)}{\partial p} \right].$$

(3)

To focus on year-by-year variations, the long-term trend and decadal variations with a period longer than 9 yr are removed by using Fourier harmonic analysis of the seasonal mean anomalies. We have done analyses by using both NCEP-NCAR and ERA-40 reanalyses. The large-scale interannual variation modes derived from the two reanalysis datasets are highly consistent for their common period from 1958 to 2002, adding fidelity to the robustness of the results. For brevity, we only present the results derived from NCEP-NCAR reanalysis, which has a longer time period.

3. Two major interannual variability modes of summer UTT over East Asia

Owing to the large topography over East Asia, the so-called surface air temperature actually measures the temperature at different altitudes. In comparison with the surface air temperature, the upper-tropospheric (200–500-hPa mean) temperature can well indicate the combined effect of zonal thermal contrasts and meridional thermal contrasts between East Asia and the ambient ocean on East Asian summer monsoon changes. Thus, it is generally to measure the tropospheric mean temperature by using 200–500-hPa average (Zhao et al. 2007; Zhou and Zou 2010). In our analysis we focus on the 200–500-hPa mean temperature. The standard deviations of summer (June–August) mean UTT derived from three reanalysis datasets (i.e., NCEP, ERA-40, and JRA-25) are shown in Figs. 1a–c. Variability is most prominent over the extratropics with two centers located in west-central Asia (30°–50°N, 30°–80°E) and East Asia (20°–60°N, 90°–150°E). Results derived from the three reanalysis datasets are highly consistent. We focus on the variability over East Asia in our following analysis.

To reveal the dominant interannual variability modes of summer UTT over East Asia, an empirical orthogonal function (EOF) analysis of the temporal correlation matrix is performed on summer mean UTT over East Asia (20°–60°N, 90°–150°E). The first two leading modes and corresponding principal components are shown in Fig. 2. The first EOF mode in its positive sign features a monopole pattern centered near 30°–50°N, 100°–140°E. The second EOF mode in the positive sign features a meridional dipole mode, with the positive (negative) center located south (north) of 35°N.
The percentage variance accounted for by the first six eigenvalues of EOF analysis is shown in Fig. 3a. The unit standard deviation of the sampling errors associated with each percentage eigenvalue is also shown in Fig. 3a. The EOF1 (EOF2) accounts for 40.9% (19.7%) of total variance. According to the rule of North et al. (1982), the first two leading modes are well distinguished from each other in terms of the sampling error bars and hence are statistically significant. Note that the two leading modes explain a large part (i.e., 60.6%) of total variance.

To reveal the dominant time periods of the two leading modes, power spectral analysis is done on the corresponding principal component (PC) time series shown in Figs. 2c,d. The power spectrum density distributions and corresponding red noise of the first and second PC time series are shown in Figs. 3b,c. The first mode has a major spectral peak at 3.5 yr. The second mode has a low-frequency spectral peak around 3–5 yr and a secondary peak around 2–3 yr. The dominant time periods indicate that the strong interannual variability modes may be related to ENSO: this is further confirmed by the evidences to be presented in the following section.

To reveal the vertical structure of temperature anomalies associated with these two leading modes, the latitude–height cross sections of regression coefficients between the PCs and June–August (JJA) mean temperature anomalies averaged between 90° and 150°E are shown in Fig. 4. For the first leading mode significant negative regression coefficients are evident over the whole of East Asia (Fig. 4a) with the minimum center at 45°N near 300–250 hPa. For the second leading mode (Fig. 4b) significant positive (negative) regression coefficients exist south (north) of 32°N centered at 25°N (45°N) near 250 hPa. Signals of the two leading EOF modes are thus vertically robust beneath all of the upper troposphere.

To verify reliability of the modes we have also performed a regression analysis as Dommenget and Latif (2002). After removing the signal of EOF1, the regression analysis reveals an out-of-phase pattern between Southeast Asia and northeast Asia as for EOF2 shown above. We also changed the domain from 90°–150°E to 0°–180°. The temperature anomalies over East Asia associated with the first two leading modes remain unchanged (figures not shown), demonstrating that the leading modes of the upper-tropospheric temperature over East Asia are independent of the boundaries.

