How Does El Niño Affect the Interannual Variability of the Boreal Summer Hadley Circulation?

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ABSTRACT

Analyses of 30-yr four reanalysis datasets [NCEP–NCAR reanalysis (NCEP1), NCEP–Department of Energy reanalysis (NCEP2), Japanese 25-year Reanalysis Project (JRA-25), and Interim ECMWF Re-Analysis (ERA-Interim)] reveal remarkably interannual variability of the Hadley circulation (HC) in boreal summer (June–August). The two leading modes of interannual variability of boreal summer HC are obtained by performing empirical orthogonal function (EOF) analysis on the mass streamfunction. A general intensification of boreal summer HC is seen in EOF-1 mode among NCEP1, NCEP2, and JRA-25 but the corresponding EOF-2 mode in ERA-Interim, while a weakened northern Hadley cell and remarkable regional variation of a southern Hadley cell are captured by the EOF-2 mode (from NCEP1, NCEP2, and JRA-25) and EOF-1 mode (from ERA-Interim), as evidenced by the enhanced (decreased) southern Hadley cell in the southern tropics (the northern tropics and southern subtropics). Both modes are driven by El Niño–like SST forcing in boreal summer, but are relevant to different phases of El Niño events. The EOF-1 (or EOF-2 derived from ERA-Interim) [EOF-2 (or EOF-1 derived from ERA-Interim)] mode is driven by SST anomalies in developing (decaying) El Niño summers. The interannual variations of the northern Hadley cell in both modes are driven by El Niño through modulating the interannual variations of the East Asian summer monsoon, while anomalous local Hadley circulation (LHC) in the regions 30°S–20°N, 110°E–180° and 30°–20°N, 160°E–120°W in response to El Niño forcing largely determine the interannual variations of southern Hadley cell in both modes, respectively. The different behaviors of the southern Hadley cell between two leading modes can be well explained by the southward shift of the tropical heating center from north of 10°N in developing El Niño summers to south of 10°N in decaying El Niño summers.

1. Introduction

The Hadley circulation (HC) is one of the most important large-scale atmospheric circulations driven by thermal forcing (Hadley 1735). In the annual mean, the HC consists of two equally strong cells, with rising air in the tropics and sinking motion in the subtropics. The HC features an evident annual cycle and asymmetric circulation, associated with a winter hemisphere cell that is much stronger than the summer hemisphere cell (Cook 2003).

The HC plays an active role in modulating the atmospheric heat transport from the tropics to the high latitudes and in determining conservation of absolute angular momentum in the free troposphere (Hou 1998; Boos and Emanuel 2008). The influence of the HC variations on regional and global climate variability is substantially significant (Wang 2002; Haarsma et al. 2008; Su et al. 2008; Döös and Nilsson 2011). Thus, understanding the mechanisms of the HC changes is of vital importance to regional climate change studies.

Previous studies focused on the long-term changes of the HC in strength and width. The boreal winter HC has intensified significantly over the last 50 years because of the uniform warming of tropical ocean (Mitas and Clement 2005; Ma and Li 2007, 2008; Zhou and Wang 2006).
The boreal summer HC in the Southern Hemisphere has undergone an interdecadal transition at the end of the 1970s and changed from a stronger-northern-limb and weaker-southern-limb state to a weaker-northern-limb and stronger-southern-limb state. This interdecadal transition is significantly correlated with the nonuniform warming trend in the Indo–western Pacific Ocean and the Atlantic Ocean (Feng et al. 2011). The long-term variation of HC in boreal spring has been investigated and an obvious strengthening trend is identified. The robust warming trend of tropical SST, especially over the Indo-Pacific warm pool, plays an essential role in modulating the variation of the HC in boreal spring (Feng et al. 2013).

There are increasing evidence of the observed widening of the HC since 1979 (Hu and Fu 2007; Hu et al. 2011). But the accuracy of poleward shift of the HC boundaries in terms of magnitude is still a controversial issue (Davis and Rosenlof 2012). Some climate models are able to reproduce the observed poleward expansion of the HC, although they underestimate the observed widening of the HC (Lu et al. 2007; Johanson and Fu 2009). Diagnosis of models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) suggests that increasing greenhouse gases play an important role in the observed poleward expansion of the HC (Hu et al. 2013), but the mechanism is poorly understood.

The HC also exhibits robust interannual variability. Analysis based on 26 years of daily upper-air wind radiosonde data found significantly positive correlations between the strength of the HC and El Niño–Southern Oscillation (Oort and Yienger 1996). The interannual variability of the boreal winter HC is mainly driven by the tropical SST anomalies, whereas the contribution of the warmer SST anomalies over the central-eastern equatorial Pacific Ocean is larger than that over the southern Indian Ocean (Sun et al. 2013b).

The Asian summer monsoon, characterized by prevailing southwest winds in the lower troposphere and northerly winds at upper levels, affects a large percentage of the population. The Asian summer monsoon system in East Asia (EA) and the western North Pacific (WNP) are parts of an expansive Asian summer monsoon system, but with very distinct features of their own. The WNP summer monsoon (WNPSM) is a typical tropical monsoon, whereas the East Asian summer monsoon (EASM) is a hybrid type of tropical and subtropical monsoon and the meridional circulation of the EASM is considered a reverse HC (RHC) (Riehl et al. 1950). The EASM exhibits multi-time-scale variability features. At decadal time scales, the tropical SST warming has driven a weakening tendency of EASM since the late 1970s (Yu and Zhou 2007; Zhou et al. 2009; Li et al. 2010). At interannual time scales, the variation of EASM and WNPSM are significantly related to El Niño–Southern Oscillation (ENSO) (Zhang et al. 1996; Kawamura et al. 2001; Chou et al. 2003). The anomalous Philippine Sea anticyclone (PSAC) is a key system that conveys the impact of El Niño to the East Asian and WNP climate (Wang et al. 2000). The anticyclone is established in the late fall of the El Niño developing years and persists until the following spring or early summer, resulting in weakening of WNPSM and enhanced EASM (Wang et al. 2001). But whether the East Asian/WNP summer monsoon can affect the boreal summer HC has never been addressed. The main motivation of the current study is to reveal how El Niño affects the interannual variability of boreal summer HC. We examine the potential relationship between the northern Hadley cell in boreal summer and EASM and the coherent variation of the southern Hadley cell and WNPSM. We attempt to explain the interannual variability of HC in the context of local Hadley circulation (LHC) variations. We find that the interannual variability modes of boreal summer HC are dominated by the anomalous LHC in response to El Niño forcing.

