On the influence of the meridional circulation and surface pressure change on the Arctic Oscillation

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[1] Eddy-forced meridional circulation and the corresponding surface pressure change associated with the month-to-month variability of the Arctic Oscillation (AO) are examined in the framework of Eulerian mean dynamics and are compared with the AO-like variability associated with stratospheric vacillation known as the Polar-night Jet Oscillation (PJO). Surface signals associated with both the AO and the AO-like variability associated with the PJO are produced through the eddy-forced meridional circulation. In the case of the AO, however, a surface pressure change is found to be produced by meridional circulation driven mainly by the mechanical forcing of zonal wave number 2 or 3 and high-frequency transient eddies in the troposphere. This largely contrasts with the AO-like variability associated with the PJO, which is mainly produced by the eddy forcings of zonal wave number 1 in the troposphere and stratosphere. A close relationship among the eddy forcing, meridional velocity, surface pressure change, and AO index was found to exist not only for the month-to-month variability but also on a decadal timescale. The separation of the AO index into tropospheric and stratospheric components in the decadal timescale revealed that the recent increasing trend mainly comes from the stratosphere whereas the decadal variation comes from the troposphere.


1. Introduction

[2] Recent observations indicate that the dominant mode of the month-to-month or winter-to-winter variability of the sea level pressure (SLP) of both the Northern and Southern Hemispheres shows a similar structure called the Annular Mode (AM) [Thompson and Wallace, 1998, 2000a]. The AM has a remarkable zonal symmetry in both hemispheres despite the larger zonal asymmetries of the land-sea distribution in the Northern Hemisphere (NH) than in the Southern Hemisphere (SH). The Arctic Oscillation (AO), also known as the Northern Annular Mode (NAM), is the NH counterpart of the AM and has a very large impact on the formation of the NH climate.

[3] Although the concept of the AO or the AM is new, the variability of the zonal wind associated with the AM or the dominant mode of variability in the zonal wind has been an important issue for a long time for observational as well as modeling studies [e.g., Nigam, 1990; Hartmann and Lo, 1998; Robinson, 1991; James and James, 1992; Limpasuvan and Hartmann, 2000; Lorenz and Hartmann, 2001, 2003]. Most of these studies indicate the importance of the positive feedback between the zonal mean wind and eddies for the maintenance of the AM.

[4] As the AO is a signal that appears as the change of the SLP, variability in the meridional circulation accompanied with it is inevitable because the sea level pressure is equal to the air mass above the sea level. However, only a few studies have focused on the meridional circulation and SLP change for the AO, in contrast to the large numbers of studies on the wave mean flow interaction. In this paper, the AO is examined from the viewpoint of variability in the meridional circulation and associated changes in the SLP.

[5] The AO pattern is characterized by a negative anomalous SLP center at the polar cap surrounded by two positive centers at the Northern Pacific and Atlantic. A similar type of variability in the SLP appears when a stratospheric variability called the Polar-night Jet Oscillation (PJO) propagates from the stratosphere to the surface [e.g., Baldwin and Dunkerton, 1999; Kuroda and Kodera, 1999; Christiansen, 2001]. As the appearance of the AO and the AO-like variability that is related with the PJO is very similar, they are sometimes regarded as the same thing. However, the largest lagged correlation between the time coefficients of the AO and the PJO (called the AO and PJO indices) is only 0.35; then, at most, only 10% variance of the AO is explained by the stratospheric process (PJO) [Kuroda and Kodera, 2004, hereinafter referred to as KK04]. In other words, most of the variability of the AO should originate in the troposphere. This means that the AO-like variability associated with the PJO and the AO is almost independent phenomena. In fact, an AO-like pattern associated with the PJO has a smaller polar center and no center of action in the Northern Pacific [KK04].

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[6] In a previous study [KK04], the source of meridional circulation and its effect on SLP for the PJO were examined within the framework of Eulerian mean dynamics, which is superior to the well-used transformed Eulerian mean method for the evaluation of changes in surface pressure. As both the AO and the AO-like variability associated with the PJO are characterized by the surface pressure change, the same analysis methods as those used in KK04 can be applied to the AO to clarify the origin of its surface pressure variability. This also enables us to compare similarities and differences between these two phenomena. We will perform these in this paper.

[7] The index of AO has shown an increasing trend in recent years [e.g., Thompson and Wallace, 2000b]. To clarify the origin of the long-term variation of the AO index, the main results of the present analysis with a month-to-month timescale are extended to long-term variability. The role of the stratosphere on the long-term variation of the AO index will also be clarified by this analysis.