4. Thermal and dynamic processes associated with the first two leading modes

a. The anomalous circulation associated with the two leading modes

The composite winds at 200 hPa, vertical motion at 500 hPa, and precipitation anomalies are shown in Fig. 5.
The threshold for composite analysis is done based on the normalized PC time series, which are larger (less) than 0.8 (−0.8), as shown in Table 1. The climatological mean circulation shows that in boreal summer the upper troposphere over the Asian continent is dominated by a planetary-scale warm air mass centered on the southern Tibetan Plateau (Figs. 5a,b), which is the well-known South Asian high and is accompanied with anticyclonic circulation (Zhou and Li 2002). A jet stream is zonally oriented over East Asia (Figs. 5c,d), which is the so-called East Asian subtropical westerly jet stream (Zhang et al. 2006).

Associated with EOF1, deficient and excessive rainfalls are seen over northern East Asia and from South China to the south of Japan, respectively (Fig. 5a). At 200 hPa there is a cyclonic anomaly over East Asia, associated with a northerly (southerly) anomaly on its western (eastern) flank. It significantly suppresses the vertical motion over northern East Asia and enhances vertical motion over South China (Fig. 5c), inducing rainfall anomalies as shown in Fig. 5a.

For EOF2, a meridional tripolar pattern along the East Asian continent is seen in the rainfall, with deficient precipitation in South China (20°–25°N) and North China, and excessive precipitation along the mei-yu–changma–baiyu rainband (Fig. 5b). Compared with EOF1, the excessive rainfall anomalies shift northward. The anomalous vertical motions are consistent with precipitation anomalies. In the upper troposphere an anomalous zonal wind along 30°–40°N at 200 hPa dominates East Asia and accompanies a northerly wind along the East Asian continent. However, a southerly wind in the eastern part of East Asia is not obvious.

**Fig. 3.** (a) The total variance (%) explained by each EOF for their respective interannual time series. Error bars are determined by the formula of North et al. (1982). The power spectrum density (solid line) and red noise (dashed line) of (b) the first and (c) the second EOF principal component; X axis in (b) and (c) is the period (yr).

**Fig. 4.** Regression coefficient maps between the principal components and zonal mean (90°–150°E) summer temperature anomalies (K/std) as latitude–pressure cross section: (a) PC1 and (b) PC2. Contour intervals in (a) and (b) are 0.05; shading denotes the 90% confidence level using a Student’s t test.
To understand how the temperature anomalies sustain and reveal the corresponding thermal and dynamical structures, the heat and vorticity balance associated with the first two leading EOF modes are quantified by performing a composite analysis. According to the temperature anomalies shown in Fig. 2a and Fig. 2b, the terms of the temperature equation averaged between 200 and 500 hPa are shown in Fig. 6.

For EOF1 significant cold (warm) meridional advection is seen over northeast China (Japan and Korea) with a center value of \(-0.7\) K day\(^{-1}\) (0.2 K day\(^{-1}\)) (Fig. 5b), which cause the northerly (southerly) anomaly on the western (eastern) flank of the anomalous cyclone as shown in Fig. 5c. The zonal temperature advection shows an opposite pattern to the meridional advection (Fig. 6a). The adiabatic heating and diabatic heating terms overall correspond to the anomalous vertical motion and precipitation, respectively. Because of their anomalies shown in Fig. 5a, significant diabatic cooling (adiabatic warming) is seen over most areas of East Asia except for South China. A transient eddy heat flux anomaly (Fig. 6d) is negative over the East Asian continent and positive over northeastern East Asia, but the amplitude is far weaker than for the other four heating terms. Thus the transient eddy heat flux contributes less to the temperature balance over East Asia and is nearly negligible.
The vorticity budget of EOF1 is related to the anomalous circulation shown in Figs. 5a,c. The mechanism can be summarized as follows. The westerly wind of the cyclonic anomaly advects negative vorticity from the Tibetan Plateau to East Asia \((-\nabla \vec{v}/\partial x) < 0\) and advects positive vorticity from East Asia to the North Pacific Ocean \((-\nabla \vec{v}/\partial x) > 0\) (Fig. 7a). The anomalous meridional wind generates positive (negative) vorticity; that is, \(-\beta \nu' > 0\) \((-\beta \nu' < 0\) over China (North Pacific) (Fig. 7b). In the meantime, the convergence in the upper troposphere creates positive vorticity centered in North China (Fig. 7c). The anomalous circulation alters the basic state in which transient eddies propagate, and finally changes the pattern of transient eddy vorticity flux convergence and divergence (Fig. 7d).