The remainder of the paper is organized as follows. Section 2 describes the data and the analysis method used in the study. The mean state and the two leading modes of boreal summer HC are documented in section 3. Section 4 depicts the horizontal distribution of the two leading modes, and the vertical profile of the anomalous LHC that determines the interannual variability of boreal summer HC is depicted in section 5. Section 6 discusses how El Niño affects the interannual variability of HC in boreal summer. The last section summarizes the major conclusions.

2. Data and analysis methods

a. Data description

To validate the robustness of the analysis, following the study in Mitas and Clement (2005), multireanalysis datasets are selected for present study. The datasets used in this study include the following:

1) Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) from 1979 to 2008, with 2.5° × 2.5° horizontal resolution (Xie and Arkin 1997);
2) Met Office Hadley Centre derived monthly mean sea surface temperature data from 1979 to 2008, with 1.0° × 1.0° horizontal resolution (Rayner et al. 2003);
3) National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (NCEP1, 1979–2008); the resolution is 2.5° latitude by 2.5° longitude with 17 vertical
levels (144 × 73 grid points, L17) (Kalnay et al. 1996);  
4) NCEP–Department of Energy (DOE) Atmospheric Model Intercomparison Project (AMIP-II) reanalysis (NCEP2, 1979–2008), with 2.5° × 2.5° horizontal resolution and 17 vertical layers (144 × 73 grid points, L17) (Kanamitsu et al. 2002);  
5) Japanese 25-year Reanalysis Project (JRA-25) reanalysis (1979–2008) has the same vertical resolution with NCEP1 and NCEP2, with horizontal resolution 1.25° latitude by 1.25° longitude (288 × 145 grid points, L17) (Onogi et al. 2005); and  
6) Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim) from 1979 to 2008, with 1.5° × 1.5° horizontal resolution and 37 vertical layers (240 × 121 grid points, L37) (Dee et al. 2011).

The following variables are used for present analysis from all reanalysis datasets (NCEP1, NCEP2, JRA-25, and ERA-Interim), including horizontal wind \((u, v)\), vertical velocity \(\omega\) and surface pressure.

### b. Method description

The mass streamfunction is a conventional way to depict the HC (e.g., Cook 2003; Sun et al. 2013a). Using pressure as the vertical coordinate, conservation of mass requires

\[
\frac{1}{a \cos \phi} \frac{\partial u}{\partial \lambda} + \frac{1}{a \cos \phi} \frac{\partial (v \cos \phi)}{\partial \phi} + \frac{\partial \omega}{\partial p} = 0, \tag{1}
\]

where \(u\) is zonal velocity, \(v\) is the meridional velocity, \(\omega\) is the vertical velocity in isobaric coordinates, \(p\) is pressure, \(a\) is Earth’s radius, \(\lambda\) is longitude, and \(\phi\) is latitude. If Eq. (1) is averaged over longitude, around the entire globe, then the first term on the left-hand side of the equation is zero. Using square brackets to denote this zonal average, the continuity equation is

\[
\frac{1}{a \cos \phi} \frac{\partial [(v \cos \phi)]}{\partial \phi} + \frac{\partial \omega}{\partial p} = 0. \tag{2}
\]

Equation (2) states that if one component in Eq. (2) \([(v)\] or \([\omega])\) is known, the other one can be identified. In other words, one variable can be used to fully define the two-dimensional flow. One can use the Stokes streamfunction \(\Psi\) to characterize the HC, defined by

\[
[v] = \frac{g}{2 \pi a \cos \phi} \frac{\partial \Psi}{\partial p} \quad \text{and} \quad (3)
\]

\[
[\omega] = -\frac{g}{2 \pi a^2 \cos \phi} \frac{\partial \Psi}{\partial \phi}, \tag{4}
\]

where \(g\) is the gravitational constant.

Theoretically, the streamfunction can be calculated from observation of either \([v]\) or \([\omega]\), but \([v]\) is used for practical reasons because meridional velocities are more frequently and accurately observed. Solving for \(\Psi\) and integrating from the top of the atmosphere yields

\[
\Psi(\phi, p) = \frac{2 \pi a \cos \phi}{g} \int_0^p [v(\phi, p)] \, dp. \tag{5}
\]

To ensure vertical-mean mass balance, the meridional wind fields were corrected by removing their mass-weighted vertical mean value (Kidson et al. 1969). Mass streamfunction (MSF) is applied to reveal the climatology and interannual variability of the HC. To clarify the interannual variations of the HC, a high-pass filter is applied to MSF before using an empirical orthogonal function (EOF), which removes the time scale longer than 8 years.

### 3. Results

#### a. Climatology of the Hadley circulation

The HC is a thermally driven large-scale meridional circulation. The mean state of the boreal summer [June–August (JJA)] HC is described by MSF and the results derived from the four reanalysis products (NCEP1, NCEP2, JRA-25, and ERA-Interim) are displayed in Fig. 1 (contour). The JJA mean HC is characterized with rising motion to the north of 20°N, upper-level poleward flow in both hemispheres, sinking motion in the subtropics, and flow back to the tropics near the surface, resulting in an enclosed cell in each hemisphere. One is the northern Hadley cell and the other is the southern Hadley cell. The southern Hadley cell is much stronger with respect to strength and coverage than the northern Hadley cell. As represented by the contour lines in Fig. 1, the features of mean state of HC in boreal summer derived from four reanalysis datasets are nearly identical (Fig. 1, contour).