[8] This paper is organized as follows. The data set and the principal method of analysis are described in section 2. Section 3 contains the results of the principal analysis. After a discussion, including a comparison of the stratospheric process and the long-term variability of the AO in section 4, the conclusion and remarks are presented in section 5.

2. Data and Method of Analysis

[9] We used reanalysis data from the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) [Kalnay et al., 1996]. The study covers a period that includes 21 winters from 1979/1980 to 1999/2000, while 6 months of extended winter from November to April were analyzed as used in KK04. This period is selected for the main analysis and for a comparison with the previous study (KK04) because the data have higher accuracy because of inclusion of satellite data. However, for comparison with the past two decades and for examination of the long-term variability of the AO, a NCEP/NCAR data set with a longer record was also used.

[10] The AO was defined from SLP north of 20°N following the original definition of Thompson and Wallace [1998] and was calculated from an Empirical Orthogonal Function (EOF) analysis of the month-to-month variability of the extended winter. Time coefficients of the AO in extended winters were calculated for every pentad (5-day mean) by projecting the AO pattern onto the 6-pentad running averaged anomalous fields.

[11] The lagged correlation and regression maps presented in this paper were calculated on the basis of the AO indices of the central 18-pentad period (from 15 December to 15 March) of the 21 years. For the calculation of the correlation coefficients, the year-to-year variation was neglected by removing the average of each year. Second-order quantities such as the Eliassen-Palm (E-P) flux were calculated on the basis of daily mean data and then they were 6-pentad running averaged.

[12] The mechanistic model we used to evaluate meridional circulation and surface pressure change due to wave forcings was the same as we used in KK04. It was originally used by Plumb [1982] and Haynes and Shepherd [1989]. This is a zonal mean quasi-geostrophic model and all calculations were performed by horizontal expansion by the Hough and its associated functions. In the model, a proper lower-boundary condition of \( \frac{\partial F}{\partial z} = 0 \) is adopted to evaluate the surface pressure change, where \( \Phi \) is the zonal mean geopotential height and \( \frac{\partial F}{\partial z} \) is a material derivative. We used 100 vertical levels with a width of 10 hPa for the pressure coordinate, and a horizontal grid of 121 equal widths for the sine of the latitude (see the Appendix of KK04 for details).

3. Results

3.1. Time Evolution of the AO

[13] Figure 1 shows the lagged regression map of the zonal mean zonal wind with the E-P flux (Figure 1, top), SLP (Figure 1, middle), and zonal mean SLP (Figure 1, bottom) calculated from the AO index. Areas of 95% significance level (correlation greater than 0.25) are shaded. The statistical significance was calculated under the assumption that each month of the reference time series (total \( 3 \times 21 \)) was independent. Here, we combined the satellite data used in KK04 above 10 hPa to examine the signal in the upper stratosphere as well.

[14] For the time evolution of the AO (Figure 1), a stronger zonal mean zonal wind first appears at the upper stratosphere at lag \(-30\) days, although it is not significant. The core of the zonal wind propagates downward with time with a significant area extending up to 1 hPa at lag \(-15\) days. As time advances, the core of the zonal wind continues to propagate downward (lag \( 0 \) to lag \( 15 \) days) and reaches 30 hPa with a significant area extending to 10 hPa at lag \( 15 \) days. It further propagates downward and stays at the lower stratosphere, although the significant signal disappears at lag \( 30 \) days. The downward propagation of the AO signal from the stratosphere means that the present AO signal includes that related to the stratospheric process [e.g., Baldwin and Dunkerton, 1999; Kuroda and Kodera, 1999]. Corresponding to the variation of the zonal wind, the anomalous E-P flux has a tendency to be directed toward the weaker low-latitude zonal wind from lag \(-15\) to \(-15\) days. Although some variability in the E-P flux exists in the stratosphere, most of the wave variability is restricted to within the troposphere.

