Role of the Polar-night Jet Oscillation on the formation of the Arctic Oscillation in the Northern Hemisphere winter

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The coupling between the Polar-night Jet Oscillation (PJO) and the Arctic Oscillation (AO) was examined using observational data. Positive and negative AO indices tended to appear within a specific phase of the time evolution of the PJO, and the lifetime of the AO tended to be longer when the AO was coupled with the PJO. We performed lagged regression analyses for Eulerian mean wave forcing to examine how the AO signal appears with the time evolution of the PJO. We found that the AO signal at the surface is created through meridional circulation driven by the combined effect of mechanical and thermal eddy forcing at some stage of the PJO. The effect of these wave forcings on sea level pressure (SLP) changes was investigated using a zonal-mean quasi-geostrophic model on the sphere. Approximately 40% of the polar SLP change was found to come from thermal forcing when the AO-like signal appeared and two thirds from the stratosphere, whereas 60% of the polar SLP change was a result of mechanical forcing that originated almost in the troposphere. The contribution from the wave of the zonal wave number 1 component was very important in both forcings, particularly for SLP in the polar cap region. Waves of zonal wave number 2 and 3 components also contributed significantly to tropospheric forcing.

INDEX TERMS: 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; 3334 Meteorology and Atmospheric Dynamics: Middle atmosphere dynamics (0341, 0342); 1620 Global Change: Climate dynamics (3309); 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); KEYWORDS: polar-night jet oscillation, arctic oscillation, dynamical interaction


1. Introduction

The Arctic Oscillation (AO) is the dominant hemispheric seesaw variability in sea level pressure (SLP) between the polar area and the surrounding midlatitude belt in the troposphere [Thompson and Wallace, 1998, 2000]. It appears throughout the year, but is particularly active during the cold season. The AO signal sometimes couples with the stratosphere during that time [Baldwin et al., 1994; Perlwitz and Graf, 1995], and tends to propagate downward from the upper stratosphere [Baldwin and Dunkerton, 1999; Christiansen, 2001]. The downward propagation of the AO signal in the stratosphere is essentially the same signal as the Polar-night Jet Oscillation (PJO) [Kuroda and Kodera, 2001] (hereinafter referred to as KK01). This is a dominant mode of variability of the stratosphere and is characterized by the poleward and downward movement of the anomalous zonal-mean zonal wind [Kodera et al., 1990].

The question of how stratospheric variability (PJO) affects the appearance of the AO in the troposphere is very important for understanding troposphere-stratosphere coupling, as well as for predicting climate changes [Baldwin and Dunkerton, 2001]. Previous studies [Black, 2002; Ambaum and Hoskins, 2002; Sigmond et al., 2003] describe several mechanisms that induce this coupling. Black [2002] and Ambaum and Hoskins [2002] considered a mechanism in which the potential vorticity in the stratosphere induces change in the surface pressure. Their works essentially address the balance between the potential vorticity in the stratosphere and other field variables, while the underlying physical mechanisms are unclear. Any change of SLP must be associated with a change of the meridional circulation since SLP should be proportional to the column total air mass above the surface. Sigmond et al. [2003] examined the polar SLP change associated with meridional circulation in the stratosphere in this respect.

We used observational data in this study to examine the role of the PJO on the formation of the AO. A lagged correlation analysis based on the PJO variability was performed to ascertain the time evolution of the PJO/AO coupling with time. Particular attention was paid to the time evolution of the wave forcing and the resulting change in meridional circulation and the SLP change, as noted by Sigmond et al. [2003]. The SLP change associated with meridional circulation induced by wave forcings in particular was analyzed by means of the Eulerian mean quasi-geostrophic model on the sphere.
This paper is organized as follows. Section 2 describes the data set and principal method of analysis. Section 3 provides the results of the analysis. After a discussion of the results in Section 4, Section 5 offers conclusion and remarks.

2. Data and Method of Analysis

We used two data sets in this analysis. The first was the updated satellite/reanalysis coupled data from the surface to 1 hPa used for KK01. The second was reanalysis data up to 10 hPa from the NCEP/National Center for Atmospheric Research (NCAR) [Kalnay et al., 1996]. The stratospheric portion of the first data was analyzed by U.S. National Centers for Environmental Prediction (NCEP)/Climate Prediction Center (CPC) (formerly NMC/CAC), and the stratospheric winds were calculated from a satellite-derived geopotential height analyzed by CPC using the non-linear balanced wind relation [Randel, 1992]. All missing data were interpolated in time. The tropospheric part (100 hPa and below) was taken from reanalysis data of the NCEP/NCAR.

The period of this study covers 21 winters from 1979 to 1980 through 1999 to 2000. The analysis was made for the six months of extended winter from November to April.

