Principal Modes of Summertime Zonal-Mean Flow and Their Connections with the AO and ENSO*

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ABSTRACT

The NCEP/NCAR reanalysis data have been employed to diagnose variations of the zonal mean flow in boreal summer. Two leading EOF modes are found to dominate the spatial and temporal changes of the summertime zonal mean winds in the troposphere. EOF1 shows the distribution of zonal-mean flow anomalies with higher variance in the North Polar Region, whereas the EOF2 shows the distribution of zonal-mean flow anomalies with higher variance in tropical and extra-tropical regions. The EOF1 and EOF2 have respectively the periodicities similar to those of AO and ENSO. Significant lag correlations have been found between EOF1 and ENSO, and between EOF2 and AO, in the seasons including spring, autumn, and winter. However, no significant correlations have been found between EOF1 in summer and ENSO in any other seasons, and between EOF2 in summer and AO in other seasons, no matter how big the lag that represents number of seasons has been set. These results suggest that the principal modes of summertime zonal mean flow could be statistically separated from each other. Hence, EOF1 and EOF2 are physically related to the AO and ENSO respectively. A theory called quasi-geostrophic non-acceleration theorem has been used to partly explain the possible mechanisms of the maintenance of the two principal modes. The composite differences of the divergence of Eliassen-Palm flux (E-P flux) between positive and negative years as obtained from the time series of EOF1 and EOF2 display the distributions that contribute to the zonal mean wind anomalies represented by EOF1 and EOF2, respectively. The planetary other than the synoptic waves dominate the behaviors of the E-P fluxes, suggesting the crucial role of the planetary waves in the maintenance of the zonal mean flow anomalies. The residual circulation as well as the friction, which cancel the divergence of the E-P flux, also play an important role in some places. These results are very helpful for our better understanding how the anomalous zonal mean flows maintain and how the ENSO and AO influence the global climate variations.

Key words: boreal summer, zonal-mean flow, Arctic Oscillation (AO), ENSO, Eliassen-Palm flux (E-P flux)

1. Introduction

As the major component of the atmospheric general circulation, zonal mean circulation plays an extremely important role in the formation of climate. Studies indicate that effects of eddies and external forcings mainly contribute to the zonal mean flow changes (Hoskins and Pearce, 1987). The effects of eddies on zonal mean flow may be explained by Eliassen-Palm flux (“E-P flux” for short). Studies of Nigam and Lindzen (1989) showed that stationary waves play a dominant role to mean flow, but the role of transient waves should not be neglected. De Weaver and Nigam (2000) made further studies on the driving role of stationary waves in zonal mean flow anomalies. As to the climatic problems, the atmosphere is also influenced by many external forcings, such as the thermal forcing from the underlying surface. The known studies show that there are close connections between the variations of zonal mean circulation and those of the earth temperature, sea surface temperature, ice, and snow at the underlying surface (e.g., Chiang et al., 2001; Solomon and Jin, 2005).

Thompson and Wallace (1998) pointed out that

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the most prominent mode of the sea level pressure in boreal winter presents a zonally symmetrical structure and they named it Arctic Oscillation (AO) or Northern Annular Mode (NAM). Since then, it has appealed to a great many researchers as the leading mode of atmospheric circulation variability, which presents a zonally symmetrical structure on the scale of Northern Hemisphere. The importance of this mode lies in its influences on many aspects of the global climatic variations, such as the monthly average changes of climate, blocking, extreme weather events and the trends of the global climatic changes since the 1970s, which are closely related to the AO (Thompson and Solomon, 2001). Many of the studies on AO/NAM reveal that the maintenance of this mode is due to the interaction of positive feedbacks between zonal mean flow and the waves. Lorenz and Hartmann (2003) found that the zonal mean flow \([u]\) anomalies would lead to anomalous vortex motion, which would further provide momentum to maintain the initial \([u]\) anomalies. This positive-feedback mechanism implies that the annular anomalies last longer than regional ones.

ENSO, which is closely connected with the variations of global circulation, is a strong signal on the interannual time scale in the air-sea system. Hoerling et al. (1995) analyzed the boreal zonal mean flow anomalies in El Niño events, and verified the effects of the internal and external forcing on zonal mean flow using the GCM model. The results show that it is the internal atmospheric forcing that induces the mid-latitude \([u]\) anomalies, while the forcing of ocean SST leads to the \([u]\) anomalies in tropical and subtropical regions.

Normally, researchers pay more attention to the anomalies of wintertime zonal mean circulation. Ting et al. (1996) observed and studied the teleconnection pattern of the zonal mean circulation with extreme phases in boreal winters, and his analyses based on the observation indicate that the interannual oscillation of the mid-latitude zonal mean circulation has remarkable influence on the interannual changes of stationary waves in the northern extra-tropical regions. They also built up the connection between the zonal mean circulation and the climatic anomalies in extra-tropical regions through the teleconnection pattern. De Weaver and Nigam (2000) ascertained the correlation between wintertime zonal mean \([u]\) anomalies and NAO, ENSO, PNA, respectively, through the three modes obtained by doing the REOF of the 200-hPa geopotential height field. However, relatively fewer studies have been conducted on issues like what the structural features the zonal mean circulation has on the whole, how they change, and what connections they have with AO as well as ENSO. In the present study, we will apply EOF method to extract the principal components of the holistic structural variations of zonal mean circulation, and try to find out the possible factors leading to those changes. The results of the study will offer a profound understanding about the regular patterns of zonal mean circulation variances and their inter-connections with the summertime climatic changes.

2. Data and methods

Data of zonal wind velocity \((u)\), meridional wind velocity \((v)\), vertical velocity \((\omega)\), temperature \((T)\), geopotential height \((h_{gt})\), and the sea level pressure (SLP) are from NCEP/NCAR reanalysis (Kalnay et al., 1996). Resolution of the data grid is \(2.5^\circ \times 2.5^\circ\) in horizontal, 12 levels in vertical direction, from 100 hPa down to the earth’s surface, from 1979 to 2005. Data of sea surface temperature (SST) are from NCEP/NCAR reanalysis. For SST, horizontal resolution is \(2^\circ \times 2^\circ\) during 1979–2005. These data have wonderful consistency in tropical regions with those in other datasets such as HadISST.

EOF analysis, Wavelet analysis (Torrence and Compo, 1998), composite analysis, and harmonic analysis are employed as the main methods in this study.

3. Results

3.1 Principal modes of zonal mean circulation and their variations

In order to better understand the principal patterns of zonal mean flow variations, EOF decomposition is performed to extract the leading modes of the
zonal mean $[u]$ anomalies. Here, the first three eigenvectors have been verified by North Criterion (Shi, 2002). The accumulated variance contributions of the first two eigenvectors to the total variance are over 50% (Table 1), which is able to depict the summertime zonal mean circulation features of spatial changes in corresponding years.

<table>
<thead>
<tr>
<th></th>
<th>EOF1</th>
<th>EOF2</th>
<th>EOF3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Variance explained</td>
<td>31.75%</td>
<td>23.09%</td>
<td>10.54%</td>
</tr>
<tr>
<td>Accumulate variance</td>
<td>31.75%</td>
<td>54.84%</td>
<td>65.38%</td>
</tr>
</tbody>
</table>

3.1.1 The spatial distribution as revealed by EOF1 and EOF2

The typical field of EOF1 eigenvectors (31.75% of the total variance), reflects the principal spatial distribution of interannual variabilities of $[u]$ anomalies in boreal summer (Fig. 1a). In regions from the equator to the North Pole, the spatial distribution of $[u]$ anomalies displays as wave-number 1.5. There are two centers, one center of positive values and the other of negative values, lying in the Polar Region and the mid-latitudes respectively, and the largest negative value is found near the Polar Region, while weaker negatives are found in the mid-latitudes. The transition node of positive-negative anomalies is near 60°N. The EOF1 shows an equivalent barotropic structure in the troposphere.

Considering the basic states in Fig.1a and taking Fig. 1b as reference, we found that the westerlies are weakened in the high-latitude and the Polar Region, and the ones in the mid-latitude are extraordinarily strengthened in layer of 200-300 hPa at both 35° and 50°N, respectively, when the time coefficients are positive. However, when the time coefficients are negative, the westerlies are strengthened in the high-latitude and Polar Region, and the easterly anomalies appear in the mid-latitudes, weakening the westerlies in the mid-latitudes at the same time. The anomalies propagate up to the stratosphere with height rising especially at 75°N, in the 300-hPa upper troposphere. The phenomena described above are similar to

![Fig.1. EOFs of the zonally-averaged wind anomalies in boreal summer. (a), (c): EOF1 and EOF2; and (b), (d): the time coefficients of EOF1 and EOF2, respectively.](image-url)
Thompson’s (2000) sectional drawing (refer to Fig. 9 of their article) of zonal mean wind field, which is corresponding with the summer Arctic Oscillation: in the meridional direction presents as the dipole form, while in the regions above 25°N in the vertical direction presents as the equivalent barotropic structure of the same phase in both upper and lower atmosphere. Some other studies (Fan et al., 2003; Li et al., 2003) also point out that AO actually reflects the intensity of the mid-latitude westerlies, which is comparatively more obvious in boreal winter. The AO also exists in boreal summer but much weaker than that in winter. The centers of summertime AO are located more northern as compared to those of wintertime AO. Positive phases of Arctic Oscillation are displayed in years of negative time coefficients in Fig.1b whereas negative phases in years of positive time coefficients.

Figure 1c shows the EOF2 (23.09% of the total variance). Anomalous $u$ as denoted by EOF2 shows a wave-number 2 structure is in spaced from the equator to the North Pole. Two centers of positive values related to the wave-number 2 structure locate respectively in high-latitudes and tropical regions. However, the larger positive values are distributed in the tropical regions whereas the negative values are in the polar and mid-latitude regions. Note that the disturbance migrates upward to the stratosphere with the latitude change from the pole to tropical regions. EOF2 as displayed in Fig. 1c reveals mostly the disturbances in $u$ in the tropical regions other than in mid- and high latitudes.

Climatologically, the axis of westerly jet stream in boreal summer lies at 40°–45°N (figure omitted). When the time coefficients are positive (Fig. 1d), the westerlies on the southern side of the jet stream are strengthened while those on its northern side were weakened, inducing the location of the westerly jet stream to be more southern in these years than normal. On the other hand, the westerlies on the southern side of the jet are weakened while those on northern side are strengthened in years when the time coefficients are negative, inducing the location of the westerly jet stream more northern than normal. The significant negative values of time-coefficients of EOF2 appear in El Niño years such as 1982–1983, 1987, 1992–1993, and 1997, whereas the significant positive values appear in La Niña years such as 1984–1985, 1988, 1996, and 1998–2001 (Fig. 1d), which indicates the strong impacts of ENSO on the variations of the westerlies. Note that two years including 1994 and 1995 are exceptional, in which year the ENSO did not occur though the time coefficients as shown in Fig.1d are significant. However, as we know, there occurred a very typical positive Indian Ocean dipole (IOD) mode event in 1994. The exception of 1994 implies that there could be some influences of IOD on the spatial distribution of $u$ anomalies (Guan and Yang, 2003).

### 3.1.2 Periodicities

The Morlet wavelet transform has been performed for time series of coefficients of both EOF1 and EOF2.

![Fig.2.](image-url) Power spectrum analysis of the Morlet wavelet transform for the time series of (a) EOF1 and (b) EOF2. Shaded areas are significant above the 90% level of confidence, and dotted areas represent the cone of influence.
as presented in Fig. 2 to demonstrate the features of the periodic changes of zonal mean circulation.

The disturbances of zonal mean circulation, shown in EOF1, have a significant quasi-period of 3 yr during 1983–1993 (Fig. 2a). A quasi-period of 7 yr is found significant after 1990. Yang et al. (2006) studied the summertime AO, concluding that summertime AO has oscillations of periods about 6–7 and 2–3 yr. Their results look similar to our results presented here by EOF1.

The disturbances of zonal mean circulation, shown in EOF2, have a significant period of 3–6 yr (Fig. 2b), along with obvious quasi-biennial variations during 1993–1998. A lot of known studies (e.g., Lau et al., 1988) reported that ENSO has some periods not only ranging from 3 to 7 yr, but also a quasi-biennial tendency. The periodicities of EOF2 seem just consistent with periodicities of ENSO.

### 3.2 Connections with AO and ENSO

The EOF analysis aforementioned suggested that there are strong connections of $\langle u \rangle$ anomalies as shown by the two leading components, from their temporal variations and spatial distribution, with AO and ENSO.

#### 3.2.1 Connections with AO

The time-series of coefficients of EOF1 has a significant negative correlation with summer AO index available from CPC; the correlation coefficient is as high as -0.67 (Fig. 3a). The mean composite of differences of $\langle u \rangle$ anomalies between cases of the positive time coefficients of EOF1 and those of negative ones looks similar to the pattern of EOF1, showing the signature of AO.

The correlations of time-series of coefficients of EOF1 respectively with the SLP anomalies, and with the 500-hPa geopotential height show that the oscillation pattern in the Northern Hemisphere with negative values in the Polar Region and positive values in some places in mid-latitudes, displaying clearly the signature of AO (Figs. 3b and 3c). Some places outside the Polar Region seem to be very important, where the notable positive or negative correlations are found; these places are the eastern part of China, east coast of North America, and western Atlantic. A wave train can be seen from Fig. 3c, which propagates at 500 hPa from the Mediterranean Sea to Polar Region, suggesting that the zonal mean circulation be interacted with zonal disturbances. It is worth noticing that the zonal mean westerly is significantly correlated with the SLP with negative values in eastern part of China. When the coefficients are positive which corresponds to the AO’s negative phase, the eastern part of China is controlled by anomalous lower than normal pressure. When the coefficients are negative which corresponds to the AO’s positive phase, the eastern part of China is controlled by anomalous higher than normal pressure. These scenarios of SLP anomalies lead to the summer climate variations in eastern China. Gong et al. (2002) examined that AO’s variations in spring had notable impacts on the rainfall over the areas around the mid and lower reaches of the Yangtze River in the following summer. Later, Nan and Li (2005a, b) found the connections between the AAO (Antarctic Oscillation) in springs and the summer rainfall over these areas.

What we need to clarify further is that the correlations of EOF1 component with the anomalous SLP, and with the anomalous 500-hPa geopotential height, are not so high in the Pacific region although the AO signature can be seen from the correlation field between the mid-high latitudes and the Polar Region. These lower correlations suggest that the connection between the Pacific SLP, geopotential height, and the leading component EOF1 of zonal mean flow in Pacific region is not so close as in other regions. However, larger positive correlations between CPC summer AO index other than the time-series of coefficients of EOF1 and the SLP anomalies can be found in mid-latitude oceanic regions of both Pacific and Atlantic Oceans (figure omitted), suggesting that EOF1 mainly represents the ENSO-independent part in AO. Some studies so far (Hsieh et al., 2005) reported that AO and ENSO possibly connected with each other through the air-sea

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1Note: AO index is from http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/ao.shtml, and visiting time is September 8, 2007.
interaction induce circulation changes in the Pacific regions.

To further clarify the connections between EOF1 and ENSO, we have performed respectively for four seasons including spring, summer, autumn, and winter the EOF analysis for zonal mean \( \zeta \). Correlations between time series of coefficients of EOF1 and the Nino3 index with different lags are presented in Table 2. The results show that the simultaneous correlation coefficients between the EOF1 and Nino3 index with lag 0 are quite low (lower than 95% level of confidence) for anyone of the four seasons. Only does the EOF1 component of springtime \( \zeta \) significantly correlate with respectively the following autumn and winter Nino3 index. These significant correlations imply that there possibly occurs the response of the ocean to the atmospheric zonal flow variations. On the other hand, EOF1 component of springtime \( \zeta \) zonal mean flow is found to be well correlated with Nino3 index of the previous spring (summer), suggesting that there could be possible lag response of the zonal mean flow to the El Niño/La Niña events with lag of 1 yr.

### 3.2.2 Connections with ENSO

The zonal mean \( \zeta \) anomalies are closely related to ENSO. This is not only because the correlation coefficient of time series of coefficients of EOF2 with Nino3 index in summer reaches as high as 0.74 (Fig. 4a), but the pattern displayed by the significant correlations of EOF2 mode with summertime global sea-surface temperature anomalies (SSTA) also shows clearly the ENSO signature. The positive correlations appear in central and east equatorial Pacific whereas the
Table 2. Lag −1, 0, and +1 yr correlation coefficients between the time series of EOF1 of \([u]\) and Nino3 indices for four seasons

<table>
<thead>
<tr>
<th>Nino3 index of lag −1 spring</th>
<th>EOF1 of spring ([u])</th>
<th>EOF1 of summer ([u])</th>
<th>EOF1 of autumn ([u])</th>
<th>EOF1 of winter ([u])</th>
</tr>
</thead>
<tbody>
<tr>
<td>−0.54</td>
<td>0.26</td>
<td>−0.25</td>
<td>0.08</td>
<td></td>
</tr>
<tr>
<td>Nino3 index of lag −1 summer</td>
<td>−0.31</td>
<td>0.15</td>
<td>−0.39</td>
<td>0.13</td>
</tr>
<tr>
<td>Nino3 index of lag −1 autumn</td>
<td>−0.18</td>
<td>0.15</td>
<td>−0.23</td>
<td>0.20</td>
</tr>
<tr>
<td>Nino3 index of lag −1 winter</td>
<td>−0.13</td>
<td>0.08</td>
<td>−0.19</td>
<td>0.23</td>
</tr>
<tr>
<td>Nino3 index of spring</td>
<td>0.03</td>
<td>0.12</td>
<td>−0.19</td>
<td>0.02</td>
</tr>
<tr>
<td>Nino3 index of summer</td>
<td>0.28</td>
<td>−0.03</td>
<td>0.16</td>
<td>−0.04</td>
</tr>
<tr>
<td>Nino3 index of autumn</td>
<td>0.43</td>
<td>−0.06</td>
<td>0.24</td>
<td>−0.06</td>
</tr>
<tr>
<td>Nino3 index of winter</td>
<td>0.51</td>
<td>−0.13</td>
<td>0.34</td>
<td>0.09</td>
</tr>
<tr>
<td>Nino3 index of lag +1 spring</td>
<td>0.43</td>
<td>−0.28</td>
<td>−0.27</td>
<td>0.27</td>
</tr>
<tr>
<td>Nino3 index of lag +1 summer</td>
<td>0.17</td>
<td>−0.35</td>
<td>0.06</td>
<td>0.15</td>
</tr>
<tr>
<td>Nino3 index of lag +1 autumn</td>
<td>−0.07</td>
<td>−0.28</td>
<td>−0.11</td>
<td>−0.28</td>
</tr>
<tr>
<td>Nino3 index of lag +1 winter</td>
<td>−0.07</td>
<td>−0.18</td>
<td>−0.12</td>
<td>−0.01</td>
</tr>
</tbody>
</table>

Note: The bolds represent correlations above 95% level of confidence.

negative ones in the West Pacific (Fig. 4b). The mean composite differences of anomalous zonal mean flow in years when coefficients of EOF2 are positive (positive phases of time-series of coefficients) from its value in years when the coefficients are negative (negative phases of time-series of coefficients) show a similar pattern as displayed by EOF2. The composites for the positive phases are related to El Niño events while those for negative phases are related to La Niña events (figure omitted).