#### b. The dominant modes of the interannual variability of Hadley circulation

To extract the major modes of the interannual variability of the HC in boreal summer (JJA), EOF analysis is performed on the JJA average MSF in the period of 1979–2008. The two leading modes from four reanalysis datasets (NCEP1, NCEP2, JRA-25, and ERA-Interim) are displayed in Fig. 1 (shaded) and the corresponding principal components (PCs) are shown in Figs. 2a and 2b. The first (EOF-1) and second EOF (EOF-2) modes are obtained by regression of the MSF anomalies onto the corresponding PC, respectively. The EOF-1 mode in NCEP1, NCEP2, JRA-25, and ERA-Interim accounts
for 64.8%, 60.5%, 41.8%, and 34.3% of the total variance, respectively, while the EOF-2 mode accounts for 16.1%, 15.4%, 24.2%, and 32.1%, respectively. In this study, we give the description of spatial patterns associated with positive phase of the two leading modes of interannual variations of boreal summer HC.

The first leading mode in NCEP1, NCEP2, and JRA-25 (Figs. 1a–c) has a quite similar spatial pattern with the respective climatological HC shown in Fig. 1 (contour), associated with negative anomalies confined to 30°S–20°N and positive anomalies northward of 20°N. As a result, there is a general increase of the northern and southern Hadley cells when the anomalies in EOF-1 mode are added to the mean state of the HC. These general features aforementioned in Figs. 1a–c also can be seen in the second leading mode derived from
FIG. 2. (a), (b) Time series of the two leading EOF modes of the MSF in JJA derived from four reanalysis products, along with (c)–(j) the wavelet analysis (shaded left, stippled area indicates >95% level) and the power spectral analysis (curve right).
ERA-Interim (Fig. 1d). Note that discrepancies are still detectable among the four reanalysis products, despite the large similarities that exist northward of 20°S. For instance, positive anomalies in the NCEP1, NCEP2, and ERA-Interim control the low levels of the southern subtropics (southward of 20°S) (Figs. 1a,b,d), whereas it is dominated by the negative anomalies in the JRA-25 reanalysis (Fig. 1c).

The spatial structure of the EOF-2 mode derived from three reanalysis datasets (NCEP1, NCEP2, and JRA-25) are shown in Figs. 1e–g (shaded), respectively. The EOF-2 mode captures significant regional characteristics of the southern Hadley cell variations and coherent variation of the northern Hadley cell. As shown in Figs. 1e–g, the positive anomalies control the regions of 30°–20°S and 5°–25°N, associated with the negative values confined in 20°S–5°N and north of 25°N, resulting in a general decrease of the northern Hadley cell and an increase of the southern Hadley cell in southern tropical (20°S–5°N) and decreased southern Hadley cell in the southern subtropics and northern tropics (5°–25°N). These features of HC variability also reflect on the EOF-1 mode in ERA-Interim (Fig. 1h, shaded). Although spatial distributions have large similarities among these reanalysis datasets, the discrepancies still can be seen among them. For instance, the positive anomalies in ERA-Interim are generally stronger than in the other three reanalyses, whereas the negative anomalies to north of 20°N derived from JRA-25 are weaker than those of other three reanalyses.

In brief, the two distinct modes of interannual variability of boreal summer HC are revealed by all the four datasets, even though the sequence of spatial modes in ERA-Interim is inconsistent with other three reanalysis products (NCEP1, NCEP2, and JRA-25). The two dominant modes capture different behaviors of HC variations. A general increase and decrease of the northern Hadley cell is seen in EOF-1 mode (EOF-2 mode in ERA-Interim) and EOF-2 mode (EOF-1 mode in ERA-Interim), respectively. There is an increase of the southern Hadley cell in tropics in EOF-1 mode (EOF-2 mode in ERA-Interim), while an increase of the southern Hadley cell in the southern tropics and decrease of the southern Hadley cell in the southern subtropics and northern tropics are found in EOF-2 mode (EOF-1 mode in ERA-Interim). What factors are responsible for different behaviors of HC will be addressed in section 6.

Figures 2a and 2b present the temporal evolution of the two leading modes of boreal summer HC derived from the four reanalyses. To diagnose the consistency of different reanalysis datasets in the description of the temporal evolution of two leading modes, the pairwise correlation coefficients of the corresponding PCs are calculated among these reanalysis datasets as shown in the following square matrix. The lower (upper) triangular matrix represents the pairwise correlation coefficients calculated between PC1 (PC2) from NCEP1, NCEP2, and JRA-25 and PC2 (PC1) from ERA-Interim. All the results are >99% confidence level.

### Table 1. To evaluate the consistency of temporal evolution of two leading modes derived from multireanalysis data (NCEP1, NCEP2, JRA-25, and ERA-Interim), pairwise correlation coefficients of the corresponding PCs are calculated among these reanalysis datasets as shown in the following square matrix. The lower (upper) triangular matrix represents the pairwise correlation coefficients calculated between PC1 (PC2) from NCEP1, NCEP2, and JRA-25 and PC2 (PC1) from ERA-Interim. All the results are >99% confidence level.

<table>
<thead>
<tr>
<th></th>
<th>PC1</th>
<th>PC2</th>
<th>JRA-25</th>
<th>ERA-Interim</th>
</tr>
</thead>
<tbody>
<tr>
<td>NCEP1</td>
<td>1</td>
<td>0.81</td>
<td>0.58</td>
<td>0.71</td>
</tr>
<tr>
<td>NCEP2</td>
<td>0.88</td>
<td>1</td>
<td>0.66</td>
<td>0.81</td>
</tr>
<tr>
<td>JRA-25</td>
<td>0.62</td>
<td>0.71</td>
<td>1</td>
<td>0.68</td>
</tr>
<tr>
<td>ERA-Interim (PC2)</td>
<td>0.79</td>
<td>0.78</td>
<td>0.46</td>
<td>1</td>
</tr>
</tbody>
</table>

Correlation coefficients are calculated between the first (second) principal component [PC1 (PC2)] from NCEP1, NCEP2, and JRA-25 and the second (first) principal component in ERA-Interim, and the corresponding results are displayed in the lower (upper) triangular matrix listed in Table 1. All the correlation coefficients listed in Table 1 are statistically significant at the 1% level. Thus the four reanalyses are consistent in their description of the temporal evolution of the two leading modes of the boreal summer HC.