[15] For the SLP signal, a small negative signal appears in the north Atlantic at lag \(-30\) days. It propagates northward while enlarging and creates a negative Arctic center at lag \(-15\) days. At the same time, the positive signal in the Arctic at lag \(-30\) days shifts to the North Pacific while enlarging, and creates a North Pacific signal at \(-15\) days. A small positive signal that appears at lag \(-30\) days at the east coast of the United States also becomes larger with an extension toward western Europe, and a weak AO-like signal is created at lag \(-15\) days. The Atlantic center shifts further northward and the signal becomes mature at lag \( 0 \). In the peak period, a typical AO signal with polar center around Iceland and two midlatitude centers around the East Atlantic and North Pacific appear. As the lag days increase, the North Atlantic signal retains its strength until day \( 10 \) and then gradually weakens. Compared with the Atlantic signal, the signal in the North Pacific weakens rapidly. As a result, a signal similar to the North Atlantic Oscillation (NAO) is...
created at lag 15 days. The positive signal in the Atlantic starts to shift northward at lag 20 days then begins to weaken, but retains significant strength at lag 30 days. At the peak, the structure of the AO is symmetrical and the zonal mean SLP shows anomalies of $-3.5 \text{ hPa}$ at the pole in contrast with $1.8 \text{ hPa}$ in the midlatitude belt.

### 3.2. Wave Forcing, Meridional Circulation, and Surface Pressure Change

[16] Since the structure of the AO is approximately zonally symmetric at the mature stage, it is valuable to see how meridional circulation and surface pressure change are explained by zonal mean dynamics. Although transform Eulerian mean (TEM) diagnosis is usually used, it is not efficient, especially in the lower troposphere because of the existence of the surface and the difficulty of applying appropriate lower boundary conditions. Therefore we used Eulerian mean (EM) diagnosis instead. In this framework, lower boundary conditions can be applied very simply [Haynes and Shepherd, 1989]. In EM dynamics, there are two types of eddy forcings: mechanical (momentum), which primarily accelerates zonal wind, and thermal, which forces temperature. Note that mechanical forcing is just the meridional divergence in the meridional component of the E-P flux, but thermal forcing is proportional to the meridional divergence in the vertical component of the E-P flux.

[17] Figure 2 shows the lagged regression of the EM eddy forcings (contours) and the meridional circulation (vectors) with the AO index. The eddy mechanical forcing (Figure 2, top) shows a meridional dipole pattern at the upper troposphere. Although the forcing center has a tendency to shift slightly upward with time, this pattern is rather persistent with peak amplitude at lag 0. The negative center is around $30^\circ \text{N}$ and $250 \text{ hPa}$ and the positive center is around $65^\circ \text{N}$ and $300 \text{ hPa}$ at lag 0. These forcings come from the meridional gradient of the eddy momentum flux, but the polar-side center dominates because of the contribution of the metric factor. For the eddy thermal forcing, it appears in the lower troposphere but it is significant only around the mature stage of lag 0.

[18] Meridional velocity has a tendency to be directed equatorward (poleward) in an area of positive (negative) mechanical forcing. For vertical velocity, positive (negative) thermal forcing corresponds well with upward (downward) flow at higher latitude, although it is worse at lower latitude. [19] To evaluate the strength of the signal of meridional circulation with the AO, we calculated the correlation...
coefficients of the meridional and vertical velocities at lag 0 (Figure 3). The meridional velocity shows an apparent quadrupole structure. The maximum value in the upper troposphere is $0.73$ at $55^\circ N$ and $250$ hPa, and in the lower troposphere it is $0.77$ at $55^\circ N$ and $1000$ hPa. For the vertical velocity, the signals are not so strong, but the downward center is located around $40^\circ N$ and the upward center around $70^\circ N$.

To see the effect of eddy forcings on the meridional circulation and surface pressure change associated with the AO, we used the quasi-geostrophic EM model on a sphere as a diagnostic tool as in KK04. Regression of the eddy forcing was used to examine the meridional circulation and surface pressure change associated with the AO. The equations were applied to the anomalous fields. The stability of the mean fields was estimated from the winter mean (December to March) temperature.

Figure 2 shows the meridional circulation and surface pressure tendency calculated from the time varying eddy forcings in Figure 2. There is a weak clockwise meridional circulation centered at $60^\circ N$ at lag $-30$ days. As the lag days proceed, a counterclockwise circulation appears from the surface that develops toward the upper troposphere (lag $-15$ days). The circulation peaks at lag $0$ and the amplitude gradually reduces but extends further toward the stratosphere (lag $15$ to $30$ days). Accompanying the time evolution of the main cell around $60^\circ N$, a clockwise subtropical cell is prominent from lag $0$ to $30$ days. The results of the calculation reproduce the overall meridional circulation in the upper troposphere to the polar area in the observation, although it is not as well reproduced in some areas such as in the subtropical troposphere around $40^\circ N$.

Corresponding to the change of meridional circulation, a small positive tendency of the SLP at the pole cap at lag $-30$ days becomes negative (lag $-15$ days) and peaks with a value of $-0.8$ hPa/day (lag $0$), which gradually increases with increasing lag days. From lag $-15$ to $15$ days, the positive SLP tendency area around $50^\circ N$ evolves along with the negative one at the polar cap.