For the dipole pattern of EOF2 (Figs. 7f-j), it is still the horizontal vorticity advection that plays the most important role in maintaining the anomalous circulation. However, the anomalous zonal vorticity advection is astride each side of the climatological summer-mean jet core \((40^\circ N)\) (Fig. 7f). The beta effect, divergence, and transient eddy flux terms show an opposite pattern to the vorticity advection term and, thus, tend to balance the vorticity advection term. The associated mechanism is similar to that of EOF1.

The above budget analyses demonstrate that distinct circulation anomalies of the leading EOF modes induce different heating and momentum anomalies and sustain the East Asian summer UTT anomalies. To reveal the forcing factors for the unique circulation anomalies associated with the two EOF modes, we further examine the corresponding SST anomalies in the following section. The dominant time periods of the principal components of the first two leading modes indicate that the strong interannual variability modes may be related to ENSO. The connection between first two leading modes and ENSO are discussed in the coming section.

5. Relationship of the first two leading modes with ENSO

To reveal the relationship between the East Asian summer UTT change with ENSO, the lead–lag correlation coefficients between the Niño-3.4 index time series and the first two PCs are shown in Table 2. Here year \((-1)\) [year \((1)\)] denotes that the Niño-3.4 index leads (lags) in the PC time series by one year. The lead–lag correlation between PC1 and the Niño-3.4 index \((0.23,\) which is statistically significant at the 10% level) but no significant correlation with the previous winter, December–February \([D(−1)JF(0)],\) Niño-3.4 index. A

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<th>Type</th>
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The heat budget analysis for EOF2 is shown in Figs. 6f-j. The horizontal temperature advection (Fig. 6f,g) shows similar patterns as for EOF1, but the magnitude of meridional advection anomalies over Japan and Korea is weaker than that of EOF1 because of the local weaker southerly wind anomaly shown in Fig. 5d. Associated with the tripoal pattern of anomalous vertical motion, the adiabatic heating anomalies exhibit a meridional “positive–negative–positive” tripoal pattern along the East Asian continent (Fig. 6h), which is balanced by diabatic heating anomalies (Fig. 6j). The amplitude of the transient eddy heat flux anomaly is still the smallest term.

The temperature advection anomalies, \((-\nabla \vec{v}/\partial x)\)’ or \((-\nabla \vec{v}/\partial y)\)’, can be divided into two parts: the anomalies induced by anomalous wind, \(-\nabla \vec{v}/\partial x\) or \(-\nabla \vec{v}/\partial y\), and that by the anomalous temperature gradient, \(-\nabla \vec{T}/\partial x\) or \(-\nabla \vec{T}/\partial y\). The relative role of each term was calculated (figures not shown). It shows that the positive zonal advection over southern East Asia is a result due to the westerly winds of the anomalous cyclone, which advect warm temperature from the Tibetan Plateau \((-\nabla \vec{v}/\partial x > 0\). The zonal advection over northern East Asia is mainly induced by the zonal gradient of anomalous temperature, that is, \(-\nabla \vec{T}/\partial x\). Note that the meridional advection is dominated by \(-\nabla \vec{T}/\partial y\). The compensation between the zonal and meridional advection indicates the importance of the baroclinic jet.