The annual cycle was first removed from the data for the present analysis, and anomalous data was constructed. The PJO and AO are defined by an EOF analysis based on month-to-month variability of the anomalous data in extended winter. The PJO is defined from the polar temperature, whereas the AO is defined from SLP north of 20o N, as used by Thompson and Wallace [1998]. The PJO is a propagation phenomenon that varies with time. Therefore we used the two leading EOFs of the polar temperature to capture the time evolution of the PJO, similar to Kodera et al. [2000]. EOF analysis was performed based on the polar temperature (defined by the average north of 80 degrees) from the surface to a 1 hPa level. Figure 1 depicts the profile for EOF1 and EOF2 of the polar temperature defined by the regression. Note that there are only very small signals in the troposphere, and they capture most stratospheric variability. EOF1 (EOF2) explains 60% (33%) of the total variance. Therefore the phase space spanned by the time coefficients of these two vectors (PC1 and 2) explains 93% of the total variance (we refer to this space as “PJO space”). Note that these time coefficients have been normalized. PC1 and PC2 time coefficients were calculated for every pentad (five-day period) by projecting EOFs to six-pentad running averaged anomalous patterns of polar temperature. AO indices of respective pentads were defined similarly to those of the PJO. Second-order quantities, such as the Eliassen-Palm (E-P) flux, were calculated based on daily-mean data, which were six-pentad running averaged.

Most lagged correlation and regression maps presented in this paper were calculated based on the PC1 time series of the central 18-pentad period (from 15 December to 15 March) of 21 extended winters. The time evolution due to the PJO was then examined. Figure 2 illustrates the lagged regression of the polar temperature against PC1 and PC2. Note that patterns of lag 0 agree with EOF1 and 2. Areas of 95% significance levels (correlation exceeding 0.25) are shaded. The statistical significance was calculated here based on the assumption that each month of the reference time series (total 3 x 21) was independent. Downward propagation of the anomalous polar temperature with almost constant velocity can be clearly seen. Note that the pattern at lag -30 days of PC1 (PC2) is very similar to -EOF2 (EOF1), and that at lag +30 days to EOF2 (-EOF1). Therefore the time evolution from lag -30 days to +30 days corresponds to a typical half-cycle revolution of PJO. The state vector rotates counterclockwise in the PJO space during the cold season, corresponding with the successive downward propagation of the anomalous polar temperature [Kodera et al., 2000]. Although this rotation is prominent, particularly in major warming years, most winters exhibit counterclockwise rotation, albeit small in amplitude.

The mechanistic model utilized to diagnose meridional circulation and surface-pressure changes due to wave forcings was essentially the same as that used by Haynes and Shepherd [1989] (hereinafter referred to as HS89). It is a zonal quasi-geostrophic model, and all calculations were performed by horizontal expansion by the zonal Hough or its associated function. We used 100 vertical levels with a vertical interval of 10 hPa in the pressure coordinate, and...
3. Results

3.1. Coupling Between the Polar-night Jet Oscillation (PJO) and Arctic Oscillation (AO)

Previous studies [e.g., Baldwin et al., 1994; Perlwitz and Graf, 1995] indicated the existence of dynamic coupling between the stratosphere and troposphere. We performed an analysis based on present data sets to clarify the evidence of PJO/AO coupling. The PJO space was filled with a total of 803 pentads, corresponding to the number of pentads in 21 extended winters. We plotted values of the AO index and Student’s-t with the probability distribution function (PDF) of state vectors on the PJO space to examine how PJO and AO are related. The distribution of the state vectors were irregular on the space; thus the number of state vectors (PDF) was calculated for every 0.5 grid points of the coordinates by using an accumulating window with a 0.5 half-width. The mean AO-index and the Student’s-t value were calculated similarly. The PDF distribution (Figure 3a) exhibited a sharp peak around (0.5, −0.5) that decreased rapidly with distance from the peak. Figure 3b indicates the mean AO value (contour) and the Student’s-t (shading) on the PJO space. Only grids of the Student’s-t that the total number of state vectors exceeds six are shaded. The area of light shading indicates a Student’s-t greater than 2 (4). Student’s-t is shown in the area where PDF is greater than 6, which is shaded in Figure 3a.

121 horizontal grids of equal intervals for the sine latitude (see Appendix for details).

3.2. Time Evolution of the PJO

We performed lagged-regression analyses of reanalysis data sets based on PC1 to examine how the PJO evolves with time. Figure 4 illustrates the lagged regression of the zonal-mean zonal wind with the E-P flux (upper panels), SLP (middle panels), and the zonal-mean SLP (lower panels). The poleward and downward movements of the anomalous zonal wind with changes of the E-P flux are clearly evident. The overall feature of the time evolution from −30 days to +30 days corresponded very well with lag 1 to lag 3 months of KK01 in the northern hemisphere winter.

There was also a significant correspondence between the variablity of the zonal wind and the E-P flux [Kuroda and Kodera, 1999] (hereinafter referred to as KK99). Weaker upward propagation of the E-P flux corresponded to stronger zonal wind from lag −30 days to −15 days. An enhanced upward propagation of the E-P flux from lag 0 corresponded to a weaker zonal wind and poleward and downward shifts of the zonal wind, with an anomalous negative zonal wind extending into the upper stratosphere in turn (from lag +15 to +30 days). The correspondence of the variability of the zonal wind with that of the E-P flux suggests that this variability comes from wave-mean flow interaction [KK99; KK01].

An anomalous AO-like pattern appeared at lag 0 in the SLP corresponding to the downward propagation of the positive wind anomaly from lag −30 days to 0, in agreement with the AO index and Student’s-t value in Figure 3b.

The overall feature of the time evolution from −30 days to +30 days corresponded very well with lag 1 to lag 3 months of KK01 in the northern hemisphere winter.

Table 1. Comparison Between the Mean Duration of the AO Expressed in Pentads for a Large PJO Amplitude and Small PJO Amplitude

<table>
<thead>
<tr>
<th></th>
<th>Small PJO</th>
<th>Large PJO</th>
</tr>
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<tbody>
<tr>
<td>Positive AO</td>
<td>4.0 (8)</td>
<td>6.0 (11)</td>
</tr>
<tr>
<td>Negative AO</td>
<td>3.5 (8)</td>
<td>5.9 (16)</td>
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</tbody>
</table>

*aNumbers in brackets indicate the number of samples; see text for details. PJO, Polar-night Jet Oscillation; AO, Arctic Oscillation.

Table 2. Same as Table 1 but With a Comparison of the Mean Duration of Pentads for the Position of the State Vector on the PJO Space of a Large PJO Group

<table>
<thead>
<tr>
<th></th>
<th>PC1 &gt; 0</th>
<th>PC1 &lt; 0</th>
</tr>
</thead>
<tbody>
<tr>
<td>Negative AO</td>
<td>6.9 (9)</td>
<td>2.0 (2)</td>
</tr>
<tr>
<td>Positive AO</td>
<td>4.6 (5)</td>
<td>6.5 (11)</td>
</tr>
</tbody>
</table>
with the result depicted in Figure 3b and in previous studies [KK99; Baldwin and Dunkerton, 1999; Christiansen, 2001]. The appearance of the AO-like anomaly corresponded with the enhanced equatorward propagation of the E-P flux in the troposphere [Limpasuvan and Hartmann, 2000]. It should be noted that the signal of the AO-like anomaly in the north Pacific was very small compared with that of the typical AO pattern.

3.3. Wave Forcing and Surface Pressure Change

[16] We compared the lagged regression of the E-P flux divergence with that of the zonal-mean zonal wind tendency based on PC1 to investigate how the variability of the zonal wind is driven by wave forcing with time. Figure 5 demonstrates that the acceleration due to E-P flux divergence corresponded very well with the tendency of the zonal wind in the stratosphere. An analysis based on zonal wave number decomposition demonstrated that the variability of the E-P flux came primarily from the zonal wave number 1 component (not shown) [KK01]. Although correspondence of the E-P flux divergence with the tendency of the zonal wind was very good in the stratosphere, it was not good in the troposphere, indicating that the E-P flux divergence analysis does not work well in the troposphere. One reason for this is the existence of the surface as a lower boundary in the transformed Eulerian mean (TEM) formulation. The lower boundary condition is very complex in the TEM formulation and acts as a significant source of forcing, making it difficult to analyze the troposphere [HS89]. We overcame this difficulty by analyzing the eddy forcings in the Eulerian mean (EM) formulation, whose lower boundary condition can be treated very simply [HS89].

[17] There are two types of eddy forcings in the EM formulation, mechanical (or momentum) forcing, which primarily forces zonal wind, and thermal forcing, which forces temperature. These forcings are proportional to the meridional divergence of the northward eddy momentum (mechanical forcing) and heat fluxes (thermal forcing) (see Appendix). The northward eddy momentum flux represents northward transport of positive zonal momentum; therefore more zonal momentum leaves the region than enters if its divergence is positive. Meridional divergence of the northward eddy momentum thus decelerates the zonal wind. A similar interpretation can be applied to thermal forcing. However, these two forcings were combined through equations and are subject to variabilities, including both zonal wind and temperature. It is apparent that the characteristics of the EM eddy forcings arise from the characteristics of the eddy fluxes. We therefore examined eddy fluxes first.

[18] The meridional and vertical components of the E-P flux are proportional to eddy fluxes, so the time evolution of
the E-P flux in Figure 4 corresponds well with the evolution of these eddy fluxes. Figure 6 depicts the lagged regression of eddy fluxes due to PC1. There is a local maximum in the eddy momentum flux at the upper troposphere around 50°N. It does not travel, but gradually increases and peaks at lag -5 days, then gradually decreases. This corresponds to the enhanced equatorward propagation of the E-P flux at lag 0 in Figure 4. In contrast, the eddy heat flux has a relatively simple standing structure throughout, and its amplitude increases toward the upper stratosphere. However, the time evolution of the amplitude was very prominent. It was significantly negative at lag -30 days, increased with time, peaked with a significantly positive value at lag 5 days, and then decreased with time. This feature corresponds with the change of the vertical propagation of the E-P flux in Figure 4. Note that the mechanical forcing in the EM formulation is proportional to the meridional divergence in the meridional component of the E-P flux, and the thermal...
forcing is proportional to the meridional divergence in the vertical component of the E-P flux.

[19] Figure 7 illustrates the lagged regression of the mechanical and thermal forcings by the PC1. The arrows indicate meridional circulation. There are two forcing centers for mechanical forcing (upper panels), one in the troposphere and the other traveling downward from the stratosphere. The tropospheric forcing appeared around 65°N in the 300 hPa level at lag -15 days and developed in conjunction with forcing from the stratosphere at lag 0. It then became weaker but was still connected with the stratospheric forcing until lag +15 days. This forcing center corresponded to the polar-side gradient of the tropospheric momentum eddy flux in Figure 6a, and is significant compared with that of the subsropical center due to the metric factor. Stratospheric forcing appeared at 80°N in 70 hPa at lag 30 days and propagated downward, reaching 85°N and 300 hPa at lag +15 days. A negative signal above this forcing followed this signal. This can also be traced back to the downward propagating eddy momentum flux from the stratosphere. Wave number decomposition analysis revealed that this came primarily from a wave of the zonal wave number 1 component in the stratosphere and a combination of components 1 to 3 in the troposphere. Thermal forcing was small in the troposphere, but became substantial toward the upper stratosphere. In contrast, the magnitude of the mechanical forcing in the stratosphere was as great as that in the troposphere.

[20] The features of thermal forcing are particularly interesting. A strong standing meridional dipole forcing pattern with a negative high-latitude center from the surface to the upper stratosphere appeared at lag -30 days. The stratospheric forcing decreased in amplitude as time passed and almost vanished at -15 days, after which its amplitude increased with changing polarity, and a deep standing meridional dipole structure with positive high-latitude forcing was created again around the peak period of lag +5 days. It gradually decreased its amplitude, but retained its structure at lag +15 days. Although stratospheric forcing affects the signal as an oscillator, tropospheric forcing exhibits some equatorward propagation with time. The meridional dipole structure is a direct result of the meridional gradient of the standing structure of the eddy heat flux in Figure 6b. Wave number decomposition analysis revealed that this came primarily from a wave of the zonal wave number 1 component in the stratosphere and a combination of components 1 to 3 in the troposphere. Thermal forcing was small in the troposphere, but became substantial toward the upper stratosphere. In contrast, the magnitude of the mechanical forcing in the stratosphere was as great as that in the troposphere.

[21] The meridional circulation (vector in Figure 7) indicates that the mean vertical velocity was upward (downward) if the thermal forcing was positive (negative), and the meridional velocity was equatorward (poleward) if the mechanical forcing was positive (negative). In fact, the lagged correlation patterns of the meridional and vertical velocities were very similar to those of mechanical and thermal forcings (Figure 8) because advection and eddy forcing should be nearly balanced in a slow variability system (PJO). Data of the vertical velocity was available only up to 100 hPa in the reanalysis data. Also, the correlation of meridional wind near the surface was substantial (0.5) at lag 0 to +15 days, which indicates the importance of meridional circulation for coupling of the stratosphere to the surface [Kodera and Kuroda, 2003]. This result suggests that the meridional circulation was primarily driven by eddy forcings.

[22] We examined this in further detail by using an EM model (see Appendix) to analyze the eddy-forced meridio-
nal circulation and surface pressure change. Equations were applied for the anomalous fields defined as departures from climatology. The stability of the mean field was estimated from the winter mean (December to March) temperature. Figure 9 illustrates the meridional circulation with the stream function (upper panels) and the SLP tendency (lower panels) by using eddy forcings calculated from the regression by PC1. EM wind (vectors) is shown only up to 100 hPa for comparison with the observation (Figure 7). The present results are identical to those obtained by regression of the circulations evaluated at each moment by the model, due to the linearity of the equation. It is clear that the overall features of the observed meridional circulation are reproduced well from this calculation, particularly for the upper troposphere to lower stratosphere and high-latitude areas. The area of increasing surface pressure corresponded well with the downward flow, and the area of decreasing surface pressure, with the upward flow. Note that the surface pressure change can be reproduced by setting an adequate lower boundary condition of

![Figure 8](image1)

**Figure 8.** Same as Figure 2 except for the lagged correlation of the meridional (top) and vertical (bottom) velocities due to the PC1. Contour interval is 0.1 but is shown only for the 95% significance region (greater than 0.25), except for the 0 contour.

![Figure 9](image2)

**Figure 9.** Calculated meridional circulation (top) and the surface-pressure tendency (bottom) from the observed eddy forcings in Figure 7. Arrows indicate velocities on the meridional plane. The horizontal (vertical) reference arrow indicates $6 \times 10^{-2}$ ms$^{-1}$ ($3 \times 10^{-4}$ ms$^{-1}$). The contour interval is $5 \times 10^8$ kgs$^{-1}$ for the mass stream function, and the unit of vertical axis is 0.01 hPaday$^{-1}$ for the surface-pressure tendency. Small vectors are neglected.
The time evolution of the SLP tendency from lag $-30$ to $0$ days corresponded well with the formation of AO and was maximized at lag $0$ with a value of $-0.7$ hPa/day at the polar cap. After that, it gradually decreased and almost vanished at lag $+30$ days, in correspondence with the weakened meridional circulation.

This analysis demonstrates the SLP tendency due to eddy forcings alone and should not be directly compared with the observed tendency. In fact, the observed SLP tendency is composed mainly of eddy forcing and frictional forcing and they are almost balanced, as will be described later. However, frictional forcing follows a given forcing, and thus the calculated SLP tendency should be a good indicator for the actual SLP anomaly, as can be seen by comparing the lower panels of Figures 4 and 9.

We analyzed the roles of eddy forcings in the meridional circulation and the SLP by applying mechanical and thermal forcing separately. Deep dipole stratosphere-troposphere meridional circulation due to the thermal forcing (Figure 10b) was prominent throughout. Anomalous meridional circulation at high-latitude changed its polarity from clockwise at lag $-30$ days to counterclockwise at lag $0$ to $+15$ days, corresponding to the change of polarity of the eddy thermal forcing. The polar SLP tendency changed from positive at lag $-30$ days to negative at lag $0$ to $+15$ days, corresponding with the change of polarity of the meridional circulation. Comparing these results with that of the total forcing indicates that the deep meridional circulation seen in Figure 9 (and Figure 7) is due to thermal forcing, but the effect of the polar cap SLP is only about $40\%$ at the stage when the AO-like signal appears (called as the annular stage).

Figure 10a clearly indicates that the meridional circulation due to mechanical forcing was restricted primarily to inside the troposphere, centered around $55$ to $60^\circ$N at $600$ hPa. Its counterclockwise circulation gradually enlarged from lag $-15$ days, peaked around lag $0$, and then gradually decreased. The tendency of the polar SLP also changed, corresponding with the change of meridional circulation. We found that approximately $60\%$ of the polar SLP change resulted from eddy mechanical forcing at the annular stage.

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We also applied wave forcing separately in the stratosphere and troposphere (Figure 11). We separated the stratosphere and troposphere at 270 hPa for all latitudes, following Sigmond et al. [2003] for simplicity. The results indicated that about 40% of the effect of polar-SLP tendency came from the stratosphere, and that it contributes to about 60% of the total thermal forcing and 30% of the total mechanical forcing (not shown). Note that the general features of meridional circulation and the SLP tendency due to stratospheric (tropospheric) forcing are very similar to those of thermal (mechanical) forcing, corresponding with the dominance of each forcing in each domain (Figure 7).

Zonal wave number decomposition revealed that the greatest contribution to the meridional circulation and the effect of the SLP came from eddy forcing of the wave of the zonal wave number 1 component at the annular stage (Figure 12). The general features due to the zonal wave number 1 component were very similar to those from total eddy thermal forcing (Figure 10b). This corresponded with the large forcing in the stratosphere and revealed deep meridional circulation. It is notable that nearly 65% of the total decreased polar cap SLP tendency around the annular stage came from a zonal wave number 1 component. The residual portion is well explained by tropospheric forcing of zonal wave number 2 and 3 components, while the contribution from wave numbers equal to or greater than 4 was very small. The contribution of the zonal wave number 2 component dominated the meridional circulation in the troposphere, particularly in the lower troposphere.

No external forcings, except from eddies, were included in the previous analysis. Therefore the reaction due to wave forcings must be stationary, and the zonal wind and temperature must increase linearly with time (A9). In reality, this cannot occur, and an increase of the zonal wind and temperature should be balanced by frictional and radiative forcings. We calculated the residual of the primitive equation from six-hourly reanalysis data sets to estimate these forcings. The calculated meridional circulation and SLP tendency from non-eddy forcing at lag days −30, 0, and +30 are indicated in Figure 13. It can be seen that the frictional forcing at the surface was sufficiently large to create considerable meridional circulation, particularly in the subtropics. It is also notable that the SLP tendency was almost the same as that due to eddy forcing (Figure 9), except with the opposite signal. Assimilation data was created through a violation of the balance of equations by introducing the observational data. Therefore we cannot attempt to evaluate non-eddy forcing precisely by this method. However, we believe that its overall qualitative character is correct.

We included the Rayleigh friction and Newtonian cooling terms in the equation and integrated the model with prescribed eddy forcings related with the PJO to examine the consistent time evolution of the zonal wind and meridional circulation.
ional circulation for comparison with the observation. The Rayleigh friction coefficient ($\lambda$) and Newtonian cooling rate ($\alpha$) are expressed by simple functional forms of height from surface ($z$) and are expressed as follows:

$$\lambda = \frac{R_0}{\exp(-z/600)} + R_1 \{1 + \tanh((z - 20000)/5000)\}$$ (1)

$$\alpha = N_1 \{1.5 + \tanh((z - 25000)/7000)\}.$$ (2)

Constant values are $R_0 = 1/(0.5 \text{ days})$, $R_1 = 1/(30 \text{ days})$, and $N_1 = 1/(10 \text{ days})$; $z$ is expressed in meters. The value of $R_0$ and the functional dependence on height are roughly estimated from the frictional forcing calculated from the reanalysis data, and the small value of $R_1$ is introduced for stability of the integration. The Newtonian cooling rate is determined so that the timescale of the radiative damping becomes 20 days in the troposphere and four days in the upper stratosphere. Time integration was then performed by the Matsuno (Euler backward) scheme, with a time step of 0.1 days and the initial condition set at day −35 (Figure 14). It was evident that the overall poleward and downward movement of the zonal wind and formation of the AO-like SLP structure at day 0 are reproduced well by the integration, though the wind in the upper stratosphere differs substantially from the observation. The meridional circulation is stronger than in the eddy-only case (Figure 9) due to the impact of frictional forcing on the zonal winds and becomes comparable to the observation (Figure 7). Thus the decreasing SLP of the polar cap due to eddy forcing around day 0 is balanced with the increasing SLP due to the meridional circulation arising from frictional forcing at the surface. Thus the role of frictional forcing through zonal winds, as well as eddy forcing, is important in explaining the formation of the AO-like SLP change.

4. Discussion

[30] The AO-like pattern induced on the surface with the PJO (Figure 4) differed from the typical AO pattern [compare Thompson and Wallace, 1998, Figure 1] because the signal at the north Pacific center is very small and the Arctic center is more symmetric and does not shift toward the north Atlantic sector. These are equivalent to the dominance of the zonal wave number 0 component in the Arctic and the zonal wave number 1 component in the middle latitude, in contrast to the dominance of the wave number 2 component in the typical AO. The zonal wave number 1 structure in the SLP should be related to dominance of the wave number 1 component that causes PJO in the stratosphere.

[31] Zonal wave number decomposition may have physical significance in the stratosphere because of the transmission property of planetary waves by near-uniform zonal winds, but this may not be so in the troposphere. Analysis indicates that the behaviors of zonal wave number 1 and 3 components were very similar in the troposphere. In fact, the E-P flux of the wave number 3 component propagated more equatorward when that of wave number 1 propagated more equatorward at lag −15 days to 0 (not shown). This is also apparent for the enhanced upward propagation at lag +15 days, which suggests that these waves are excited and propagate as one wave. It is interesting to note that the phases of these waves were identical to their climatological phases (around 55°N at lag 0), and corresponded to the formation of an NAO-like signal over the North Atlantic.

[32] Though the analysis presented here was performed based on daily-mean eddy data, a comparison of the pentad-mean eddy data was also performed to examine the role of synoptic waves. A high-frequency wave was found to contribute primarily to mechanical forcing in the troposphere and thermal forcing in the stratosphere. As a result, approximately 15% of the polar cap SLP tendency came from the contribution of high-frequency waves at the annular stage (not shown). Although the PJO is a very slow variability, this result suggests that the change of mean fields affects the generation and propagation characteristics of synoptic waves to create AO-related variability in the polar area.

[33] Sigmond et al. [2003] reported that the meridional mass flux variation in the stratosphere precedes that in the
troposphere by about one day in winter. This small lag cannot be directly expressed in the quasi-geostrophic model used here. However, use of the quasi-geostrophic model to analyze meridional circulation and SLP changes will yield a satisfactory approximation since the lag is very small compared with the typical timescale of the PJO. The lag of the mass flux between the stratosphere and troposphere described by Sigmond et al. [2003] should originate primarily from wave forcing from the stratosphere since the meridional circulation driven by tropospheric wave forcing is almost totally confined within the troposphere (Figure 11).

[34] The lagged correlation of the zonal wind (Figure 4) demonstrates that it has a local minimum in the middle troposphere, and increases toward the surface. This indicates that the zonal wind at the lower troposphere is not directly driven by the stratospheric wind, but through meridional circulation, particularly by the meridional wind near the surface [Sigmond et al., 2003]. This is consistent with the significant correlation of meridional wind near the surface (Figure 8).

[35] We evaluated the SLP change and its ratios due to respective components of wave forcings by a diagnostic equation. This may not be proportional to the actual SLP anomaly due to frictional effects. We verified this by integrating the EM model with applications of respective components of wave forcings. The resultant SLP anomalies were approximately proportional to the SLP tendencies evaluated by the diagnostic equation (not shown). This can be explained as follows. In a slow variability system such as PJO, eddy forcing and friction are nearly balanced for the body momentum torque as well as the SLP tendency at all times. Frictional forcing should therefore be determined so that it is nearly balanced with the given eddy forcing. The meridional gradient of the SLP is determined from the zonal wind at the surface (which is obtained from the frictional forcing), whereas the SLP tendency is directly determined by the eddy forcings. This result indicates that use of a diagnostic equation to analyze the SLP change is of value.

[36] We used EM equations in this paper to analyze meridional circulation and surface pressure changes due to the PJO. These equations should be equivalent to TEM equations in the mathematical sense. However, only one forcing (E-P flux divergence) exists in the TEM system, as opposed to two forcings in the EM system. In fact, residual circulation is calculated by an equation whose source consists only of E-P flux divergence in the TEM system [Andrews et al., 1987, equation (3.5.8)], in contrast to the two-source omega equation in the EM system (A.6). As a result, one residual circulation in the TEM system corresponds with multiple meridional circulations in the EM system.
system. We note, however, that it is the EM vertical velocity and not the residual one that corresponds with the correct mass transport (A.4). We should also note that the proper lower boundary condition in the TEM system cannot be expressed by the TEM field variables. Thus the TEM system is not suitable for estimating the effect of meridional circulation on surface pressure.