It should be noted that the SLP tendency calculated here contains the activity of the eddy forcings on the SLP alone and does not include frictional effects, which have the opposite effect on the SLP tendency as will be seen later. However, because the frictional effect always follows a given forcing, the calculated SLP tendency should be a good indicator for the actual SLP anomaly [KK04]. In fact, a comparison of Figures 1 and 4 suggests that the observed SLP anomaly is nearly proportional to the eddy-forced SLP tendency. The relationship can be derived from the following two assumptions: (1) SLP tendencies from eddy forcings are large and positive, and (2) the SLP tendency is proportional to the eddy forcing. This assumption is supported by the observed data [KK04].
and friction are nearly balanced and (2) an SLP tendency created from friction is nearly proportional to an anomalous SLP multiplied by $-1$. In fact, the observed SLP tendency was very small in comparison with the calculated tendency shown in Figure 4 (not shown).

[24] To see the roles of respective components of the wave forcings on meridional circulation and surface pressure change, we applied the mechanical and thermal wave forcings separately. Figure 5 shows the result at lag 0. It can be seen that overall features of the meridional circulation by total eddy forcings are reproduced well by the mechanical forcing alone (Figure 5, top left), although the extension of the circulation toward the stratosphere is somewhat diminished. This results from the dominance of the mechanical forcing compared to the thermal one on the formation of AO [Limpasuvan and Hartmann, 2000]. Corresponding to the similarity of the meridional circulation, SLP change is also reproduced well by the mechanical forcing alone (Figure 5, bottom left). Compared with meridional circulation associated with mechanical forcing, that from thermal forcing (Figure 5, top right) shows a weak meridional circulation corresponding with weaker forcing, although it shows a remarkable extension toward the stratosphere. SLP change is also relatively small compared with that of mechanical forcing except for that around 55°N (Figure 5, bottom right).

[25] To examine the role of wave forcings from the respective zonal wave number components, we decomposed waves into zonal wave number components and applied the respective wave forcings associated with these waves. Figure 6 shows the result at lag 0. The major contribution of the meridional circulation in the troposphere and the surface pressure change arises from zonal wave number 2 (WN2) and higher wave number components (WN4+). Although meridional circulation at the stratosphere is strongly affected by zonal WN1 component, the effect in the troposphere is very small. Almost half of the polar SLP change comes from the higher wave number components. [26] To see the role of high-frequency transient waves, we defined the high-frequency transient wave as the difference between the pentad and daily mean values. Figure 7 shows the eddy forcing contribution from the high-frequency transient component with the calculated meridional circulation and the SLP change. The overall features of the wave forcings and the resultant meridional circulation and surface pressure change are similar to those from the high wave number components (WN4+) in Figure 6, although some minor differences exist. This shows that most of the high wave number components belong to the high-frequency transient eddies, as would be expected.

[27] We also evaluated the noneddy forcing for its effect on the meridional circulation and surface pressure change associated with the AO (Figure 8). Noneddy forcings were determined from the residual of the three-dimensional primitive equations by using a 6-hourly data set: frictional forcing was estimated from the zonal wind equation whereas diabatic heating was estimated from the thermodynamic...
equation. In Figure 8, areas of higher diabatic heating (greater than 0.05 Kday$^{-1}$) are indicated by shading. Frictional forcing near the surface and diabatic heating in the lower troposphere around 55°N and cooling around 40°N were prominent at lag 0 (Figure 8). The meridional circulation in the lower latitudes is reproduced well by the estimated noneddy forcings. The increasing surface pressure tendency is approximately balanced by the decreasing tendency created by the eddy forcings. Separation of the noneddy forcings into frictional and diabatic forcings reveals that meridional circulation in the subtropics comes from the combined effect of frictional and diabatic forcings, but most of the surface pressure change comes from the frictional forcing (not shown).

Although the surface pressure tendency estimated from the frictional forcings is too large compared with that from the wave forcings, the frictional forcings calculated above should be regarded as qualitative because the data were produced through violation of the balance of the dynamical equation in the reanalysis process. There is also the problem that the lowest pressure level (1000 hPa) sometimes intersects the Earth’s surface and artificial forcings are analyzed at the lowest level.

To evaluate the total effect of forcings on the meridional circulation and the surface pressure change consistently in the framework of the model, we integrated the EM quasi-geostrophic model with time. The effect of friction is included as a function of Rayleigh friction ($\lambda$) against the zonal wind, and the effect of heating as a function of Newtonian cooling ($\alpha$). The functional dependencies of these coefficients on height ($z$) are given the same forms as those used in KK04:

$$\lambda = R_0 \exp(-z/600) + R_1 \{1 + \tanh[(z - 20 000)/5000]\}$$  \hspace{1cm} (1)

$$\alpha = N_1 \{1.5 + \tanh[(z - 25 000)/7000]\}$$  \hspace{1cm} (2)

Here, the constant values are set as $R_0 = 1/(0.5 \text{ days})$, $R_1 = 1/(30 \text{ days})$, $N_1 = 1/(10 \text{ days})$, and $z$ is expressed in meters.

All parameter values are the same as those used in KK04. The radiative dumping scale in the troposphere is 20 days, but it becomes 4 days in the upper stratosphere with the value of $N_1$ adopted here. A small value of $R_1$ is introduced for the stability of the integration. Because the diabatic heating associated with the AO is quite large in the troposphere (Figure 8), we included regressed diabatic forcing in the troposphere (below 100 hPa) in addition to that calculated in the model through equation (2). Time integration was then performed by the Matsuno (Euler backward) scheme, with a time step of 0.2 days and the initial conditions of day $-35$ (Figure 9). Here we used a
version of the model with 50 vertical levels with a layer thickness of 20 hPa, because a model with 100 vertical levels failed to integrate probably because of very large gradient of the circulation at the surface.

A meridional dipole structure of zonal mean zonal wind first appears in the upper stratosphere at lag $/C0_{30}$ days, and then it propagates down to the surface at lag 0 and persists until lag 15 days. At lag 0 to 15 days, a

Figure 8. Same as Figure 4 except for the contribution from noneddy forcings at lag days of $-30$, 0, and $+30$. Shading indicates areas of larger diabatic heating (greater than 0.05 Kday$^{-1}$). See color version of this figure in the HTML.

Figure 9. (top) Zonal mean zonal wind, (middle) meridional circulation, and (bottom) sea level pressure calculated by integrating the zonal mean quasi-geostrophic mechanistic model with observed eddy forcings and tropospheric diabatic heating associated with the AO. Contour interval is 1 ms$^{-1}$ for zonal wind and $5 \times 10^8$ kgs$^{-1}$ for the mass stream function. The unit for sea level pressure is hPa. The horizontal (vertical) reference arrow indicates $6 \times 10^{-2}$ ms$^{-1}$ ($3 \times 10^{-4}$ ms$^{-1}$). Small vectors are neglected. See text for details. See color version of this figure in the HTML.
meridional dipole structure with a node at around 40°N is very prominent in the troposphere. Its low-latitude center exists around 30°N at 150 hPa and its high-latitude center around 55°N at 300 hPa. This feature is very similar to what is observed (Figure 1), although the zonal wind profile is somewhat modified, especially in the upper stratosphere.

[31] The overall strength of the meridional circulation becomes stronger than what is obtained from eddy forcings alone (Figure 4), and it becomes comparable to what is observed (Figure 2) due mainly to the inclusion of surface friction and diabatic heating in the troposphere. Corresponding to meridional circulation, surface pressure shows a meridional dipole structure with lower pressure over the polar cap and higher pressure in the middle latitudes. The surface pressure has a minimum value of −4 hPa on the polar cap region at lag 0 corresponding to the strongest meridional circulation. Although this value is a bit larger than what is observed, the overall features of the surface pressure are well reproduced by this model.

4. Discussion

4.1. Comparison With the PJO

[32] We found that surface pressure change associated with the AO is produced through meridional circulation driven by eddy forcings mainly by the mechanical forcing of waves of the zonal WN2 component and high-frequency transient waves. The contribution to meridional circulation from thermal forcings was small in the troposphere. For the formation of the surface pressure anomaly in the AO, the effects of frictional forcing against zonal wind near the surface also play a significant role. Note that, for the estimation of the surface pressure change from forcings, setting the proper lower boundary condition \((D\Phi/Dt = 0)\) is crucial [Haynes and Shepherd, 1989].