Individual terms of the vorticity equation averaged between 200 and 500 hPa for the first two leading modes are analyzed and shown in Fig. 7. Associated with the monopole cooling of EOF1, the magnitude of anomalous zonal vorticity advection (Fig. 7a) is the largest, with a negative center over northeast China and a positive center over the North Pacific. Both the beta and divergence terms show opposite anomalies to the zonal vorticity advection, but with weaker strength. The transient eddy flux is a key factor for the vorticity balance, especially over the North Pacific (Fig. 7d). The residual term also plays an important role in the vorticity balance (Fig. 7e).
Fig. 6. Composite patterns of 200–500-hPa mean summer (a), (f) zonal advection \((-\nabla \cdot \vec{u} \cdot \partial \vec{T} / \partial x)\); (b), (g) meridional advection \((-\nabla \cdot \vec{v} \cdot \partial \vec{T} / \partial y)\); (c), (h) adiabatic heating \([- \left(p/p_0 \right)^{R/C_p} \partial \left(\vec{u} \cdot \nabla \vec{n} / \partial p\right)/\partial p]\); (d), (i) transient eddy heating \([- \left(p/p_0 \right)^{R/C_p} \nabla \left(\vec{u} \cdot \nabla \vec{n} / \partial p\right)\] \cdot \nabla)\); and (e), (j) diabatic heating anomalies \(Q_1\) (K day\(^{-1}\)) for (left) EOF1 and (right) EOF2. Contour intervals are 0.05 K day\(^{-1}\); shading denotes the 90% confidence level using a two-tailed Student’s \(t\) test.
FIG. 7. As in Fig. 6 but for the composite patterns of (a),(f) vorticity advection \((-\nabla \cdot V f)\); (b),(g) beta term \((-\beta v')\); (c),(h) divergence term \((-[(f' + \zeta') \cdot \nabla]/C_1\); (d),(i) transient eddy vorticity flux \([-u' \partial \zeta'/\partial x + v' \partial \zeta'/\partial y + \zeta' \cdot \nabla]/C_1\); and (e),(j) residual term \((res') \times 10^{-10} \text{ s}^{-2}\). Contour intervals are 0.05 \times 10^{-10} \text{ s}^{-2}.
correlation coefficient statistically significant at the 10% level is evident since May–July and persists through the following winter, indicating that the first leading mode may be associated with the El Niño developing summer. The PC2 is significantly correlated with the Niño-3.4 index in the previous winter, D(−1)JF(0), with a correlation coefficient exceeding 0.4. The significant correlation persists until the following fall and then becomes statistically insignificant at the 10% level, indicating that the second mode is associated with the El Niño decaying summer.

To further confirm the relationship between upper-tropospheric temperature changes and ENSO, the composite SST anomalies between positive and negative years of the first two leading modes are calculated. As shown in Fig. 8a, during the previous winter associated with the EOF1, strong cold SST anomalies are evident in the central and eastern Pacific Ocean and tropical Indian Ocean. During the summer, the eastern equatorial Pacific Ocean witnesses significant warm anomalies (Fig. 8b). The warm anomalies expand across the whole tropical Pacific in the following winter (Fig. 8c). Time evolution of the SST anomaly pattern demonstrates that the EOF1 of summer UTT over East Asia tends to occur during the El Niño developing summer.

For SST anomalies associated with EOF2 of the East Asian summer UTT, the eastern equatorial Pacific Ocean is dominated by significant warm anomalies in the previous winter (Fig. 8d). In the following summer, the amplitude of warm anomalies becomes weaker in the central and eastern tropical Pacific, but stronger in the tropical Indian Ocean (Fig. 8e). In the following winter, the warm anomalies in the tropical Pacific and Indian Ocean are still evident but weak in amplitude (Fig. 8f). The seasonal evolution of SST anomalies strongly indicates that the second EOF mode of East Asian summer UTT is associated with the decaying phase of El Niño.

Based on the Niño-3.4 index, we classify the entire 55 summers from 1951 to 2005 into ENSO developing (DV) and ENSO decaying (DC) summers (see Table 3). The definition of ENSO DV and DC is the same as in Chou et al. (2003) except that the ENSO persisting years are regarded as ENSO DC years in this study. Therefore, some ENSO DC years are not followed by ENSO DV years. Out of the 55 summers, 37 are associated with ENSO events. Among them, 17 summers are classified as ENSO DV year summer and 20 are classified as ENSO DC year summer.