[37] The variability of the SLP due to PJO is dominant at the polar cap (Figure 4). In contrast, the polar center of the SLP in the typical AO appears more on the Atlantic side. Therefore we defined the polar SLP index as the mean SLP variability averaged north of 60°N. We also defined the W-AO index as the mean vertical velocity averaged from 600 hPa to 400 hPa north of 60°N, following Kodera and Kuroda [2003], since the polar SLP change is very sensitive to the polar vertical velocity in the middle troposphere. The W-AO index and polar SLP index exhibit a very good simultaneous correlation of 0.73. The correlation between the AO index and PC1 has a maximum value of 0.35 at lag -5 days. In contrast, the W-AO index (polar SLP index) reveals a maximum correlation of 0.52 (0.48) with PC1 at lag 0. The lagged correlation with the W-AO index (or polar SLP index) exhibits a very clear downward propagation of the signal from the stratosphere (not shown). Thus the stratospheric signal appears more strongly in the polar cap region through meridional circulation.

[38] We noted a substantial difference between the AO-like SLP signal associated with the PJO and the typical AO. The SLP signal in the former was more symmetric in the polar area, and was related to the vertical wind inside the polar cap and to the signal at midlatitudes consisting of more zonal wave number 1 components. The center of action in the typical AO signal shifted more to the north Atlantic side compared with the PJO-related signal, and a second action center in the north Pacific was more apparent.

A separate paper will examine the typical AO signal and compare it with the PJO-related signal.

5. Conclusion and Remarks

[39] The role of the PJO in the formation of the AO was examined based on observational data. The relationship between the PJO and AO was investigated first. The polarity of the AO was found to be significantly dependent on the phase of the PJO, and the AO tended to have a longer lifetime when the PJO had sufficient amplitude and a preferable phase.

[40] Lagged regression analyses based on the PJO were performed to observe changes of the field variables, such as wave forcings and meridional circulation associated with the time evolution of the PJO. The effect of wave forcings on surface pressure changes was analyzed using a zonal-mean quasi-geostrophic model on the sphere. An SLP anomaly is evidently created through activity of the meridional circulation associated with a change of the wave forcings. We further examined the role of respective wave forcings in surface pressure changes. For surface-pressure changes at the polar area when AO-like signal appears we found that (1) about 60% (40%) of the effect comes from mechanical (thermal) forcing, (2) about 60% (40%) of the effect comes from wave forcing in the troposphere (stratosphere), (3) about 65% of the effect comes from the zonal wave number 1 component while the remainder results primarily from wave number 2 and 3 components, and (4) about 15% of the effect comes from transient waves. Surface pressure changes associated with meridional circulation driven by wave forcings are almost balanced with those due to frictional forcing. However, an anomalous SLP was found to appear almost in proportion to the surface pressure changes due to wave forcings.

[41] Previous studies [KK99 and KK01] demonstrated that the stratospheric variability of the PJO is well explained by the wave-mean flow interaction of waves of mainly zonal wave number 1 components. This variability can occur even if the wave forcing from below is constant in time [Christiansen, 1999]. However, the present analysis indicates that the wave amplitude and wave forcing in the lower troposphere varies significantly with time (Figure 4 and Figure 7), corresponding with the variability of the zonal wind and the E-P flux in the stratosphere. In fact, the meridional wind at the surface around 60°N and the eddy heat flux in the troposphere correlate very well in extended winter (Figure 15). This suggests a close relationship between the wave generation at the surface and eddy-forced meridional circulation. The wave strength at the surface has a significant impact on the formation of eddy forcings in the upper troposphere to stratosphere (which creates meridional circulation). Thus the strength of meridional circulation is controlled in part by wave generation at the surface. Wave generation at the surface and eddy forcings in the upper troposphere to stratosphere should therefore create feedback through meridional circulation. Feedback of the meridional circulation-surface interaction, as well as the wave-mean flow interaction in the stratosphere, must be examined to clarify the maintenance mechanism of PJO.

Figure 15. Simultaneous correlation of northward eddy heat flux with a 1000-hPa meridional wind averaged from 50°N to 75°N in extended winter (November to April). Contour interval is 0.1 but shown only for the 95% significance region (greater than 0.19) except for the 0 contour. Areas of 95% significance are shaded.
and PJO/AO coupling. This will be performed in a future study.

Appendix A

[42] The equation used in this study is a quasi-geostrophic Eulerian mean (EM) equation that is essentially the same as that used by Plumb [1982] and HS89:

\[
\frac{\partial u}{\partial t} - 2\Omega \mu v = F + G \\
2\Omega \mu u = -\frac{a}{\sqrt{1 - \mu^2}} \frac{\partial \Phi}{\partial \mu} \\
\frac{\partial \Phi}{\partial p} = \frac{RT}{p} \\
\frac{\partial T}{\partial t} - \Gamma \omega = Q + S \\
1 \frac{\partial}{\partial \mu} \left( \sqrt{1 - \mu^2} v \right) + \frac{\partial \omega}{\partial p} = 0
\]

(\text{A1})

where

\[ F = -\frac{1}{a\sqrt{1 - \mu^2}} \frac{\partial}{\partial \mu} \left[ \omega \sqrt{1 - \mu^2} \right] \]
\[ Q = -\frac{1}{a \partial \mu} \left[ \sqrt{\Gamma} \sqrt{1 - \mu^2} \right] \]

(\text{A2})

are eddy mechanical and thermal forcings, \( G \) and \( S \) are frictional forcing and diabatic heating, \( \mu \) is the sine latitude, \( \omega \) is the vertical pressure velocity, \( \Gamma = -\partial T_0/\partial p + \kappa T_0/p \) is the stability of the basic atmosphere, \( \Omega \) is the angular velocity of the Earth, \( a \) is the radius of the Earth, field variables with primes represent departures from the zonal mean, the overbar denotes zonal averaging, and other symbols follow the usual convention [e.g., Andrews et al., 1987]. Also note that the meridional winds considered here are essentially ageostrophic in nature.