[33] The mechanism for the creation of the hemispheric near zonally symmetric SLP anomaly is very similar between the usual AO and the AO-like variability associated with the PJO. Both signals are created through meridional circulation that is driven by the eddy forcings. However, in the case of the AO, the major part of the eddy forcings arises mainly from the mechanical forcing of the low-frequency WN2 and transient high-frequency waves. This is in contrast with the PJO-related variability, whose eddy forcings come mainly from both types of eddy forcing of low-frequency WN1 [KK04]. In the case of PJO-related variability, the dominance of WN1 should come from its connection to the stratospheric variability because the only wave component that can propagate into the upper stratosphere and cause a wave mean flow interaction in midwinter is WN1 [e.g., Charney and Drazin, 1961; Matsuno, 1970]. In contrast, the dominance of the mechanical forcing of high-frequency transient waves in the usual AO variability is quite understandable because mechanical forcing dominates the thermal one for the synoptic transient wave [Pfeffer, 1987]. The results suggest that an AO-like SLP signal appears through a similar mechanism regardless of the dominance of the wave number in the quasi-stationary wave which creates different types of climatic variability (AO or PJO). [34] A part of the AO-like variability should be related to the variability of the stratospheric process (PJO), since the downward propagation of the zonal wind signal can be seen from the time evolution of the AO (Figure 1). Therefore we defined the AO-like signal that is not related to the PJO (hereinafter, the time coefficient is called the tropospheric AO index) by statistically subtracting the component related to the PJO [Kuroda, 2002]:

\[
r' = \frac{1}{\sqrt{1 - (\langle r_s \rangle)^2}} (r - \langle r_s \rangle s).
\]

Here, \(r\) and \(s\) are the normalized AO and PJO indices, respectively, and the bracket indicates the time averaging. It is clear that the new standardized time coefficient, the tropospheric AO index, \(r'\), has no correlation with that of the PJO. Here, the PJO index is taken as the time coefficient of the EOF1 of the averaged polar temperature from the surface to 1 hPa, which was calculated in KK04.

[35] The results of the time evolution of the zonal mean zonal wind with E-P flux and SLP for the tropospheric AO are shown in Figure 10. The time evolution of the zonal wind (Figure 10, top) shows that the signal of the tropospheric AO does not propagate down from the stratosphere as a typical AO (Figure 1), but appears first in the troposphere (lag −15 days) and propagates into the stratosphere (lag 15 to 30 days) after the mature period of the AO. The time evolution of the horizontal structure of SLP shows that the signal first appears in the North Pacific as well as in the Arctic region (lag −15 days), and the North Pacific signal extends to the North Atlantic area and creates the matured AO signal (lag 0). The signal on the Pacific side decreases its strength but that on the Atlantic side retains its strength until lag 15 days.

[36] The time evolution of the tropospheric AO is consistent with the results by Kodera and Kuroda [2000] for the time evolution of their tropospheric AO, and it is also very similar to that of the Aleutian-Iceland seesaw (AIS) examined in the series of papers of Honda, Nakamura, and their collaborators [Honda et al., 2001; Honda and Nakamura, 2001; Nakamura and Honda, 2002]. Its horizontal time evolution from Pacific to Atlantic is also reminiscent of the work by Simmons et al. [1983, Figure 11] for the barotropic instability originating in the Pacific region. These results suggest that the typical AO is a mixture of tropospheric and stratosphere-related signals, and the nature of the tropospheric AO is a Pacific-to-Atlantic traveling mode that is related to barotropic instability. However, more research will be needed to confirm this result.

4.2. Multidecadal Variations in the AO

[37] We used only the most recent 22 years of data in the present analysis, because sufficiently large samples can be obtained from a month-to-month analysis and better analysis of the observational data can be expected from the satellite observations made during this period. However, data in the NH should be reliable since around the year 1958. Thus 20 more years will be available, although the accuracy will be somewhat less because of the absence of satellite data for the NH. It is interesting to note the difference in the AO signals between the two
decades. Therefore we performed the same analysis using data from 1958/59 to 1978/79. The resulting time evolution of the zonal wind with the E-P flux and SLP is shown in Figure 11.

The overall time evolution and the matured AO pattern are similar. However, a visual inspection shows that there are some significant differences with recent ones (Figure 1). The downward propagation of zonal wind is as prominent as in recent ones, but it is weaker (lag $-15$ to $15$ days). It is also apparent that the extension of the AO signal toward the stratosphere is smaller even at the mature stage of lag 0. In addition, the anomalous horizontal propagation of the E-P flux at lag $-15$ days and at lag 30 days is very prominent.