The interannual variations of the two leading modes are well captured by four reanalysis datasets, as evidenced by wavelet analysis and power spectral analysis (Figs. 2c–j). The EOF-1 mode in NCEP1, NCEP2, and JRA-25 (Figs. 2c–e) exhibits interannual variability quite similar to the results of the EOF-2 mode in ERA-Interim (Fig. 2f), with a dominant peak of 2.5 yr after the mid-1990s. We also examine the wavelet analysis and power spectral analysis of PC2 (NCEP1, NCEP2, and JRA-25) and PC1 (ERA-Interim) (Figs. 2g–j), but there is less consistency among them compared with same analysis of PC1 (NCEP1, NCEP2, and JRA-25) and PC2 (ERA-Interim) (Figs. 2c–f). The result in NCEP1 coincides well with NCEP2 in terms of wavelet and power spectral analysis, with a dominant peak of 2.5 yr. Discrepancies of wavelet analysis are still evident between JRA-25 and ERA-Interim, despite sharing a similar power spectral distribution, with dominant peaks of 2 and 4 yr.

### 4. Horizontal distribution of anomalous Hadley circulation

As shown in Fig. 1, the HC is usually revealed by the zonal mean MSF. However, there is an evident deficiency in using MSF to depict the HC, since the variability of the local HC cannot be depicted. To illustrate
the horizontal distribution of the interannual variability of the HC, following the analysis method in Quan et al. (2004), another HC index is denoted by V100-V850. The mean states of this index (vertical shear of the meridional wind; meridional wind at 100 hPa (V100) minus meridional wind at 850 hPa (V850)) from different reanalyses (NCEP1, NCEP2, JRA-25, and ERA-Interim) are first examined to illustrate whether it is able to depict the climatological features of the zonal mean HC. As shown in Fig. 3a, the negative and positive values can be used to roughly distinguish the southern and the northern Hadley cells. This is easy to understand considering that the southern (northern) Hadley cell is anticlockwise (clockwise) large-scale circulation, associated with prevailing northerlies (southerlies) at upper levels and southerlies (northerlies) at low levels. As such, in NCEP1 and NCEP2, the negative values control the region between 30°S and 20°N, while the positive values dominate the region north of 20°N. Compared with NCEP1 and NCEP2, the latitudinal position where the value of V100-V850 equals zero, as derived from JRA-25 and ERA-Interim, slightly shifts southward. These features coincide well with the HC characteristics measured by the corresponding MSF presented in Fig. 1a. Therefore, the vertical shear of meridional wind is a useful metric to gauge the mean state of zonal averaged HC, and thus it was selected to describe the interannual variability of the boreal summer HC.

Second, to further examine the robustness of the two leading EOF modes based on MSF, a composite analysis is applied to V100-V850 (Figs. 3b,c). Based on Table 2, we have six positive cases for the EOF-1 mode (EOF-2 mode for ERA-Interim) and four positive cases for the EOF-2 mode (EOF-1 mode for ERA-Interim). The years selected for composite analysis are the positive phase of corresponding PC and El Niño events. As shown in Fig. 3b, composited anomalies of meridional wind shear share large similarities between NCEP1 and NCEP2, characterized by negative anomalies dominating between 30°S and 20°N and positive anomalies confined northward of 20°N (Fig. 3b). As a result, there is a general increase of each Hadley cell, which is consistent with the results of EOF-1 mode (shaded in Figs. 1a,b). A similar distribution of composited V100-V850 is also seen in both JRA-25 and ERA-Interim, except that the positive anomalies in the Northern Hemisphere are larger in terms of extent than in NCEP1 and NCEP2. In particular, these discrepancies are in also reflected in the MSF in ERA-Interim (Fig. 1d), manifest as positive anomalies in the Northern Hemisphere extending from low (north of 15°N) to upper levels (10°N).

A composite analysis for the EOF-2 mode (EOF-1 mode for ERA-Interim) is found in Fig. 3c. A weakened northern Hadley cell and regional characteristics of variations in the southern Hadley cell are well captured by the vertical shear of meridional wind (V100-V850), as reflected in the MSF (Figs. 1e–h). All the reanalyses can reasonably describe the weakened northern Hadley cell and decrease of the southern Hadley cell in the northern tropics and southern subtropics, while the increased southern Hadley cell in the southern tropics is narrower in NCEP1 and JRA-25 in terms of extent than in NCEP2 and ERA-Interim.

As discussed above, the composited V100-V850 can realistically capture the main features of the two leading
In summary, V100-V850 is useful to describe the horizontal distributions of two leading modes of interannual variability of boreal summer HC. Both modes suggest that the interannual variability of the northern Hadley cell is associated with maximal anomalies in the EASM region, indicating that the interannual variability of the northern Hadley cell may be linked to the coherent variability of EASM activity. The increased southern Hadley cell in EOF-1 mode (EOF-2 mode for ERA-Interim) is directly connected to the cross-equatorial negative anomalies (30°S–20°N, 110°E–180°), while distinct positive anomalies (5°–25°N, 160°E–120°W) and negative anomalies (0°–30°S, 160°E–120°W) in the Pacific, which tend to be symmetrical about the equator, could explain the decrease and increase of the southern Hadley cell in the northern and southern tropics in EOF-2 mode (EOF-1 mode for ERA-Interim), respectively. What factors could explain the large differences of the southern Hadley cell between both modes and whether the East Asian/western North Pacific summer monsoon affect the interannual variability of the boreal summer HC will be further addressed in detail in section 6.