[43] We applied these equations for those variables defined as a departure from the climatology. Therefore all field variables, including \( F \) and \( Q \), should be regarded as anomalous variables.

[44] Boundary conditions of the equation should be

\[ \frac{\partial \Phi}{\partial t} = 0, \]

(\text{A3})

for the lower boundary and \( \omega = 0 \) for the upper boundary; lateral conditions are \( v = 0 \) for \( \mu = \pm 1 \). Here, equation (A3) is used for its approximated form as

\[ \frac{\partial \Phi}{\partial t} + \omega \frac{\partial \Phi_0}{\partial p} = 0, \]

(\text{A4})

where \( \Phi_0 \) is the geopotential height at the basic state, since the surface-pressure change should be small compared with the lowest pressure level of the model (1000 hPa).

[45] Equation (A1) with boundary conditions of equation (A4) are easily solved if all field variables are expanded by the zonal Hough function \( \Theta_n(\mu) \) or its associated function \( B_n(\mu) \) [Plumb, 1982; HS89] as

\[ \omega = \sum_n \omega_n \Theta_n(\mu), \ T = \sum_n T_n \Theta_n(\mu), \ \Phi = \sum_n \Phi_n \Theta_n(\mu), \]
\[ Q = \sum_n Q_n \Theta_n(\mu), \ S = \sum_n S_n \Theta_n(\mu), \ v = \sum_n v_n B_n(\mu), \]
\[ u = \sum_n u_n B_n(\mu), \ F = \sum_n F_n B_n(\mu), \ G = \sum_n G_n B_n(\mu). \]

(\text{A5})

Equations (A1) and (A4) can then be put into ordinary differential equations with respect to \( p \) as

\[ 4\Omega^2 a^2 p d^2 \omega_n \over dp^2 + \varepsilon_n \omega_n = 2\Omega a \left( \frac{dF_n}{dp} + \frac{dG_n}{dp} \right) - \frac{\varepsilon_n}{\Gamma} (Q_n + S_n), \]

(\text{A6})

and

\[ \rho \varepsilon_n^{-1} \omega_n = -4\Omega^2 a^2 \frac{d\omega_n}{dp} = -2\Omega a (F_n + G_n), \]

(\text{A7})

where \( \varepsilon_n \) is the \( n \)-th eigen value of the zonal Hough function.

[46] Equations (A6) and (A7) are diagnostic equations; thus \( \omega_n \) and (then \( \omega, v, \partial u/\partial t, \partial T/\partial t \)) are uniquely determined for the given external forcings. For example, if \( G = S = 0 \) and \( F \) and \( Q \) are constant in time, \( \omega, v, \partial u/\partial t, \partial T/\partial t \), are also constant in time.

[47] If frictional forcing \( G \) and diabatic heating \( S \) are introduced in the form of Rayleigh friction and Newtonian cooling as

\[ G = -\lambda u, \]
\[ S = -\alpha T, \]

(\text{A8})

then \( \partial u/\partial t \) and \( \partial T/\partial t \) are no longer determined by eddy forcings without considering the change of \( u \) and \( T \), and \( u \) and \( T \) must be determined by integration of the equations, as:

\[ \frac{\partial u_n}{\partial t} = 2\Omega \nu_n - \lambda u_n + F_n, \]
\[ \frac{\partial T_n}{\partial t} = \Gamma \omega_n - \alpha T_n + Q_n. \]

(\text{A9})

[48] The field coefficients are then determined by the time integration of prognostic equation (A9) with the solution of the diagnostic equations (A6) and (A7) for each time step. Field variables at any time step can then be calculated from equation (A5).

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Figure 7. Same as Figure 2 except for mechanical (top) and thermal (bottom) eddy forcings with the Eulerian mean meridional circulation (arrows) due to the PC1. Contour interval is 0.2 ms$^{-1}$ day$^{-1}$ (0.2 K day$^{-1}$) for the mechanical (thermal) eddy forcing. The horizontal (vertical) reference arrow indicates $6 \times 10^{-2}$ ms$^{-1}$ ($3 \times 10^{-2}$ ms$^{-1}$). Small vectors are neglected.
Figure 14. Zonal-mean zonal wind (top), meridional circulation (middle), and sea level pressure (bottom) calculated by integrating the observed eddy forcings in the mechanistic model. Contour interval is 2 m s$^{-1}$ for zonal wind and 5 x $10^8$ kgs$^{-1}$ for the mass stream function, and the unit is hPa for sea level pressure. The horizontal (vertical) reference arrow indicates 6 x $10^{-2}$ m s$^{-1}$ (3 x $10^{-4}$ m s$^{-1}$). Small vectors are neglected. See text for details.