The composite JJA UTT anomalies over East Asia during ENSO developing and decaying summers are shown in Fig. 9. In El Niño developing summers (Fig. 9a), a cooling anomaly is seen over the East Asia continent centered at 35°N, 110°E. In contrast, in La Niña developing years (Fig. 9b), the warm air center shifts northward, centered at 45°N, 120°E over northeast Asia. The difference between Fig. 9a and Fig. 9b indicates the asymmetry of the East Asian summer UTT anomalies between El Niño and La Niña events. Because EOF analysis only depicts the symmetric component of circulation anomalies, the symmetric component of UTT during ENSO developing summer (the composite field between Fig. 9a and Fig. 9b) is shown in Fig. 9c. It can be seen that the upper troposphere over East Asia is dominated by negative anomalies with a cold center located at 45°N, 120°E that resembles the spatial pattern of EOF1 shown in Fig. 2a. In ENSO decaying summer (Figs. 9d–f), a dipole pattern is seen and is evident in La Niña decaying summers (Fig. 9e). The symmetric component of ENSO decaying summers (Fig. 9f) shows that the East Asian UTT is dominated by a dipole-like anomaly with a cold (warm) center in the north (south). This pattern resembles that of EOF2 shown in Fig. 2b.

The spatial correlation coefficient between temperature anomalies of EOF1 (EOF2) and that of ENSO developing (decaying) summers over the East Asian region (20°–60°N, 90°–150°E) is 0.67 (0.87). Composite maps over the global scale of ENSO developing (decaying) summers also resemble that of EOF1 (EOF2) (figures not shown). The corresponding spatial correlation coefficient over 60°S–60°N, 0° east to 360° is 0.71 (0.83). The significant spatial correlation coefficients add reliability to the relationship between ENSO and the first two leading modes, confirming that the first and second leading modes of the summer UTT over East Asia are representative features of ENSO developing and decaying summers, respectively.

To reveal the relevant atmospheric circulation changes, the composite 850-hPa and 200-hPa wind, vertical motion

### Table 2. Lead–lag correlation coefficients between Niño-3.4 SST index and principal components of the first and second EOF modes of the summer UTT anomalies (as 3-month means stepped one month at a time). Boldface numbers indicate correlations statistically significant at the 10% level.

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</tbody>
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at 500 hPa, and precipitation anomalies during ENSO developing and decaying summers are shown in Fig. 10. During the ENSO developing summer, the eastern Pacific SST anomaly has attained a certain magnitude (Fig. 8b) that induces a suppressed Indian summer monsoon (Fig. 10a) (Wang et al. 2003). The reduced rainfall in the Indian monsoon region may force a wave train extending from northwestern India to East Asia in the upper troposphere (Fig. 10c), which is the so-called Silk Road teleconnection pattern, featuring a cyclonic anomaly over East Asia. At a lower level (850 hPa, Fig. 10a), a cyclonic anomaly over the western North Pacific is seen, created by the anomalous heating over the equatorial central Pacific, that induced local evaporation and transport of dry air into eastern China (Chou et al. 2003; Lau et al. 2000; Wu et al. 2003, 2009a). The cyclonic circulation at the upper level and northerly wind at low level leads to suppressed vertical motion and less rainfall over North China (Fig. 10a). The cyclonic anomaly over East Asia, decreased precipitation, and suppressed vertical motion provide the circulation condition for the heat and vorticity balance shown in Figs. 6–7 and maintains the UTT anomalies.

![Composite patterns of SST anomalies](image)

Table 3. Classification of the 1951–2005 summers into ENSO developing (DV) and decaying (DC) summers.

<table>
<thead>
<tr>
<th>Type</th>
<th>El Niño</th>
<th>La Niña</th>
</tr>
</thead>
</table>
During ENSO decaying summer, by the combined effects of local SST anomaly (SSTA) forcing in the northwestern Pacific and remote SSTA forcing from the Indian Ocean basin mode (IOBM) (Fig. 8e), the precipitation over the western Pacific, especially near the Philippines, decreased (Chang and Li 2000; Yang et al. 2007, 2010; Li et al. 2008; Xie et al. 2009; Wu and Zhou 2008; Wu et al. 2009a,b, 2010). The related anomalous tropical heating forced a meridional Rossby wave train as shown in Fig. 10b. This is referred to as the Pacific–Japan (PJ) or East Asian–Pacific teleconnection pattern (Nitta 1987; Huang and Sun 1992). In the upper level (Fig. 10d), the atmosphere is significantly warmed in the whole tropical region and increases the tropospheric meridional temperature gradient in the subtropics, then leads to an enhancement of westerly flow to the south of the westerly jet stream axis by the thermal balance and thereby a cyclonic anomaly over northeast Asia (Fig. 10d). The tripolar precipitation and vertical motion anomalies and the zonal extended cyclonic anomaly at the upper troposphere provide the heating and momentum flux balance and maintain the temperature anomaly pattern during ENSO decaying summer.