A weak negative SLP signal similar to recent ones first appears in the North Atlantic at lag $-30$ days and it propagates northward while enlarging. However, another small negative signal appears east of Scandinavia at lag $-25$ days and gradually increases in value almost without propagation. As a result, these two signals merge (lag $-15$ days) and create a strong Arctic center at lag 0. At the same time, a positive weak signal over Greenland at lag $-30$ days separates into two pieces: one moves over the Mediterranean Sea and the other over Japan with
enlargement at lag −15 days. These two midlatitude centers further extend toward the North America and create an AO structure at lag 0. At the mature stage of lag 0, the Arctic center is more extended toward the direction of China and the United States, and the signal at midlatitude is weaker compared with the recent AO pattern (Figure 1). Although the signal in the Arctic weakens quickly with increasing lag days, the midlatitude signal retains its strength until lag 15 days. As a result, the SLP pattern becomes similar to that of the recent AO at lag 15 days except that the southern anomaly shifts further eastward. The positive signal at the Atlantic propagates more toward the Arctic and has a marginal strength with a small positive signal in northern Siberia at lag 30 days.

[40] Compared with recent AO, the time evolution seems to be rather different. First, there seems to be no signal associated with the AIS [Honda and Nakamura, 2001; Honda et al., 2001]. Second, the signal that appears in northern Siberia is very prominent. Third, coupling with the stratosphere is weaker. For the first point, Honda et al. [2005] noted that AIS variability was very small during the 1960s. There may be other causes, such as the regime shift that appeared in the late 1970s [e.g., Nitta and Yamada, 1989; Trenberth, 1990].

[41] To examine whether the difference feature of the AO (the larger area of the polar center in the older period than the recent one at lag 0) is significant or not, we used the Monte Carlo test: we randomly chose 1000 samples of 20 winters from 43 winters from 1958 to 2000 and compared the AO patterns among them. It was found that only 3.6% of the samples had wider polar centers than the older AO pattern, whereas only 0.7% had a smaller polar center than the recent AO pattern. This indicates that the difference between recent and older AOs is statistically significant.

[42] Wave forcing analysis indicates that the effect of eddy forcings on the surface pressure change is very small (−0.6 hPa/day) in spite of deeper zonal mean SLP (−4.5 hPa) at the polar cap at the mature stage (not shown). This may partly come from poorer analysis of waves in this period. It also indicates that about half of the surface pressure change comes from the eddy forcing of high-frequency synoptic waves and one third comes from waves of zonal wave number 3 (WN3). The contribution from WN2 was found to be very small. The high activity of WN3 should be related to the presence of the northern Siberian signal on the SLP signal during this period. This result suggests that the presence of WN2 is not necessarily crucial for the formation of the AO.

4.3. Relationship Among AO-Related Indices and Long-Term Variation

[43] Investigation of the formation of the AO shows that the surface pressure anomaly associated with the AO is driven by near zonally symmetric meridional circulation, which is primarily driven by mechanical forcing. This result suggests that the AO index has a good correlation among the polar cap SLP anomaly and mean meridional and vertical velocities, as well as the mean mechanical forcing, or momentum transport. Therefore we defined the following: (1) polar cap SLP index (Pol index), (2) meridional velocity index (V index), (3) vertical velocity index (W index), (4) mechanical forcing index (M index), and (5) momentum transport index (MT index) in the central 18-pentad period (from 15 December to 15 March) of 21 extended winters. In addition, to see the relationship with the stratospheric process, the correlations with the (6) PJO index were also calculated. The definitions of these indices are listed in Table 1. Simultaneous correlations of these indices with the AO or PJO indices are listed in Table 2.

[44] The results show that the AO index has very large simultaneous correlations among the V, M, MT, Pol, and W indices, as expected. The appearance of the largest correlation with the M and MT indices at about lag −3 days (the M and MT indices lead the AO index), and the V index at about −2 days (not shown) corresponds well with the causal order of the formation of AO. The significant correlation between the AO and PJO indices indicates that the AO signal includes some stratospheric variability, as discussed before and in KK04. However, it is noteworthy that the total variance from the PJO contributes only about 10% of the AO. From the results in Table 2, it can be determined that about 25% of the polar SLP variance comes from the PJO, whereas about 60% (40%) of the polar SLP variance comes from the total AO (tropospheric AO). We believe that the variability of the polar SLP is the key factor of the AO paradigm.