5. Two distinct modes of Hadley circulation dominated by anomalous LHC

To further confirm whether the two leading modes of interannual variability of boreal summer HC can be represented by the anomalous LHC in the key regions extracted by the V100-V850, following Chung et al. (2011), the vertical profiles of anomalous LHC relevant to the two leading modes are further revealed by the regression of the corresponding PC onto the zonally averaged vertical velocity \(v\) and divergent component of meridional wind \(V_d\) at each pressure level along 110°–140°E (110°E–180° and 160°E–120°W) to identify the different behaviors of the northern (southern) Hadley cell in both modes. As shown in Figs. 5a–d, the statistically significant anomalous rising motions (shading indicates >90% confidence level) are seen at 10°–20°N, while the statistically significant descending motions dominate both subtropics. As a result, the two anomalous enclosed cells generated in each hemisphere and each cell are comparable to the climatology of the northern and southern Hadley cells leading to a general increase of the northern and southern Hadley cells, respectively. These results coincide well with the features revealed by the EOF-1mode derived from NCEP1, NCEP2, and JRA-25 (Figs. 1a–c) and the EOF-2 mode in ERA-Interim (Fig. 1d) and the vertical shear of the meridional wind (V100-V850) (Fig. 3b).

Following the same analysis method, we examined the vertical profile of the anomalous LHC relevant to the

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Table 2. To display the horizontal structure of the atmospheric circulation changes associated with the HC variability modes, a composite analysis method is applied to two leading modes of four reanalysis datasets, respectively. We choose six positive phase years for EOF-1 mode (NCEP1, NCEP2, and JRA-25) and EOF-2 mode (ERA-Interim), while four positive phase years are selected for EOF-2 mode (NCEP1, NCEP2, and JRA-25) and EOF-1 mode (ERA-Interim). The years selected for composite analysis are the positive phase of the corresponding PC among four reanalysis datasets (NCEP1, NCEP2, JRA-25, and ERA-Interim) and El Niño events as listed in Table 2.

<table>
<thead>
<tr>
<th>Positive phase</th>
<th>PC1 (NCEP1, NCEP2, and JRA-25) and PC2 (ERA-Interim)</th>
<th>PC2 (NCEP1, NCEP2, and JRA-25) and PC1 (ERA-Interim)</th>
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<tbody>
<tr>
<td>1982</td>
<td>1982</td>
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modes of interannual variability of the boreal summer HC. Thus the horizontal distributions of V100-V850 are further examined for the two leading modes to find out where the interannual variations of boreal summer HC are largely dominant. As shown in Figs. 4a–d, the significant positive anomalies in the Northern Hemisphere are seen in the EASM regions (20°–45°N, 110°–140°E), while the cross-equatorial negative anomalies dominate the region 30°S–20°N, 110°E–180°. Since these aforementioned positive and negative anomalies both have the same sign as the mean state of the northern and southern Hadley cells, respectively (Fig. 3a), they are contributed to the increased northern and southern Hadley cells in EOF-1 mode (EOF-2 mode for ERA-Interim), respectively.

We also examine the horizontal distributions of the EOF-2 mode (EOF-1 mode for ERA-Interim). As shown in Figs. 4e–h, the EOF-2 mode (EOF-1 mode for ERA-Interim) is well distinguished from the EOF-1 mode (EOF-2 mode for ERA-Interim) in horizontal distribution. The distinct negative anomalies dominate the EASM region (20°–45°N, 110°–140°E). These aforementioned anomalies when compared to the zonal mean V100-V850 (Fig. 3a) result in a weakened northern Hadley cell. Meanwhile, distinct positive anomalies (5°–25°N, 160°E–120°W) and negative anomalies (0°–30°S, 160°E–120°W) in Pacific are found on both sides of equator and tend to be symmetrical about equator, which could largely explain the decreased southern Hadley cell in the northern tropics and increased southern Hadley cell in the southern tropics in comparison with the zonal mean southern Hadley cell (Fig. 3a).
EOF-2 mode from NCEP1, NCEP2, and JRA-25 (Figs. 5e–g) and EOF-1 mode in ERA-Interim (Fig. 5h). The features of the EOF-2 mode derived from NCEP1, NCEP2, and JRA-25 and the EOF-2 mode in ERA-Interim are well captured by Figs. 5e–h, characterized by statistically significant ascending motions near the equator (south of 10°N) and north of 30°N and statistically significant descending motions in both subtropics. As
a result, anomalous clockwise circulation is seen between the equator and 25°N, together with anomalous anticlockwise circulation in the regions of 25°–40°N and 25°S–0°. These features above are consistent with the results in the EOF-2 mode based on NCEP1, NCEP2, and JRA-25 (Figs. 1e–g) and the EOF-1 mode derived from ERA-Interim (Fig. 1h).

Given the fact that the regressed vector (Fig. 5) is only confined in the East Asian monsoon sectors for the northern Hadley cell in both modes, whereas it was performed along 110°E–180° and 160°E–120°W to describe the two leading modes of the southern Hadley cell, respectively, it follows that the interannual variability of the northern Hadley cell in boreal summer is largely dominated by

![Fig. 5: The vertical profiles of the anomalous LHC relevant to two leading modes represented by the regression of corresponding PC onto the zonally averaged $\omega$ and $V_d$ at each pressure level along 110°E–140°E and 110°E–180° to identify the variations of the northern and southern Hadley cell in EOF-1 modes derived from (a) NCEP1, (b) NCEP2, and (c) JRA-25, and (d) EOF-2 mode in ERA-Interim. (e)–(h) As in (a)–(d), but for 160°E–120°W for the southern Hadley cell. The contours represent the regression coefficients of corresponding PC onto the averaged $\omega$ and statistically significant regression coefficients are shaded (>90% confidence level).]
EASM variations, while the anomalous LHC at 110°E–180° and 160°E–120°W largely determines the interannual variability of the boreal summer southern Hadley cell in the two leading modes, respectively.