6. Summary and discussion

a. Summary

Two dominant interannual variability modes of summer upper-tropospheric (500–200 hPa) temperature over East Asia are revealed by using 55-yr NCEP–NCAR reanalysis data. The thermodynamic processes dominating the maintenance of the two leading modes are analyzed by doing budget and composite analysis. The atmospheric
circulation changes controlling the heating and vorticity budget are studied.

The interannual variability of upper-tropospheric (500–200 hPa mean) temperature over East Asia is dominated by two leading modes, which account for about 60.0% of total variance. The first leading EOF mode in its positive sign features a monopole cooling temperature anomaly over all of East Asia with its center located near 45°N at 250–300 hPa. The second EOF mode in its positive sign features a meridional dipole mode with the positive (negative) center located south (north) of 35°N.

Budgets and correlation analyses show that the temperature anomalies of the two interannual variability modes are both related to ENSO. The first mode is associated with ENSO developing summers, while the second mode coincides with ENSO decaying summers. They are the result of dynamical teleconnections remotely induced by ENSO and local moist processes.

In the positive sign of EOF1, the uniform UTT cooling is sustained due to the heating and vorticity balance induced by the anomalous cyclonic circulation at the upper troposphere, associated vertical motion, and decreased (excessive) precipitation over northern East Asia (South China). They are induced by teleconnections with the ENSO developing summer. During El Niño developing summer, the Indian summer monsoon precipitation decreases and may force the Silk Road teleconnection pattern at 200 hPa, which features an anomalous cyclone center over the East Asian continent. Coupled with the anomalous northerly wind in eastern China at 850 hPa, the
anomalous circulation over East Asia favors suppressed (excessive) rainfall over North China (South China).

In the positive sign of EOF2, the heating and vorticity balance for maintaining the upper-tropospheric temperature warming (cooling) over East Asia south (north) of 35°N is caused by a zonally extended anomalous cyclone in the upper troposphere and a meridional tripolar precipitation (vertical motion) anomaly along the East Asian continent. This is related with the remote forcing of ENSO during its decaying phase. In the El Niño decaying summer, driven by the combined effects of a local SST anomaly over the western Pacific Ocean and remote warm SST anomaly forcing from the Indian Ocean basin mode, precipitation is reduced over the western Pacific Ocean. Less latent heat is released and forces the Pacific–Japan teleconnection pattern along East Asia, which induces tripolar heating anomalies over East Asia. The East Asian subtropical westerly jet at 200 hPa is enhanced (weakened) south (north) of the climatological jet core, inducing a cyclonic anomaly over north East Asia.

b. Discussion

The identification of ENSO-related modes for the East Asian summer UTT is expected to be useful in East Asian summer monsoon prediction. This study provides an explanation on how the two distinct UTT anomalies are sustained. How the UTT anomalies grow deserves further study. In addition, although the first two leading modes of East Asian summer UTT resemble the UTT anomalies associated with ENSO, there are still some differences between Figs. 2a,b and Figs. 9c,f. Thus, in addition to ENSO forcing, other factors that influence East Asian summer UTT anomaly should exist. These additional forcing factors also deserve further study.

There are other opinions about the formation of Silk Road pattern and Pacific–Japan pattern (Kosaka et al. 2009; Yasui and Watanabe 2010; Kosaka and Nakamura 2006). This work indicates that the decline of Indian monsoon precipitation during the ENSO developing summer and reduced precipitation over western Pacific Ocean during the ENSO decaying summer may be the main forcing factor for them, respectively. This result is consistent with previous studies (Lau et al. 2000; Wu and Wang 2002; Wu et al. 2003; Hu et al. 2005). This study indicates that the Silk Road pattern tends to occur in the ENSO developing summers, which may provide some insight into the hypothesis that the optimal heating anomaly pattern for the Silk Road teleconnection may have a lagged relationship with ENSO (Yasui and Watanabe 2010). However, the relative roles of diabatic heating anomalies in different regions during ENSO developing summers cannot be fully identified based purely on data diagnosis and, thus, needs to be further investigated using numerical models. Numerical study is also needed for the ENSO decaying summers.

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