[45] The correlation coefficients calculated here were based on month-to-month variability and the contribution associated with interannual variation was removed. However, these indices may be linked together on the interannual scale as well if the same interrelationship exists for a longer variability. Thus we calculated 6-year running averaged Pol, V, and MT indices and compared them with those of the AO index calculated from the longer record of 1958 to 2001 (Figures 12a–12d). Calculations were based on standard-ized monthly indices. The overall features of the decadal variability of the indices are very similar to each other. Higher peaks around 1965, 1973, and 1992 and lower valleys around 1968, 1980, and 1998 are prominent in all indices. A long-term trend is also apparent in all indices, although it is somewhat weaker in the MT index. In this

### Table 1. Definition of Indices Concerned With the AO and the PJO

<table>
<thead>
<tr>
<th>Index</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pol</td>
<td>mean sea level pressure north of 60N, multiplied by −1</td>
</tr>
<tr>
<td>V</td>
<td>mean meridional velocity north of 45N and levels from 400 to 200 hPa, multiplied by −1</td>
</tr>
<tr>
<td>W</td>
<td>mean vertical velocity north of 60N and levels from 600 to 400 hPa</td>
</tr>
<tr>
<td>M</td>
<td>mean mechanical forcing from 50N to 75N, and 700 to 100 hPa</td>
</tr>
<tr>
<td>MT</td>
<td>mean momentum transport from 40N to 60N, and 700 to 100 hPa</td>
</tr>
<tr>
<td>PJO</td>
<td>time coefficient of the EOF1 of the mean polar temperature up to 1 hPa (see KK04)</td>
</tr>
</tbody>
</table>

### Table 2. Simultaneous Correlations of the Indices

<table>
<thead>
<tr>
<th></th>
<th>Pol</th>
<th>V</th>
<th>M</th>
<th>MT</th>
<th>W</th>
<th>PJO</th>
<th>AO</th>
</tr>
</thead>
<tbody>
<tr>
<td>AO</td>
<td>0.77</td>
<td>0.87</td>
<td>0.79</td>
<td>0.80</td>
<td>0.67</td>
<td>0.34</td>
<td>1</td>
</tr>
<tr>
<td>PJO</td>
<td>0.49</td>
<td>0.43</td>
<td>0.44</td>
<td>0.41</td>
<td>0.52</td>
<td>1</td>
<td>0.34</td>
</tr>
</tbody>
</table>
sense, some part of the origin of the long-term variability of the AO index may be traced back to that of the MT index. Therefore it is necessary to examine what controls the long-term variability of the eddy momentum transport in the upper troposphere to explain the long-term variability of the AO index.

Some of the long-term variability of the AO also will have a relationship with the stratospheric process as a month-to-month variability. Although the PJO index cannot be calculated before 1978 because of the absence of upper stratospheric data, we instead calculated the monthly PJO_L index as a projection of the polar temperature pattern defined for PJO (KK04) onto the anomalous polar temperature of the reanalysis data. Correlation of this index with the PJO index for the period from 1979 to 2000 was 79%, so this index can be regarded as a good extension of the AO index over a longer period. The 6-year running averaged PJO_L index is also compared with the other indices in Figure 12e. The increasing trend and decadal variability are also prominent in this index. Particularly, the increasing trend is more prominent. Although the decadal variability is not so prominent as in other indices, there exist three decadal peaks as in other indices. Peaks and valleys before 1975 seem to match well but others seem to be delayed. Since the monthly AO index is well correlated (0.37) with the PJO_L index, the AO index without the effect of the PJO_L (called the tropospheric AO index for longer data) can be calculated by statistical subtraction, as in equation (3):

\[
r'_{L} = \frac{1}{\sqrt{1 - C^2}} (r - C s_L),
\]

where \(C(=0.37)\) is the correlation between the AO and PJO_L indices, and \(s_L, r, \) and \(r'_{L}\) are the standardized PJO_L index, AO index, and tropospheric AO index, respectively. The 6-year running average of the tropospheric AO index is also shown in Figure 12f. It can be seen that the decadal

Figure 12. Long-term variability of the (a) AO index, (b) Pol index, (c) V index, (d) MT index, (e) PJO_L index, and (f) tropospheric AO index from 1958 to 2001. All indices are standardized by the month-to-month variance and then 6-year running averaged. See text for details.
variability of the tropospheric AO index is more prominent but the increasing trend is much smaller. A smaller trend of the tropospheric AO index can easily be seen from equation (4): if the trend of index drift, \( T(A) \), is defined by some linear operation of \( A \), such as the difference of the average of \( A \) for the newest and oldest periods, we have

\[
T(r') = \frac{1}{\sqrt{1-C^2}} (T(r) - CT(s_l)).
\]  

(5)

Therefore, if \( T(s_l) \) is largely positive, \( T(r') \) should be much smaller than \( T(r) \). The result suggests that the decadal variability of the AO originates from the tropospheric process; however, a major part of the increasing trend comes from the stratosphere for long-term variability.

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