Meridional and Downward Propagation of Atmospheric Circulation Anomalies.  
Part I: Northern Hemisphere Cold Season Variability

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ABSTRACT

The Northern Hemisphere cold season circulation anomalies are diagnosed in a semi-Lagrangian θ-PVLAT (potential vorticity surfaces as latitudes) coordinate by following contours of the daily potential vorticity (PV) field on isentropic (θ) surfaces using the NCEP–NCAR reanalysis II dataset from 1979 to 2003. It is found that circulation anomalies propagate poleward and downward simultaneously in the stratosphere and equatorward in the extratropical troposphere. On average, it takes about 40 days for warm anomalies and 70 days for cold anomalies to travel from the equator to the Pole. The beginning of the equatorward propagation of the tropospheric temperature anomalies coincides with the arrival of the poleward and downward propagating temperature anomalies of the opposite sign over the polar stratosphere. Accompanied with warm (cold) anomalies is a successive leveling (steepening) of isentropic surfaces propagating poleward and downward. The zonal wind anomalies follow the poleward and downward propagating temperature anomalies of the opposite sign.

A global mass circulation paradigm is proposed to qualitatively explain the simultaneous meridional and downward propagation of circulation anomalies that appears responsible for the annular mode variability. The meridional propagation of circulation anomalies can be viewed as an intensity variation of the zonally averaged isentropic mass circulation. When the mass circulation is weaker, the isentropic surfaces in the extratropical stratosphere (troposphere) are steeply (gently) sloped, corresponding to the positive phase of the annular mode. The cold air mass is effectively imprisoned within the polar cap when the mass circulation is weaker, responsible for warm surface temperature anomalies prevailing in the extratropics. Meanwhile, the weaker mass circulation also implies a temporary reduction of air mass supply over the polar cap, leading to a negative surface pressure anomaly. The warm anomalies brought by the stronger mass circulation cause a lowering of isentropic surfaces in the polar stratosphere, resulting in more gently sloped isentropic surfaces in the extratropical stratosphere. This corresponds to the negative phase of the annular mode in which the meridional temperature gradient in the extratropical stratosphere is weaker, accompanied with a weakened polar jet and a falling of the tropopause. The stronger warm air branch of the mass circulation aloft requires a strengthening of the compensating equatorward advancement of the surface air mass, causing massive cold air outbreaks in the extratropics. The more air mass aloft brought by the stronger mass circulation contributes to a rising of the surface pressure over the polar cap till the surface cold air moves out. This explains why the surface pressure anomalies in high latitudes are positive during the negative phase of the annular mode.

The well-known apparent downward propagation of geopotential height and zonal wind anomalies into the troposphere from the stratosphere in the polar region can be explained as the local dynamic PV response to the arrival of the simultaneous poleward/downward propagating heating anomalies in the polar stratosphere and the compensating equatorward propagating tropospheric heating anomalies of the opposite sign, rather than suggesting the stratospheric origin of the anomalies. The apparent equivalent barotropic structure of the annular mode mainly results from the dynamic response to the heating anomaly that has an opposite polarity between the stratosphere and lower troposphere.

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1. Introduction

The Northern Hemisphere annular mode (NAM) is the most dominant recurrent pattern of the Northern Hemisphere (NH) atmospheric variability in winter with a time scale from a week to a few months (Thompson and Wallace 1998; Baldwin and Dunkerton 1999). It is characterized with a ringlike seesaw oscillatory pattern between the subtropics and extratropics in both thermal and momentum fields (e.g., Thompson and Lorenz 2004, and the references therein). The positive phase of the NAM is associated with a zonally quasi-symmetric polar vortex in the stratosphere. The negative phase corresponds to a weak stratospheric polar vortex with a strongly wavy flow surrounding the polar cap and is often accompanied with a stratospheric sudden warming event (Kuroda 2002; Zhou et al. 2002; Limpasuvan et al. 2004). Recently, there are a number of observational studies showing a poleward propagation of zonal mean zonal wind anomalies associated with the annular mode variability (Feldstein and Lee 1998; Kodera et al. 2000; Dunkerton 2000; Kuroda 2002). Simultaneous poleward and downward propagation of the relative angular momentum anomalies with a time scale of 40–50 days is also found in the tropical region (Anderson and Rosen 1983).

The Arctic/North Atlantic Oscillation (AO/NAO) is the dominant mode of climate variability near the surface. The positive phase of the AO, which is characterized with the presence of negative surface pressure anomalies over the polar region, coincides with the positive phase of the NAM and vice versa. Therefore, the AO is regarded as the surface signature of the NAM (Kodera and Kuroda 1990; Thompson and Wallace 1998; Baldwin and Dunkerton 1999; Thompson and Wallace 1999; Baldwin and Dunkerton 1999). The linkage of the NAM to the AO/NAO is characterized with a systematic downward propagation of geopotential height and zonal wind anomalies in the extratropics (Baldwin and Dunkerton 1999, 2001; Coughlin and Tung 2005). It is also known that in winter, severe cold air outbreaks in the extratropics accompanied with massive tropopause folding events are more (less) frequent during the negative (positive) phase of the AO/NAO (e.g., Thompson and Wallace 2001; Thompson et al. 2002; Cai 2003). The stratospheric connection to the AO and the associated extreme weather events bring out a new opportunity for intraseasonal and interannual climate predictions in winter seasons (Baldwin and Dunkerton 2001; Thompson and Wallace 2001; Baldwin et al. 2003).

The stratospheric polar vortex oscillation as well as the associated downward propagation signal has been recognized as a manifestation of a two-way coupling operating locally between the stratosphere and troposphere in the extratropics. In the view of the wave–mean flow interaction theory (e.g., Holton et al. 1995; Shepherd 2002; Haynes 2005; and the references therein), the extratropical troposphere is a source of planetary-scale Rossby waves that propagate upward into the stratosphere. When these waves break in the stratosphere, they transfer westward momentum to the mean flow and cause a deceleration of the stratospheric polar jet. The deceleration of the polar jet would in turn restrict the upward propagation of Rossby waves penetrating the critical level into the stratosphere. The lack of planetary Rossby waves in the stratosphere, together with local radiative cooling processes, would in turn result in a restoration of the westerly jet and polar vortex. The downward propagation of stratospheric anomalies in high latitudes has been interpreted as some kind of delayed feedbacks of the stratosphere to the upward propagation of tropospheric Rossby waves (Hartley et al. 1998; Limpasuvan and Hartmann 2000; Ambaum and Hoskins 2002; Black 2002; Perlwitz and Harnik 2003; Song and Robinson 2004). However, it is still debatable whether the systematic downward propagation indicates the direct influence of the stratosphere on the troposphere or merely reflects the signal propagation from the stratosphere to the troposphere. Also the local coupling of the stratosphere and troposphere in the extratropics alone seems not to be able to fully explain why the extratropical anomalies tend to be out of phase with anomalies in the subtropics/Tropics.

Part I of the series papers is an extension of the work reported in Cai and Ren (2006). Here, we present a comprehensive diagnostic study of atmospheric circulation anomalies in a semi-Lagrangian coordinate constructed using constant isentropic ($\theta$) and potential vorticity (PV) surfaces. The primary objective of this study is to present a physical explanation on the dynamical nature of the annular mode climate variability from the perspective of the global mass circulation. The remaining part of this paper is divided into eight sections. The next section describes the data used in this study. Section 3 discusses the spatial pattern and time scales of the dominant mode of daily NH PV anomalies in the $\theta$-PVLAT (potential vorticity surfaces as latitudes) coordinate as well as its relation with the NAM indices at various levels. In section 4, we present the temporal and spatial evolution of the composite circulation anomalies, showing a simultaneous poleward and downward propagation of anomalies in the stratosphere and equatorward propagation in the extratropical troposphere. In sections 5 and 6, we provide evidence showing that the propagation of circulation anomalies is intimately related to the global mass circulation variability. The
relation between the mass circulation, polar vortex, and surface pressure variability is discussed in section 7. Section 8 is devoted to reveal the linkage of the tropospheric circulation anomalies to the stratospheric anomalies from the perspective of the global mass circulation. A comprehensive summary of the main findings of this paper is presented in section 9.

2. Data and analysis procedures

The data used in this study are derived from the daily isentropic analysis of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis II covering the period of 1 January 1979 to 31 December 2003 (Kalnay et al. 1996; Kistler et al. 2001). The daily isentropic analysis includes PV, zonal and meridional winds, temperature, relative humidity, Montgomery potential, and Brunt–Väisälä frequency square (or the static stability parameter) on 2.5° × 2.5° grids at 11 standard isentropic surfaces (θ = 270, 280, 290, 300, 315, 330, 350, 400, 450, 550, and 650 K) and the surface potential temperature and pressure fields.

Following Norton (1994), we construct a semi-Lagrangian θ-PVLAT coordinate using the daily isentropic PV fields, where θ is the “vertical coordinate” representing vertically increasing constant potential temperature surfaces and PVLAT is the “meridional coordinate” representing northward increasing PV. The PVLAT coordinate is constructed by assigning a PV contour on an isentropic surface to a latitude value such that the area of the spherical cap encircled by the PV contour is identical to that encircled by the latitude. Details about the algorithm for mapping a PV field to the PV latitudes are reported in Ren and Cai (2006).

The daily 2D fields in the θ-PVLAT coordinate are obtained by averaging the original 3D daily isentropic analyses along PV latitudes according to

\[ [X]_{\phi_{\text{PV}}} = \int_{\phi_{\text{PV}}} X \cos \phi \, dl / \int_{\phi_{\text{PV}}} \cos \phi \, dl, \]

where \( \phi \) is the latitude, \( \phi_{\text{PV}} \) denotes PV latitude, and the generic variable “\( X \)” can be any meteorological parameter, such as PV, wind, temperature, Montgomery potential, and static stability parameter, and \( \int_{\phi_{\text{PV}}} \cos \phi \, dl \) is the line integral along a PV latitude.

It is known that the upper-level baroclinic zones are along the PV contours with the largest gradient (e.g., Hoskins et al. 1985; Hoskins 1997; Davies and Rossa 1998; Morgan and Nielsen-Gammon 1998). The θ-PVLAT coordinate can be regarded as natural boundaries separating air masses of different properties. Particularly, the “zonal mean” along the PV latitudes is very close to a “Lagrangian averaging,” capturing both the thermodynamic and dynamic properties of the same air mass more accurately compared to the conventional zonal mean along the geographical latitudes.

After obtaining the daily fields in the θ-PVLAT coordinate (total of 9131 days), the daily climatological annual cycle is obtained by first averaging the daily data on each calendar day across all years from 1979 to 2003, which yields a series of 365 maps (the data on 29 February in the leap years are excluded in calculating the annual cycle). Then a 31-day running mean operator is applied to the 365 consecutive maps to obtain smoothly varying annual cycle. The daily anomalies are obtained straightforwardly by removing the annual cycle from the total fields. Because both the daily annual cycle and anomalies are obtained in the θ-PVLAT coordinate, the anomalies defined in this fashion effectively are (semi) Lagrangian transients. Alternatively, one could first determine both the daily annual cycle and anomalies in the regular longitude–latitude coordinate on each isentropic surface and then obtain the anomaly fields in the θ-PVLAT coordinate by averaging anomalies along PV latitudes. The temporal anomalies obtained in this fashion are obviously the transients in the Eulerian sense but are viewed in the semi-Lagrangian θ-PVLAT coordinate. