Variability of the polar night jet in the Northern and Southern Hemispheres

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Abstract. The spatial and temporal characteristics of the month-to-month variability of the polar night jet and its relationship with tropospheric circulation is investigated for both the Northern and Southern Hemispheres. The variabilities of the hemispheres have many common characteristics of the Polar Night Jet Oscillation (PJO). These common characteristics include the following: (1) the anomalous zonal-mean zonal winds shift poleward and downward; (2) the anomalous polar temperatures propagate downward from the stratosphere to the upper troposphere; and (3) they are made through a wave-mean flow interaction with mainly the planetary wave of zonal wave number one. Annular modes associated with the PJOs appear in both hemispheres when the zones of maximum polar temperature anomaly descend to the lowermost stratosphere and upper troposphere. The major difference in the PJOs of the two hemispheres is found in their temporal characteristics. In the Southern Hemisphere, the phase of the PJO is closely locked to the annual cycle, while in the Northern Hemisphere it exhibits quasiperiodic variability with its envelope controlled by the annual cycle. The origin of the differences between the PJOs is discussed based on the theory of the wave-mean flow interaction.

1. Introduction

Stratospheric circulation exhibits large month-to-month variabilities in accordance with the occurrence of stratospheric warming in the Northern Hemisphere (NH) winter [e.g., Labitzke, 1982]. However, this intraseasonal variability does not occur randomly but has a temporal and spatial structure [e.g., Perlwitz and Graf, 1995; Perlwitz et al., 2000; Baldwin et al., 1994]. In particular, zonal-mean zonal wind anomalies propagate poleward and downward to the troposphere in turn from the stratosphere region [Kodera, 1995; Kuroda and Kodera, 1999] (hereinafter referred to as KK99). The periodic nature of this variation can be more clearly seen in general circulation model (GCM) simulations under perpetual winter conditions [Christiansen, 1999; Yamazaki and Shinya, 1999].

In contrast, stationary planetary wave activity is very weak in the Southern Hemisphere (SH), and major stratospheric warming does not occur during midwinter, but minor warmings generally take place in spring at an altitude that descends over time [Farrara and Mechoso, 1986]. In spite of the large differences in stratospheric variability between the two hemispheres [Shiotani and Hirota, 1985], a similar poleward and downward propagation of zonal wind anomalies is found in the SH [Kuroda and Kodera, 1998] (hereinafter referred to as KK98). For convenience, we designate such structured slow variability in the polar night jet in both hemispheres as the “Polar Night Jet Oscillation” (PJO). However, as will be described later, there are also large differences in the PJOs in the two hemispheres. For example, in the SH, the phase of the PJO is closely locked to the annual cycle, while in the NH, only the envelope of the PJO is controlled by the annual cycle.

The aim of the present study is to clarify, through a detailed comparison, whether the phenomenological similarity between the PJOs in both hemispheres is produced from essentially similar processes and whether their different appearances are due to conflicting circumstances, such as a difference in the planetary wave forcing. Another interesting aspect of the PJO is that in the NH the Arctic Oscillation (AO) (or the NH annular mode) [Thompson and Wallace, 1998, 2000] preferentially occurs at a certain period of the PJO (KK99). In the present study, we also investigate the relationship between the PJO and the Antarctic Oscillation (AAO), a counterpart of the AO.

This paper is organized as follows. The data set and the principal method of analysis are described in section 2. Section 3 provides the results of analyses in both hemispheres on (1) the spatial structure of the PJO, (2) the relationship with the troposphere, and (3) the temporal characteristics of the stratospheric variability. After a discussion in section 4, a conclusion is offered in section 5.

2. Data and Method of Principal Analysis

The stratosphere and troposphere data used in the present study are updated from KK99 and cover 20 years, from January 1979 to December 1998. The original stratosphere data were analyzed by U.S. National Centers for Environmental Prediction (NCEP)/Climate Prediction Center (CPC) (formerly NMC/CAC). The stratospheric winds were calcu-
lated from a satellite-derived geopotential height analyzed by CPC using the nonlinear balanced wind relation [Randel, 1992]. The stratospheric data of the NH up to 10-hPa levels from April 1996 to April 1997 were completely missing at NCEP. The missing data for this period are compensated by those derived from the Met Office, UK geopotential height data [Bailey et al., 1993], after being adjusted for being continuously connected at the equator to the SH counterpart of the NCEP data. Operational data for the troposphere at 100 hPa and below are replaced by reanalysis data of the NCEP/National Center for Atmospheric Research (NCAR) [Kalnay et al., 1996].

All missing data, except for the above mentioned substantial missing period, were linearly interpolated in time, and the monthly mean data were then calculated as 30-day mean data, with the exception of July, which is 35-day mean data. Note that the Eliassen-Palm (E-P) flux, residual velocity, and wave-activity flux by Plumb [1985] were calculated from 5-day mean data and the monthly average was taken thereafter. The monthly mean E-P flux calculated from the daily mean NCEP/NCAR data below 10 hPa was also used as a supplement to investigate the high-frequency transient eddy component.

Different analysis methods were applied to each hemisphere in previous studies. These methods included the multiple empirical orthogonal function (M-EOF) analysis to the SH (KK98) and the extended singular value decomposition (E-SVD) analysis to the NH (KK99). A recent study [Kodera et al., 2000] revealed that the stratospheric variability in the NH can endure for a very long time during the cold season, as in the SH. Therefore the same E-SVD analysis is applied for both hemispheres to extract the variabilities that extend over a 5-month period in the present study.

For this purpose, a SVD analysis [Bretherton et al., 1992] is applied to the extended vectors, which incorporate lagged vectors into the original one, to extract a propagating phenomenon with time. This technique is a direct extension of E-EOF analysis [Weare and Nasstrom, 1982]. The E-SVD analysis is calculated based on extended vectors \( \mathbf{x}_i \) and \( \mathbf{z}_i \) as

\[
\begin{align*}
\mathbf{x}_i &= [\mathbf{u}^i(m_0 + t), \mathbf{u}^i(m_0 + t + 1), \cdots, \mathbf{u}^i(m_0 + t + 4)] \\
\mathbf{z}_i &= [\mathbf{e}^i(m_0 + t), \mathbf{e}^i(m_0 + t + 1), \cdots, \mathbf{e}^i(m_0 + t + 4)]
\end{align*}
\]

where \( \mathbf{u}^i(m) \) and \( \mathbf{e}^i(m) \) are the anomalous monthly mean zonal-mean zonal wind and the vertical component of the E-P flux, respectively, at each height and latitude for the \( m \)th month of the \( i \)th year. The field at the month of \( m_0 + t + k \) is called the lag \( k \) month component. To treat the stratosphere and troposphere equally, all vectors are normalized by their standard deviations of the year-to-year variabilities of respective months prior to the calculation of the cross-covariance matrix to perform SVD. Two sets of vectors corresponding to \( t = 0 \) and \( t = 1 \) are used in the present analysis for each cold season; November (June) was selected as \( m_0 \) for the NH (SH), and so the vector for \( t = 0 \) corresponds to the five consecutive months from November to March (June to October), and that for \( t = 1 \), to December to April (July to November). The present calculation of the SVD of the large cross-covariance matrix due to extended components can be efficiently performed by the method of Kuroda [1998].

Most of the analyses in the present study were conducted based on the monthly mean data. The analyses of the zonal-mean zonal wind and vertical component of the E-P flux were directly performed by the E-SVD. Other variables were regressed onto the “developing coefficients of the zonal wind of the leading E-SVD” (hereinafter denoted as the PJO index) to study related variability. Similarly, lagging components outside of the original definition of the E-SVD, such as lag -1 month, were also calculated by a regression onto the PJO index.

3. Results

3.1. Spatial Structure

Figures 1a and 1b show the leading E-SVD in the Northern and Southern Hemispheres. They are displayed as heterogeneous regression maps of the zonal-mean zonal wind and the E-P flux. Note that the meridional component of the E-P flux is calculated by regression of the zonal-mean zonal wind expansion coefficient. The lag -1 month component was also calculated by regression (hereinafter, the words lag -1 month are denoted by lag -1 for simplicity). The name of the month corresponding to each lagged component is indicated by initials. Regions where the heterogeneous correlation coefficients exceeded the 95% significance level (0.32 for the NH for 19 winters and 0.31 for the SH for 20 winters) are shaded.

The leading mode in the NH explains 28% of the total squared covariance fractions and is comparable to the second one (explaining 23%). However, the spatial patterns of the second mode are very similar to those of the 1-month-delayed components of the leading mode. This condition reflects the periodic nature of the PJO in the NH, as will be shown later. In contrast, the leading mode (explaining 57%) in the SH dominates the second one (explaining 10%).

The present results reproduce major features of previous studies; in the NH, three consecutive months, lag 0 to lag 2, correspond to the results in KK99. In the SH, lag -1 to lag 3 components correspond to the 2-month average of the May to October pattern in KK98, due to the use of two-set vectors in the present analysis. This ensures that the result is not sensitive to a small change in a parameter or the method of analysis.

In the NH (Figure 1a), the negative zonal wind anomaly first appears in the subtropical upper stratosphere in association with an enhanced upward propagation of anomalous E-P flux from the high-latitude troposphere (lag -1 to 0). The negative wind anomalies shift toward high latitude while developing (lag 1) and then move poleward and downward to the lower stratosphere, giving their place to positive anomalies in the subtropical stratosphere (lag 2). Positive anomalies also shift poleward and downward, following the negative anomalies (lags 3 to 4). A close relationship between the anomalous E-P fluxes and zonal winds was observed during the whole sequence; the anomalous E-P flux was directed toward the region of negative anomalies of zonal wind. This happens because the region of negative wind anomaly coincides approximately with the region of anomalous E-P flux.
convergence for this slow variability, as will be shown later. The meridional propagation of the E-P flux in the troposphere is also enhanced at the stage when the meridional dipole structure is formed in the anomalous zonal wind field (lags 1 and 2).

In the SH (Figure 1b), negative zonal wind anomalies appear in the subtropics in association with an enhanced upward propagation of anomalous E-P flux, similar to the NH (lags 1 to 0). However, the negative anomalies do not shift poleward. This dipole pattern persists during the winter (lag 1 to 2). In spring, negative wind anomalies slowly shift poleward and downward and positive anomalies gradually disappear from the stratosphere (lags 3 to 4).

Positive zonal wind anomalies appear in the subtropical upper stratosphere in the NH when the vertical component of anomalous E-P flux changes sign from positive to negative (lag 2). In contrast, vertical component of anomalous E-P fluxes in the SH are positive throughout the sequence. However, it must be noted that the latitudes of enhanced vertical propagation in the lower stratosphere gradually shift poleward in association with the enlargement of the negative wind anomalies. Changes in meridional propagation of the tropospheric E-P flux are not evident in the SH, except for the high-frequency component, which will be described below.

To gain an insight into the origin of the zonal-mean zonal wind variability, associated accelerations of zonal-mean zonal wind due to the divergence of E-P flux (wave forcings) were calculated by regression onto the PJO index (Figure 2). It can be seen that the positive and negative anomalies of the wave forcing in the Northern and Southern Hemispheres correspond well to those of the zonal-mean zonal wind tendencies, except for the accelerations of the zonal winds in the high latitudes of the SH at lag 0. This overall similarity between wave forcing and the zonal wind tendency indicates that the variability of PJO is made through wave-mean flow interaction. The zonal wind acceleration in the high latitude of SH at lag 0 is considered to be due to Coriolis forcing, although other possibilities such as unresolved gravity wave exist. The wave forcings and the E-P flux due to zonal wave number 1 and 2 components are shown in Figure 3. In the NH, the main contribution comes from the zonal wave number 1 component, as documented by Kodera et al. [2000]. The zonal wave number 2 component contributes little to the total wave forcing in the upper stratosphere, but it is relatively significant in the middle stratosphere and below. For example, the poleward propagating wave in the troposphere in lag 2 of the NH comes mainly from the zonal wave number 2 component. In the SH, the zonal wave number 1 component is dominant, and the zonal wave number 2 component contributes to the lower part of the area of total wave forcing, but it is relatively small.

Figure 4 presents regression maps of the anomalous residual circulation and zonal mean temperature. The residual velocities were calculated following the method by Soel and Yamazaki [1999]; transformed Eulerian equations are iteratively solved with the upper boundary condition of no vertical motion at the 0.5-hPa level. This upper boundary condition is appropriate for all the seasons in both hemispheres except for the polar area in the SH in June and July [Haynes et al., 1991]. The areas of downward and upward motion of anomalous residual circulation in both hemispheres corresponded well to the high- and low-temperature anomalies.
The largest temperature anomalies in the NH were found in the polar region. Positive and negative anomalies descend in turn (Figure 4a). There is a tendency for temperature anomalies in the subtropics to be opposite to those in the polar region. (In particular, a quadrupole structure of the temperature anomaly was noted at lag 2 of the NH.)