6. How does El Niño affect the interannual variability of the HC in JJA?

a. The interannual variability of HC relevant to different phases of El Niño

The main source of the interannual variability in the climate system is from ENSO, which is primarily attributed to tropical air–sea interactions. ENSO has an important and direct influence on boreal winter HC through SST variability (Oort and Yienger 1996). Note that the SST anomalies over the equatorial eastern-central Pacific are established in the early summer of the El Niño developing years and persists until the following spring or early summer (Wang et al. 2000). A key question is whether ENSO has an impact on the interannual variability of the boreal summer HC. As shown in Fig. 6, PC1 and PC2 derived from NCEP1 and ERA-Interim are both significantly correlated to the homogeneous SST anomalies over the equatorial eastern-central Pacific (i.e., the positive phase of the corresponding PC is linked to El Niño and vice versa). In the current study, we only attempt to investigate the possible linkage of the positive phases of the two leading modes to the El Niño events. To identify whether HC variations are related to boreal summer SST anomalies in developing or decaying phases of El Niño, Fig. 6 shows the lead–lag correlation coefficients between the SST anomalies and the principal components of the two leading modes of boreal summer HC. PC1 in NCEP1 (shading; left panel) associated with PC2 in ERA-Interim (contours; left panel) shows significant correlation with SST anomalies during El Niño developing summers (Fig. 6, left), while PC2 in NCEP1 and PC1 in ERA-Interim are both related to El Niño decaying stages (Fig. 6, right). Note that it is time–length dependent for the relationship between the interannual variation of boreal summer HC and SST anomalies leading PC2 (or PC1 in ERA-Interim) by two seasons (Fig. 6e); that is, the aforementioned significant correlation shown in Fig. 6e is more evident during 1979–97 than during 1979–2008. As such, we merely investigate the issue in the current study of how the SST anomalies during El Niño decaying summers in the period of 1979–97 affect the interannual variation of the boreal summer HC.

We also examined the lead–lag correlation based on NCEP2 and JRA-25. The results in NCEP2 are quite similar to NCEP1 (not shown), but different results are seen in JRA-25. As shown in Fig. 7, the temporal evolutions of two leading modes in JRA-25 are both closely related to the SST anomalies during El Niño developing summers. Thus, it is quite difficult to understand why the EOF-2 mode is closely correlated with the SST anomalies in the developing El Niño summers rather than decaying summers, given the high correlation coefficient of PC2 in JRA-25 compared with other three reanalyses (0.58, 0.66, and 0.68) listed in Table 1. To explain this paradox, we examined the temporal evolution of the EOF-2 mode derived from NCEP1, NCEP2, and JRA-25 (Figs. 2g–i) and the EOF-1 mode in ERA-Interim (Fig. 2j) and found that one of the three maximal variations is unique for JRA-25 and confined in the period of 1985–95; that is, PC2 includes some special events of El Niño (1987, 1991, and 1992), and all of them not only belong to El Niño developing summers but also belong to decaying summers, which could be a possible cause for the high correlations between PC2 and warmer SST anomalies during El Niño developing summer. To test this hypothesis, Fig. 8 represents the composite MSF derived from JRA-25 for the special events (1987, 1991, and 1992) and the results are almost same as the EOF-2 mode as displayed in Fig. 1g. Thus the PC2 in the JRA-25 reanalysis emphasized too much the interannual signal of El Niño that lasted longer than one year, which is a possible cause of high correlation between PC2 in JRA-25 and SST anomalies in developing El Niño summers.

In summary, there are two distinct modes of interannual variability of HC in boreal summer for four reanalysis datasets. Both modes are significantly correlated with SST anomalies over the eastern-central Pacific, but relevant to different phases of El Niño events. Given the fact that the two leading modes of interannual variability of boreal summer HC are dominated by anomalous LHC, whether and how the El Niño affects the anomalous LHC will be described in the following sections in detail.

b. Impact of El Niño developing phase SST on the anomalous LHC

In this section, the composite analysis method is performed on the atmospheric horizontal circulation to study how the El Niño developing phase affects the interannual variability of the boreal summer HC. The mean state of the 850- and 100-hPa wind vectors are first examined (Figs. 9a–d: 850 hPa; Figs. 9e–h: 100 hPa) derived from four reanalysis datasets (NCEP1, NCEP2, JRA-25, and ERA-Interim). Figures 9a–d present large-scale summer monsoon circulation in the lower troposphere. East
Fig. 6. (left) Correlation maps (contours; solid and dashed represent positive and negative correlations, respectively) of the PC of the EOF-2 mode derived from ERA-Interim with reference to SST anomalies at (a) lag = −1 (PC2 lags SST anomalies by 1 season), (b) lag = 0 (simultaneous), (c) lag = 1 (PC2 leads SST anomalies by 1 season), and (d) lag = 2 (PC2 leads SST anomalies by 2 seasons). (right) The similar correlation maps (contours) of the PC1 derived from ERA-Interim with reference to SST anomalies: (e) lag = −2 (PC1 lags SST anomalies by 2 seasons), (f) lag = −1 (PC1 lags SST anomalies by 1 season), (g) lag = 0 (simultaneous), and (h) lag = 1 (PC1 leads SST anomalies by 1 season). The correlation maps of PC1/PC2 derived from NCEP1 with reference to SST anomalies are revealed by the shading. Shaded and contoured regions indicate >95% confidence level.
Asia mainly has three branches of southwest flow for normal climatological conditions (Zhou and Yu 2005). The first is from the Indian monsoon, which is linked to the robust cross-equatorial Somali jet and brings abundant moisture from the Arabian Sea and the Bay of Bengal, crossing the Indochina Peninsula and South China Sea into eastern China. The second is the southwest flow from the western Pacific, which is often linked to the western Pacific subtropical high (WPSH). The third is the cross-equatorial flow straddling 105°–150°E, which is the weakest one among these three branches (Figs. 9a–d). In the upper troposphere, the South Asian high dominates East Asia, associated with robust northerlies located to its east. (Figs. 9e–h).