The significant positive correlations of EOF2 with anomalous 200-hPa geopotential height are found in the entire tropical zone with the center locating mainly in the central east equatorial Pacific. The significant negative correlations are found in area 30°–50°N (Fig. 4c). These results indicate that the increases (decreases) of 200-hPa geopotential height occur in years when the coefficients of EOF2 are positive (negative), which are found to be related to El Niño (La Niña) years. On the other hand, in the extratropical region (mid-latitudes) there appear apparently the lower than normal geopotential heights in the Northern Hemisphere (Fig. 4c). This oppositely signed distribution of the correlation coefficients between tropical and extratropical regions for boreal summer shows a scenario consistent with the results by De Weaver and Nigam (2000) though his study was for boreal winters, in which season the intensity and distribution of the zonal mean flow is different from those in summer season. This pattern of zonal flow related to EOF2 is probably associated with the intensification of subtropical jet that induces the ENSO events (Hoerling et al., 1995).

The series of coefficients of EOF2 of a season as shown in Table 3 is significantly correlated with Nino3 indices of this season or another. Particularly, the series of coefficients of EOF2 of summer is highly correlated with Nino3 indices of anyone of the four seasons (simultaneously, 0.75), implying the possible influence of the Nino3 of seasons before summer on the zonal mean flow and the possible influence of this flow on the Nino3 SSTA of the following seasons after summer. Note that no significant correlations can be found in Table 3 when the lags between two time-series are larger than 3 seasons. The time-series of coefficients of EOF2 of both spring and summer are found to be well correlated with the Nino3 indices of previous winter.

We have also noticed that no significant simultaneous correlations can be found between AO index and the EOF2 mode of a season out of the four seasons but winter. The simultaneous correlations of winter zonal mean flow anomalies represented by mode EOF2 with winter AO index is significant (0.55, Table 4).

The results above indicate that the EOF2-related anomalies of summertime zonal mean flow \([u]\) mainly has significant responses to ENSO (Table 3), with the largest variances in subtropical region (Fig. 1c). Ding and Wang (2005) conducted a research on Circum-Global Teleconnection type (CGT) in mid-latitudes in boreal summers, showing that summer CGT is
related to the second mode of the interannual variations of upper troposphere circulation and the CGT is correlated significantly with ENSO but without AO. Their result is somewhat similar to outcomes of our EOF analysis in the present work. What we need to notice more is that, for the boreal winter, the first EOF mode can effectively extract part signatures of AO. However, although EOF2 characterizes part of ENSO signatures in winter season, a portion of AO signatures has not been fully eliminated from the EOF2 mode, which leads to some significant correlations of EOF2 with AO index as seen in Table 4. This may be related to the fact that winter AO is affected by ENSO (Pozo-Vazquez et al., 2001).

### 3.3 Possible mechanism of maintenance of two principal modes

The divergence and convergence of E-P fluxes exhibit the forcing of waves acting on the zonal mean flows. Figure 5a shows that the E-P fluxes propagate upward till the upper troposphere in high latitudes where these E-P fluxes meet with those fluxes that propagate along the waveguide from the near tropopause in mid-latitudes poleward and downward to this region of upper troposphere, inducing convergence of E-P fluxes over 60°–80°N. This convergence weakens the westerly in regions from high latitudes to the polar area. The convergence center at 70°N at

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**Table 3.** Lag –1, 0, and +1 yr correlation coefficients between the time series of coefficients of EOF2 of \([u]\) and Nino3 indices of four seasons

<table>
<thead>
<tr>
<th>Nino3 index of lag</th>
<th>EOF2 of spring ([u])</th>
<th>EOF2 of summer ([u])</th>
<th>EOF2 of autumn ([u])</th>
<th>EOF2 of winter ([u])</th>
</tr>
</thead>
<tbody>
<tr>
<td>lag –1 spring</td>
<td>0.02</td>
<td>–0.28</td>
<td>–0.28</td>
<td>0.30</td>
</tr>
<tr>
<td>lag –1 summer</td>
<td>0.35</td>
<td>–0.03</td>
<td>–0.03</td>
<td>0.19</td>
</tr>
<tr>
<td>lag –1 autumn</td>
<td>0.32</td>
<td>0.20</td>
<td>–0.16</td>
<td>0.14</td>
</tr>
<tr>
<td>lag –1 winter</td>
<td><strong>0.45</strong></td>
<td><strong>0.38</strong></td>
<td>–0.23</td>
<td>0.09</td>
</tr>
<tr>
<td>spring</td>
<td><strong>0.45</strong></td>
<td><strong>0.55</strong></td>
<td><strong>0.55</strong></td>
<td>–0.01</td>
</tr>
<tr>
<td>summer</td>
<td>0.21</td>
<td><strong>0.75</strong></td>
<td><strong>0.75</strong></td>
<td><strong>–0.56</strong></td>
</tr>
<tr>
<td>autumn</td>
<td>0.07</td>
<td><strong>0.59</strong></td>
<td>–0.01</td>
<td><strong>–0.56</strong></td>
</tr>
<tr>
<td>winter</td>
<td>0.05</td>
<td><strong>0.55</strong></td>
<td>–0.12</td>
<td><strong>–0.62</strong></td>
</tr>
<tr>
<td>lag +1 spring</td>
<td>0.09</td>
<td>0.33</td>
<td>0.33</td>
<td>–0.36</td>
</tr>
<tr>
<td>lag +1 summer</td>
<td>0.14</td>
<td>–0.11</td>
<td>–0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>lag +1 autumn</td>
<td>0.12</td>
<td>–0.13</td>
<td>0.02</td>
<td>–0.03</td>
</tr>
<tr>
<td>lag +1 winter</td>
<td>0.07</td>
<td>–0.18</td>
<td>0.06</td>
<td>–0.01</td>
</tr>
</tbody>
</table>