In the SH, high-temperature anomalies are first created in the midlatitudes of the middle stratosphere in early winter. In contrast with the NH, downward velocity is larger in the midlatitudes than in the polar region owing to the stronger convergence of E-P flux in the lower latitudes (lag -1 to 0). The warm area gradually shifts poleward (lags 1 and 2) and downward (lags 3 and 4) in accordance with the poleward shift of the convergence zone of E-P flux. Low-temperature anomalies can be seen at the equatorial side of the high-temperature anomalies, similar to the NH; however, the overlying lower-temperature anomalies in the polar region at 1 hPa descend only a short distance.

A time-height cross section of the anomalous polar temperatures (averaged poleward of 80°) illustrates well the different characteristics of the temperature variations associated with the PJOs (Figure 5, top panels). In the NH, positive and negative anomalies propagate downward in turn, while interannual variability is very large and positive or negative anomaly propagates downward only once a year in the SH. This structure of temperature variabilities can be obtained from a simple height-time cross section of the polar temperature anomalies of individual winters. For example, Figures 6a and 7a show 30-day running mean northern and southern polar temperature anomalies for three winters from 1986 to 1989. Successive downward propagation of positive and negative anomalies are found in NH winters, while this occurs only once per year in the SH.

3.2. Relation With the Troposphere

Two types of interactions between the troposphere and stratosphere are noted in KK99. One is the response of the stratosphere to changes in the vertical propagation of tropo-
spheric waves. This situation was found in both hemispheres at similar periods of lags -1 and 0 in the present analysis (Figure 1).

The other interaction is a formation of the annular mode (or the AO in the NH) in the troposphere in association with a change in the stratospheric polar vortex. Figure 5 (bottom panels) gives the correlation coefficients between the PJO indices and time coefficients of the annular mode (annular mode indices). The annular modes are defined here as the leading EOF of the month-to-month variability of the 850-hPa levels in both hemispheres in the present data set for 1979 to 1998; they are essentially the same as those obtained by Thompson and Wallace [1998, 2000]. The correlation coefficients with the annular mode indices become largely negative when the warming signal reaches the lower stratosphere in the hemispheres (lag 1 for NH and lag 4 for SH). This period (hereinafter denoted as the annular period) corresponds in both hemispheres to the period of formation of dipole anomalies in the tropospheric zonal-mean zonal wind, with a node around 45° latitude (Figure 1). In the NH, changes in meridional propagation of E-P flux occur concurrently with the formation of the dipole structure in the troposphere. However, this change is not found in the SH in lag 4, shown in Figure 1. However, it must be remembered

Figure 4. Regression map of the zonal-mean temperature (contour) and the residual velocity (vector) associated with the PJO in the (a) Northern and (b) Southern Hemispheres. The numbers above the panels indicate the lag in months, and the initials represent the corresponding months. Shading denotes regions where the correlations of the temperatures are significant at the 95% level. Contour interval is 1 K.

Figure 5. Same as Figure 4, except for the regression map of the polar temperatures (upper panels), displayed as a function of time and height. Contour interval is 1 K. Correlation coefficients between the PJO and the AO or AAO indices are displayed in the bottom panels.
that the E-P flux depicted in Figure 1 was calculated from 5-day mean data.

Figure 8 shows a regression map of the high-frequency transient component of an anomalous monthly mean E-P flux, defined as the difference between those calculated from the daily mean and from the 5-day mean data. The figures display only three components of lags -1, 0, and 1 for the NH and -1, 3, and 4 for the SH. It can be seen that in the SH the anomalous E-P flux in the midlatitudes of the troposphere is directed toward the region of negative anomalies of the zonal-mean zonal wind. In the NH, meridional propagation is also found in the high-frequency transient component at

Figure 6. (a) Thirty-day running averaged anomalous polar temperature in the Northern Hemisphere from 1987 to 1989. (b) Interannual standard deviation of monthly mean polar temperature in the Northern Hemisphere. Contour interval is 5 K in Figure 6a and 1 K in Figure 6b. Shading indicates negative values in Figure 6a, and values larger than 4 K in Figure 6b.

Figure 7. Same as Figure 6 except for the Southern Hemisphere from 1986 to 1988, and the contour interval is 3 K for Figure 7a.
lag 1, but the contribution of the stationary component is dominant (compare lag 1 of Figure 1 and Figure 8).

A regression map of 500-hPa geopotential height was calculated to investigate more closely the relationship between the PJO and the tropospheric circulation (Figure 9). The associated wave activity flux of Plumb [1985] is also shown by arrows. Since our focus here is on planetary waves, the wave-activity flux was calculated after truncation at zonal wave number 5. Note that zonal averaging of the wave activity flux reproduces the E-P flux meridional section (Figure 1). The numbers below the respective panels indicate the correlation coefficients between the PJO and the annular mode indices. The geopotential height fields in both hemispheres exhibited wave trains at lag -1 when the upward anomalous E-P flux was still confined mainly in the troposphere and lower stratosphere. It can be seen that height anomalies around Scandinavia or the Ross and Amundsen Seas accompany an enhanced upward propagation of waves (Figure 10). It should be noted that the areas of enhanced upward propagation of the waves are located at the areas of significant heat transport (i.e., east of the trough and west of the ridge at lag -1 in Figure 9). The contribution of the stationary eddy

![Figure 8. Same as Figure 1, except for the regression maps of the E-P flux (vector) of high-frequency transient waves for selected lag components (see text).](image)

![Figure 9. Regression map of the geopotential height of the 500-hPa level (contour) and the horizontal component of the wave activity flux by Plumb [1985] (vector) associated with the PJO for selected lag components. Contour interval is 10 m, and the dashed lines indicate negative values. Shading denotes regions where the correlations of geopotential heights are significant at the 95% level. Numbers below each panel indicate the correlation coefficient between the PJO indices and the AO or AAO indices.](image)
component is evidently important, since the upward wave propagation was observed even with the E-P flux calculated from the monthly mean data (not shown).

A seesaw of geopotential heights between the polar region and midlatitudes, i.e., AO or AAO (or annular mode), appears at lag 1 for the NH and 4 for the SH in Figure 9. A significant correlation between the PJO and the annular mode was observed in this stage as -0.50 for the NH and -0.42 for the SH. As expected from the meridional cross sections in Figure 1, the height anomaly in the NH accompanied a change in the meridional propagation of waves, particularly over the Atlantic and eastern Asian sector. Such change in wave propagation was not found in the SH. In the following period (lags 2 for NH and 5 for SH), the annular modes evolved into more regional north-south dipole patterns over the North Atlantic sector in the NH and the central southern Pacific sector in the SH.

3.3. Temporal Structure

The standard deviation of interannual variability of the polar temperatures in each month of the year repeated for three cycles is shown in Figures 6b and 7b for the NH and SH. Considerable variability occurs over the whole depth of the stratosphere during the cold season from November to April in the NH. In contrast, the zone of large variability in the SH is confined to a relatively small height range, which gradually descends from August to December. A comparison of the seasonal structures of the standard deviation with the anomalous patterns of individual years revealed that those anomalous patterns fit well into the standard deviation structure in the SH, whereas in the NH, the “envelope” of anomalous patterns may correspond to the patterns of standard deviation. This demonstrates that intraseasonal variability is more closely locked to the annual cycle in the SH.

The above conclusion was drawn from the temperature variability of the polar region. Although polar temperature is a good indicator of the wave-mean flow interaction in the midlatitudes of the stratosphere, it is relatively insensitive to subtropical wind variability in early winter. To investigate the variability of the entire stratosphere, analysis is made in low-dimensional space spanned by two leading EOFs calculated from anomalous zonal-mean zonal winds of all months. About 65% of the total variability is explained in each hemisphere by the two leading EOFs. Figure 11 shows the distribution of the time coefficients of EOF 1 and 2 for December and March of each year for the NH (top) and for July and October for the SH (bottom). The distribution of the coefficients has no clear structure in the NH, whereas the coefficients in the SH are clustered along a line that changes direction with the season. This indicates that the spatial pattern of the interannual variability changes with the seasons in the SH; i.e., it is a dipole type pattern in July, but a monopole type in October, as confirmed by Figure 1.