![Fig. 7](https://example.com/fig7.png)

**Fig. 7.** As in Fig. 6, but for correlation maps of the PCs from the two leading EOF modes derived from JRA-25 with reference to SST anomalies: (left) PC1 and (right) PC2. Shaded regions indicate >95% confidence level.
We first examined how the anomalous LHC is established during the developing El Niño summers and then considered whether it was relevant to the variations in the East Asian/western North Pacific summer monsoon. As shown in Figs. 10a–d, an anomalous low-level cyclonic circulation dominates the Philippine Sea in the El Niño developing summers, resulting from an El Niño–induced anomalous westerly in the equatorial western-central Pacific generating strong positive shear vorticity and thereby enhancing precipitation (shaded in Figs. 10a–d) and cyclonic anomalies over the Philippine Sea (PSC) (Wang and Zhang 2002). Accordingly, an anomalous upper-level anticyclone and upper-level cyclone anomaly dominate the Philippine Sea (PSAC) and the region 20°–45°N, 0°–140°E, respectively (Figs. 10e–h). As shown in Fig. 10, anomalous low-level (upper-level) cyclonic (anticyclonic) circulation over the Philippine Sea is the key system that bridges the SST anomalies in the developing El Niño summers and low-level (upper-level) branch of LHC in the EASM region consists of northerly winds on the northwestern flank of low-level PSC (southerly winds in the northwestern flank of upper-level PSAC and eastern flank of anomalous upper-level cyclone in the region 20°–45°N, 0°–140°E). The low-level (upper-level) branch of anomalous LHC in the region 30°S–20°N, 110°E–180° comprises the robust cross-equatorial flow on the southeastern flank of the low-level PSC (upper-level PASC).

The ascending and descending branches of anomalous LHC are both closely associated with anomalies in the latent heating release. In developing El Niño summers, the maximal increase in precipitation is seen in the southern flank of anomalous low-level PSC (10°–20°N), which leads to an increase of the latent heating release (shaded in Figs. 10a–d). Thus the upper atmosphere is heated from the increasing release of latent heating, accompanied with enhancement of local rising motion as manifested by the composited vertical velocity at 500 hPa (shaded in Figs. 10e–h). In addition, anomalous descending motions are also seen in southern subtropics (30°–20°S, 110°E–180°) and EASM region north of 30°N.
(Figs. 10e–h), associated with anomalous ascending motion at 20°N in the EASM region (Figs. 10e–h), which is in good agreement with latent heating release anomalies in these regions above. Note that the latitudinal locations of anomalous ascending and descending motions are consistent with those of the anomalous LHC displayed in Figs. 5a–d.

As a result, the anomalous LHC is generated in the EASM region, characterized by rising motion near 20°N, poleward flow at upper levels, sinking motion northward of 30°N, and flow back to 20°N near the surface. Anomalous LHC is also seen in the region 30°S–20°N, 110°E–180°, consisting of ascending motion at 10°–20°N, southward cross-equatorial flow in the upper level,
Fig. 10. Composited horizontal circulation at 850 hPa (vectors; m s$^{-1}$) and precipitation (shaded; mm day$^{-1}$) in the developing El Niño summers based on (a) NCEP1, (b) NCEP2, (c) JRA-25, and (d) ERA-Interim. (e)–(h) As in (a)–(d), but for the composited horizontal circulation at 100 hPa (vectors; m s$^{-1}$) and vertical velocity in isobaric coordinates at 500 hPa (shaded; Pa s$^{-1}$).
descending motion in the southern subtropics (30°–20°S, 110°E–180°), and northward cross-equatorial flow returning to the tropics near the surface.

The anomalous LHC in the EASM region is actually a weakened EASM circulation in the developing El Niño summers. The EASM circulation is usually considered as a reversed Hadley cell since its direction is opposite that of the northern Hadley cell (Richel et al. 1950; Ye and Yang 1979; Zhou and Li 2002; Chen et al. 2010). Hence a weaker EASM is beneficial for a stronger northern Hadley cell. Meanwhile, the anomalous LHC in the region 30°S–20°N, 110°E–180° features the coherent variation with WNPSC circulation in the developing El Niño summers.

c. Impact of decaying El Niño phase SST on anomalous LHC

We also examined the role of warmer SST anomalies in the eastern-central Pacific in the establishment of anomalous LHC during decaying El Niño summers. Figure 11 represents the response of the atmospheric circulation to the SST forcing using four reanalysis datasets in the decaying El Niño summers. There is a pair of anomalous low-level cyclonic circulations in the eastern-central Pacific (Figs. 11a–d), associated with a pair of anticyclonic circulations in the upper troposphere (Figs. 11e–h). The local atmospheric response can be explained by the Gill model (Gill 1980). A low-level anticyclone anomaly dominates the Philippine Sea during El Niño decaying summers (Figs. 11a–d), resulting from wind–evaporation–SST feedback (Wang et al. 2000). Accordingly, an upper-level cyclone anomaly is found in the Philippine Sea (Figs. 11e–h). Both the Indian Ocean and northern Pacific Ocean SST anomalies contribute to sustain the anticyclone (Wu et al. 2009; Xie et al. 2009). The former plays a crucial role in early summer, while the latter dominates in late summer (Wu et al. 2010).

The anomalous LHC in the EASM region is mainly associated with the anomalous low-level PSAC and upper-level PSC during El Niño decaying summers, while the anomalous low-level cyclone and upper-level anticyclone in eastern-central Pacific as well as anomalous low-level PSAC and upper-level PSC are both attributed to the anomalous LHC in the region 30°–20°S, 160°E–120°W; that is, the anomalous LHC generated in the EASM region comprises the northerly anomaly and southerly anomaly on the northwestern flank of upper-level PSC and low-level PASC, descending motion near 20°N, and rising motion to the north of 30°N as manifest by vertical velocity at 500 hPa (shaded in Figs. 11e–h). The anomalous LHC in the region 30°–20°S, 160°E–120°W is directly related to the anomalous rising motion in the equatorial western and eastern-central Pacific (Figs. 11e–h). The anomalous ascending motion stems from convergence of airflow in the equatorial western and eastern-central Pacific, respectively. The former arises from the convergence of airflow between north-easterly wind on the eastern flank of low-level PSAC and southeasterly wind in the Southern Hemisphere (30°S–4°, 160°E–180°), while the latter is caused by a pair of cyclonic circulation anomalies in eastern-central Pacific, symmetrical about the equator. Correspondingly, the anomalous upper-level divergences poleward are seen in the equatorial western and eastern-central Pacific, respectively, associated with anomalous descending motion in both subtropics (160°E–120°W), respectively.