Note: The bolds represent correlations above 95% level of confidence.

**Table 4.** Lag –1, 0, and +1 yr correlation coefficients between the time series of EOF2 of \([u]\) and AO indices of four seasons

<table>
<thead>
<tr>
<th>AO index of lag</th>
<th>EOF2 of spring ([u])</th>
<th>EOF2 of summer ([u])</th>
<th>EOF2 of autumn ([u])</th>
<th>EOF2 of winter ([u])</th>
</tr>
</thead>
<tbody>
<tr>
<td>lag –1 spring</td>
<td>0.28</td>
<td>0.32</td>
<td>–0.20</td>
<td>0.03</td>
</tr>
<tr>
<td>lag –1 summer</td>
<td>0.28</td>
<td>0.26</td>
<td>–0.09</td>
<td><strong>–0.52</strong></td>
</tr>
<tr>
<td>lag –1 autumn</td>
<td>0.38</td>
<td>0.27</td>
<td>0.21</td>
<td>–0.01</td>
</tr>
<tr>
<td>lag –1 winter</td>
<td>0.16</td>
<td>0.20</td>
<td>–0.18</td>
<td>0.17</td>
</tr>
<tr>
<td>spring</td>
<td>0.02</td>
<td>0.03</td>
<td><strong>0.47</strong></td>
<td>–0.17</td>
</tr>
<tr>
<td>summer</td>
<td>0.12</td>
<td>–0.31</td>
<td>0.01</td>
<td>0.17</td>
</tr>
<tr>
<td>autumn</td>
<td>–0.01</td>
<td>–0.09</td>
<td>–0.20</td>
<td>0.27</td>
</tr>
<tr>
<td>winter</td>
<td>0.09</td>
<td>–0.06</td>
<td>–0.24</td>
<td><strong>0.55</strong></td>
</tr>
<tr>
<td>lag +1 spring</td>
<td>–0.14</td>
<td>–0.28</td>
<td>0.30</td>
<td><strong>0.41</strong></td>
</tr>
<tr>
<td>lag +1 summer</td>
<td>0.10</td>
<td>0.19</td>
<td>–0.04</td>
<td>0.18</td>
</tr>
<tr>
<td>lag +1 autumn</td>
<td>–0.09</td>
<td>–0.02</td>
<td>0.11</td>
<td>0.31</td>
</tr>
<tr>
<td>lag +1 winter</td>
<td>0.16</td>
<td>0.31</td>
<td>0.07</td>
<td>0.16</td>
</tr>
</tbody>
</table>

Note: The bolds represent the correlations above 95% level of confidence.
300 hPa basically coincides with the easterly anomalies of EOF1 as shown in Fig. 1a. The E-P flux divergence with its center at 35°N at 200 hPa is found near the tropopause in the mid-latitude, strengthening the mid-latitude westerlies, corresponding with the equatorward center of westerly anomalies represented by EOF1 (Fig. 1a). The positive-negative distribution of E-P flux divergence seems roughly to be consistent with the EOF1 pattern of anomalous \[ u \] in mid and high latitudes, which is supported by the quasi-geostrophic non-acceleration theorem by Wu (1988) as described in Appendix of this paper.

The composite quantities including the E-P fluxes and their divergence as derived from components of waves with zonal wavenumbers 1–4 are useful for our understanding the role played by the summertime planetary waves in maintenance of the aforementioned principal modes of zonal mean flow (Fig.5d). The locations of the convergence and divergence centers of E-P flux of planetary waves are almost the same as those in the filed of E-P flux of total eddies although the divergences induced by planetary waves are a little bit weaker in intensity as compared to those induced by total eddies. These scenarios show that the AO which is represented by EOF1 is apparently related to the forcing of planetary waves. Anomalies of mean zonal flow take place when planetary waves propagate in the three-dimensional atmosphere, inevitably leading to the convergence and divergence of E-P fluxes, thereby, causing the interaction between waves and zonal mean flows (Huang and Zou, 1989). Chen and Huang (2005) reported that the oscillation between the two waveguides of quasi-stationary planetary waves was intrinsically related to the AO. If the planetary waves in the troposphere tend to propagate toward mid and low latitudes, will the activities of these planetary waves that propagate into the lower stratosphere from the troposphere where the waves are refracted upward and poleward via the polar waveguide be weakened. These weakened poleward propagations of planetary waves have less than normal influences on the polar vortex, facilitating the polar vortex to be stronger and more...
Fig. 5. Composite differences between high and low value years of the time series of EOF1. (a) E–P flux (vector) and its divergence $\nabla \cdot E$ (contour); (b) residual circulation $\text{acos} \varphi f_0 \vec{v}$; (c) sum of the two terms $\nabla \cdot E + \text{acos} \varphi f_0 \vec{v}$; and (d) E–P flux (vector) induced by planetary waves (zonal wavenumbers 1–4) and its divergence (contour).

stable, and consequently strengthening the westerlies around the Polar Region. However, if more planetary waves propagate toward higher latitudes, the polar vortex will be perturbed more, which results in weakening the polar vortex and henceforth weakening the westerlies around the polar vortex. Baldwin and Thompson (2003) reported that the variations of positive and negative phases of AO are closely related to the variations in intensity of the stratospheric polar vortex.

Kodera and Kuroda (2000) found that there exist the intra-seasonal differences of the AO. Based on whether the troposphere couples with the stratosphere, they divided the wintertime AO into two types of oscillations, i.e., the troposphere type (T-type) and stratosphere type (S-type). Generally, the T-type is much more likely to occur in early winter (Nov.–Dec.), but the S-type in late winter to early spring (Feb.–Mar.). The apparent differences between the two types are found in the vertical section of wind field. The vertical shear of wind is found near the tropopause in the T-type AO, and the wind anomaly center is located in the upper troposphere, while an equivalent barotropic structure from troposphere to stratosphere is observed in the S-type AO, and the wind anomaly center, which is stronger in intensity than that in T-type lies in the stratosphere. However, because the AO in summer is apparently weaker than that in winter, the turbulence center of zonal wind field is usually located in mid-high levels of troposphere. With help
from the E-P flux technique, they found that the planetary waves in their upward propagation processes in high-latitudes are intercepted by the wave guide that propagates poleward and downward from the upper level of subtropical stratosphere, which prevents the wave activities from propagating upward from troposphere into stratosphere and stops the coupling between troposphere and stratosphere. The results by Kodera and Kuroda (2000) and our results as mentioned above suggest that the summertime AO would be mostly the T-type one. Because our present study focuses only on the troposphere, the zonal mean flow variations in the stratosphere need further investigations.

The residual circulation (Fig. 5b) shows an oppositely signed scenario as compared to that of E-P flux divergence (Fig. 5a). The values of the residual circulation in middle latitudes are also oppositely signed as compared to those in high latitudes. The E-P flux divergence term is found to balance the residual circulation. The sum of E-P flux divergence and residual circulation (Fig. 5c) indicates that the dissipation due to friction in the upper troposphere in mid and high latitudes is relatively small, which suggests that the balance between E-P flux divergence and residual circulation maintains the AO. However, the stronger friction dissipation in the lower troposphere and regions near the equator are found, which is possibly due to the thermal forcing from the underlying surface of the earth. Zhao and Taleahash (2005) calculated the zonally symmetric normal mode by using the primitive atmospheric dynamical equations in a spherical $\rho$-coordinate system. They found that the zonally symmetric normal mode has a spatial structure similar to AO, suggesting that the maintenance of this physical mode AO possibly resulted from the nonlinear interactions among oscillations corresponding with the normal modes.

Fig. 6. As in Fig. 5, but for EOF2.
The locations of E-P flux divergence centers as displayed in Fig. 6a are mostly close to the center locations of EOF2 as shown in Fig. 1c though some differences in distribution can be found in between Figs. 1c and 6a. A divergence center at 30°N at 150 hPa is found in Fig. 6a, where the westerlies are strengthened as shown in Fig. 1c. The convergence of E-P fluxes at 40°–60°N is found, where the westerlies are weakened, inducing the westerly jet in El Niño year to locate more southern. It can also be seen that the distributions of E-P fluxes and their divergence in Fig. 6a are quite similar to those in Fig. 6d, though some small differences in values, indicating the dominant role played by the planetary waves in maintaining the \( u \) anomalies. In addition, the friction dissipation is relatively stronger in regions around the equator and in the lower troposphere in the mid-latitudes south of 60°N.