The lack of structure in the NH as depicted in Figure 11 does not indicate that there is no organized variability, but that the phase of the phenomenon is not locked to the annual cycle. In fact, Kodera et al. [2000] demonstrated that the trajectory of a phase vector tends to trace a circle on the plane, suggesting a periodic nature. Multichannel singular spectrum analysis (MSSA) [Vautard and Ghil, 1989] was conducted for the anomalous zonal-mean zonal wind to extract the periodic structure more clearly. MSSA is the same operation as E-EOF except that the attention is paid to the spectrum structure of the time coefficients. Therefore the operation is very similar to the E-SVD.
calculated here only uses data from the cold season. The band length was set to 7 months (a lag component of 0 to 6 months) to cover the entire cold season. The leading two modes explain almost the same total variance (12%) and form a pair. The time evolution of the spatial pattern of the leading mode was very similar to that shown in Figure 1a. The power spectrum of the normalized time coefficient of the leading mode is shown in Figure 12a. The power spectrum exhibited a sharp peak at the period of 5.1 months. The lateral peaks around 3.5 months and 7 months may be attributed to amplitude modulation of the 5-month signal by the annual cycle.

The same analysis was conducted for the SH. In that case, the leading mode explained 18% of the total variance and differed from the second one (10%). Therefore the leading two modes do not form a pair, which indicates that there is no apparent periodic signal with the intraseasonal timescale. The power spectrum of the normalized time coefficient of the leading mode in the SH is shown in Figure 12b. It is clear that the power associated with the intraseasonal timescale is very small and that most is associated with the interannual timescale (i.e., a period longer than 1 year) in the SH. This indicates that there exists only some periodicity with the interannual timescale in the SH. We found that the peaks near the origin (note that the power at the origin is zero owing to the use of anomalous data) and at the period of 2 years correspond to a long-term trend and the biennial oscillation, respectively, as noted by KK98.

4. Discussion

The spatial and temporal characteristics of the PJO were investigated for both the Northern and Southern Hemispheres. Negative zonal wind anomalies formed in the subtropical stratosphere at lags -1 and 0 propagated poleward and downward in both hemispheres (Figure 1). However, in the NH, negative anomalies quickly shifted poleward, giving way to the positive anomalies in the tropics and forming a deep meridional dipole structure (lags 1 and 2). In contrast, negative anomalies shift poleward only minimally during the winter in the SH, but positive anomalies are formed in the polar region (lags 0 and 1), creating a meridional dipole pattern in the stratosphere.

The differences of the two hemispheres can be explained by the differences in the convergence zones of the E-P flux in both hemispheres (Figure 2). Convergence of the E-P flux occurs mainly in the lower latitudes in early winter in the SH. The induced downward and poleward motion of the residual circulation produces warming in the midlatitudes by adiabatic heating and acceleration of the zonal wind in the polar region through Coriolis forcing, respectively. Since wave forcing is smaller in the high latitudes in early winter in the SH, positive wind anomalies are produced in the polar region, in contrast with the negative anomalies in the lower latitudes. When the convergence zone shifts to the polar region in spring (e.g., lag 3 of Figure 2b), the upward and downward flows of residual circulation produce adiabatic cooling and warming in the subtropics and the polar region, respectively. The poleward flow accelerates the zonal flow in the subtropics, where wave forcing is small. However, this effect is quite small. In the NH, on the other hand, the convergence of anomalous E-P flux occurs in high latitudes from early winter; so the dipole structure observed in the SH is not found in this period. The growth of positive anomalies in zonal winds in the NH at lag 2 is mainly explained by the change in wave forcing in the upper stratosphere.

The reason why the annular modes preferably appear when the zones of maximum polar temperature anomaly descend to the lowermost stratosphere can be explained as follows: If the polar temperature at the lowermost stratosphere is warmer, the zonal wind at high-latitude stratosphere becomes weaker owing to the thermal wind relation. As a result, the refractive index in the high latitude of the lowermost stratosphere becomes larger, and the tropospheric waves tend to propagate more polewardly, which makes a negative phase of the annular mode.

The vertical propagation of anomalous E-P flux in the NH changes with the development of the opposite polarity of an anomalous zonal wind in the subtropical stratosphere (Figure 1a). It is quite interesting that in the SH the vertical propagation of the waves is maintained over 6 months with the development of the wind in the stratosphere; the central latitude of upward propagation shifts to a higher latitude with the increase in negative wind anomalies (Figure 1b).