In brief, anomalous LHC in EASM region is indeed a stronger EASM circulation in decaying El Niño summers, which reverses to the condition during developing El Niño summers. Thus the two leading modes of the interannual variability of the northern Hadley cell are dominated by the interannual variability of the EASM. However, the interannual variability of the southern Hadley cell in the two leading modes features well-distinguished latitudinal positions of the rising branch and dominated by anomalous LHC in the region 30°S–20°N, 110°E–180° and LHC in the region 30°S–20°N, 160°E–120°W, respectively. The rising branch of anomalous LHC is confined to 10°–20°N in developing El Niño summers, while the ascending branch of anomalous LHC is seen in 0°–10°N during the decaying El Niño summers. Studies have shown that the cross-equatorial winter Hadley cell depends profoundly on the tropical thermal forcing displaced from the equator (i.e., it is sensitive to a small latitudinal shift or concentration of tropical convective heating in the summer hemisphere; Lindzen and Hou 1988; Hou and Lindzen 1992). Further study suggested that the intensity of the cross-equatorial winter Hadley cell is significantly stronger when tropical convective heating is centered off the equator in the summer hemisphere than when tropical thermal forcing is centered near the equator (Hou 1993). As represented by the shading in Figs. 10a–d and Figs. 11a–d, a general increase of the southern Hadley cell in the developing El Niño summers is associated with tropical latent heating release located off the equator (north of 10°N), while the regional characteristics of the southern Hadley cell are directly related to the tropical latent heating release near the equator (south of 10°N) during decaying El Niño summers. Thus the different behaviors of the southern Hadley cell in both modes are attributed to a southward shift of tropical latent heating center from north of 10°N in developing El Niño summers to south of 10°N in decaying El Niño summers. In essence, the interannual variability of boreal summer HC is the response to the El Niño forcing.
7. Summary

The two distinct modes of the interannual variability of the HC in boreal summer are revealed using four reanalysis datasets (NCEP1, NCEP2, JRA-25, and ERA-Interim). Both modes are an El Niño–forced response, but are relevant to different phases of El Niño. We examined the linkage of interannual variability of the northern Hadley cell in boreal summer with the interannual variations of the EASM and coherent

Fig. 11. As in Fig. 10, but for the results from the decaying El Niño summers.
variation of the southern Hadley cell and WNPSM. We also provide the possible explanation of different behaviors of the southern Hadley cell in both modes. Following is a summary of the main results.

The two leading modes of interannual variability of boreal summer HC are obtained by performing EOF analysis on the boreal summer MSF. A stronger northern Hadley cell is seen in EOF-1 mode (NCEP1, NCEP2, and ERA-25) and EOF-2 mode (ERA-Interim), whereas a decreased northern Hadley cell is found in EOF-2 mode (NCEP1, NCEP2, and JRA-25) and EOF-1 mode (ERA-Interim). A general increase of the southern Hadley cell is also seen in EOF-1 mode (NCEP1, NCEP2, and JRA-25) and EOF-2 mode (ERA-Interim), while the regional variations of the southern Hadley cell are well captured by EOF-2 mode (NCEP1, NCEP2, and JRA-25) and EOF-1 mode (ERA-Interim) with an increased southern Hadley cell in the southern tropics (20°S–5°N) and decreased southern Hadley cell north of 5°N and in the southern subtropics. The above features are also reflected in the vertical shear of meridional wind (V100-V850).

The two leading modes of interannual variability of the northern Hadley cell are mainly determined by the interannual variations of the EASM, while the anomalous LHC in the regions 30°S–20°N, 110°E–180° and 30°S–20°N, 160°E–120°W largely determine the two leading modes of the southern Hadley cell, respectively.

Attribution analysis suggests that the temporal evolution of the two leading EOF modes is driven by El Niño–like anomalies over the equatorial eastern-central Pacific, but are relevant to different phases of El Niño events. The EOF-1 mode (NCEP1, NCEP2, and JRA-25) and EOF-2 mode (ERA-Interim) are closely correlated with the SST anomalies in the developing summers, while the EOF-2 mode (NCEP1, NCEP2, and JRA-25) and EOF-1 mode (ERA-Interim) are related to the SST anomalies during the decaying El Niño summers.

The interannual variations of the northern Hadley cell in both modes are driven by El Niño through modulating the interannual variations of EASM. The anomalous low-level cyclone (anticyclone) associated with anomalous upper-level anticyclone (cyclone) in the Philippine Sea is the key system that conveys the impact of the SST anomaly during developing (decaying) El Niño summers on the interannual variability of EASM, resulting in a stronger (weaker) northern Hadley cell. A general increase of the southern Hadley cell features coherent interannual variation of WNPSM in developing El Niño summers, resulting from distinct cross-equatorial airflows at the low and upper levels (30°S–20°N, 110°E–180°). Anomalous LHC in the region 30°S–20°N, 160°E–120°W driven by El Niño largely determines the interannual variation of the southern Hadley cell in decaying El Niño summers, associated with anomalous low-level (upper-level) convergence (divergence) at 160°E–120°W near the equator and descending motions seen in both sub tropics at 160°E–120°W. The different behaviors of the southern Hadley cell in the two modes are mainly caused by the southward shift of the tropical latent heating center from north of 10°N in developing El Niño summers to south of 10°N during decaying El Niño summers.

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