4. Conclusions and discussion

The NCEP/NCAR reanalysis data are employed in the present study to investigate the principal modes of the mean zonal wind in boreal summer for our better understanding the mechanisms of the interannual variations in atmospheric general circulations in the Northern Hemisphere. Our conclusions may be drawn as follows:

(1) Two principal modes of zonal mean \( u \) anomalies in boreal summer have been extracted by performing the EOF analysis. EOF1 displays the AO-related distribution of the \( u \) anomalies, while EOF2 does the ENSO-related one. Periodicities in EOF1 and EOF2 modes are very similar to those of AO and ENSO, respectively. The lag correlations of Nino3 index with respectively the series of time coefficients of EOF1 (EOF2) have been calculated for four seasons. The simultaneous correlation between EOF1 and ENSO is not found significantly. EOF1 are hence believed as a mode that relates to AO but ENSO. The simultaneous correlation between EOF2 and Nino3 index is found significant, particularly in summer. The EOF2 mode shows the response of summertime zonal mean circulation to ENSO. The strongest response is found in subtropics. Besides, lag correlations of AO index with series of time coefficients of EOF2 for four seasons show that no significant correlations between EOF2 and AO index are found in the seasons including spring, summer, and autumn, but winter.

(2) The probable mechanism of maintaining \( u \) anomalies are discussed in terms of the quasigeostrophic non-acceleration theorem. The pattern as displayed by composite differences of E-P flux divergence in years with bigger values of series of time coefficients of EOF1 (EOF2) from that in years with smaller values of the series of time coefficients of EOF1 (EOF2) matches well with EOF1 (EOF2), indicating the importance of the impacts of waves on the zonal mean flows. The planetary waves as obtained using harmonic wave analysis dominate the spatial-temporal changes of zonal mean flows in boreal summer. The residual circulation is almost oppositely signed as the E-P flux divergence is. Values of the residual circulation are comparable to those of the divergence of E-P fluxes. The residual circulation cancels the effects of the planetary wave activities. However, it is noted that the friction dissipation is also to some extent important in tropics and in the lower troposphere in middle and high latitudes.

In the present paper, only the principal modes of anomalous zonal mean flow are studied. Further investigations need to be carried out in future on the detailed processes and mechanisms of maintaining the zonal mean flow.

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REFERENCES


APPENDIX

Quasi-geostrophic non-acceleration theorem (Wu, 1988) can be described as follows:

Under the quasi-geostrophic approximation, momentum equation, thermal wind equation, continuity equation, and thermodynamical equilibrium equation in a spherical \( p \)-coordinate system could be written as

\[
\begin{align*}
[u]_t &= f_0 \vec{v} + (a \cos \theta)^{-1} \nabla \cdot \mathbf{E} + F, \\
f_0[u]_p - a^{-1} R^p \theta &= 0, \\
(\cos \theta)^{-1}(\bar{v} \cos \theta)_\phi + (\bar{\omega})_p &= 0, \\
|\theta|_p + \theta_p \bar{\omega} &= Q.
\end{align*}
\]

\[ \text{(A1)} \]

\[ \text{(A2)} \]

\[ \text{(A3)} \]

\[ \text{(A4)} \]

In the above equations, \( F \) and \( Q \) represent friction and diabatic heating, respectively. \( \theta \) represents potential temperature, and \( \Theta \) the area-mean of \( \theta \). “[ ]” means zonal mean, while “*” represents zonal-mean deviation. \( E \) represents E-P flux, \( (\bar{v}, \bar{\omega}) \) residual circulation. We have

\[
E = (E(\varphi), E(p)) = a \cos \varphi \left( -[u^* v^*], f_0 \Theta_p^{-1} [v^* \theta^*] \right), \\
\n\n\text{\text{\text{\text{\text{\text{(A5)}}}}}}
\]

\[
\nabla \cdot E = \frac{1}{a \cos \varphi} \frac{\partial}{\partial \varphi} (E(\varphi) \cos \varphi) + \frac{\partial}{\partial p} (E(p)). \\
\n\n\text{\text{\text{\text{\text{\text{(A6)}}}}}}
\]

\[
\begin{align*}
\bar{v} &= [v] - (\Theta_p^{-1} [v^* \theta^*])_p, \\
\bar{\omega} &= [\omega] + (a \cos \varphi)^{-1} (\Theta_p^{-1} [v^* \theta^*] \cos \varphi)_\varphi.
\end{align*}
\]

\[ \text{\text{\text{\text{\text{\text{(A7)}}}}}} \]

Multiplying both sides of Eq.(A1) by \( a \cos \varphi \), we obtain

\[
a \cos \varphi [u]_t = a \cos \varphi \cdot f_0 \bar{v} + \nabla \cdot E + F'. \\
\]

\[ \text{\text{\text{\text{\text{\text{(A8)}}}}}} \]

To clarify the reason of zonal mean flow variations, we have calculated \( a \cos \varphi [u]_t, \nabla \cdot E, a \cos \varphi f_0 \bar{v}, \) and \( \nabla \cdot E + a \cos \varphi f_0 \bar{v} \) month by month, respectively. Our results show that the magnitude of E-P flux divergence and residual circulation are equivalent, whereas the magnitude of \( a \cos \varphi [u]_t \) is smaller. The maintenance of zonal-mean flow anomalies is therefore believed to be as a result of the balance among E-P flux, residual circulation, and friction.