The successive downward propagation of negative and positive anomalies in the zonal wind anomalies in the NH suggests a periodic nature, which can be more clearly seen in GCM simulations under perpetual winter conditions [Christiansen, 1999; Yamazaki and Shinya, 1999]. Such variability is identified as stratospheric vacillation in a mechanistic model [Holton and Mass, 1976; Scaife and James, 2000; Kodera and Kuroda, 2000a]. Stratospheric vacillation cannot be expected during winter in the SH, where stationary planetary wave activity is much weaker. However, when the annual cycle is taken into account, planetary waves can propagate into the stratosphere during autumn and spring [Plumb, 1989; Yoden, 1990] and produce a wave-mean flow interaction in the stratosphere. In the real world, planetary waves always propagate into the stratosphere even during the winter but are deflected from the polar region by the strong polar night
jet. Hence no important wave-mean flow interactions are produced in the polar region until spring. This difference in climatological conditions during winter may explain why no poleward shift of the anomalous wind occurs during midwinter in the SH. Therefore better similarity is found with the early winter part of the NH PJO if the midwinter period is omitted from SH PJO, as illustrated in Figure 8. This suggests that the PJO in both hemispheres is essentially the same phenomenon produced through the interaction between the planetary waves and zonal-mean flow.

The similarity between the NH and SH PJOs is also found in association with the tropospheric circulation pattern in early winter. This may be a coincidence, but wave trains appear in 500-hPa height fields (lag -1 of Figure 9) in both hemispheres, with enhanced vertical propagation of the planetary waves (lag -1 of Figure 1). In addition, annular modes appear in the troposphere at the stage when the stratospheric anomalies propagate down into the lower stratosphere (Figures 5 and 9).

The AO index changes its polarity from negative to positive in late winter to spring in the NH, according to the successive downward propagation of the positive and negative temperature anomalies with the PJO (Figure 5). This result shows good agreement with the analysis of the origin of the month-to-month variability of the AO by Kodera and Kuroda [2000b].

Downward propagation of the temperature anomaly seen in Figure 5 is reminiscent of the downward propagation of the AO signal by Baldwin and Dunkerton [1999]. However, it must be noted that the downward signal shown in this figure is the polar temperature due to the PJO. Therefore the AO signal can be traced back to before the formation of the AO in the troposphere at the annular period. This indicates that the AO-related variability could be understood better in the framework of the PJO.

Changes in the meridional propagation of tropospheric waves with the formation of the annular mode were recognized in both hemispheres. However, they occurred in stationary waves in the NH and could be seen only for transient waves in the SH, which is consistent with the annular mode differences in the two hemispheres [Limpasuvan and Hartmann, 2000].

The annular mode evolves into a less zonally symmetric structure with time, as seen in Figure 9. This could also be coincidental, but it is interesting to note that these sectors where dipole anomalies develop correspond in both hemispheres to the sector where double-ridge or blockings occur more frequently [Lejenas, 1984]. This suggests further excitation of the regional mode.

Thompson and Wallace [2000] noted an out-of-phase relationship in the annular mode between the polar and equatorial temperatures in the lower stratosphere. A similar temperature anomaly pattern can be seen in Figure 4 (lag 1 for NH and 4 for SH). In fact, these temperature anomalies are produced through changes in residual circulation associated with the stratospheric process (PJO). As the annular mode and the PJO are well correlated at the annular period, the structure of the annular mode reflects that of the PJO. Note that the correlation does not mean one-to-one correspondence [Kodera and Kuroda, 2000b].

5. Conclusion

The month-to-month variability of the stratosphere and troposphere during the cold season was investigated by E-SVD analyses for both Northern and Southern Hemispheres. The extracted variability in both hemispheres can be considered to be the same variability mode, i.e., the PJO. It is characterized by a poleward and downward propagation of zonal-mean zonal wind anomalies from the subtropical stratosphere to the troposphere through interaction with planetary waves (particularly the wave number 1 component). An association of the annual mode with the PJO was also commonly found in both hemispheres. The major difference is their temporal characteristics. The NH PJO is essentially an intraseasonal variability for which the envelope is controlled by an annual cycle, while the SH PJO has an interannual variability with an intraseasonal structure.

This difference is thought to originate from a weak mean flow interaction during the SH winter. In short, the SH PJO is similar to an NH PJO for which evolution is frozen during the winter. It would be interesting to conduct GCM experiments using perpetual autumn or spring runs to investigate the structure of the SH PJO and its association with the annular mode.

In the present study, we focused our attention on the spatial and temporal structures of the intraseasonal timescale. It is also important to know the origin of the interannual variations in the PJOs, particularly the biennial oscillation and trends in the SH PJO (KK98).

Acknowledgment. The authors are grateful to T. Hirooka for providing Met Office, UK data.

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(Received September 18, 2000; revised April 20, 2001; accepted April 30, 2001.)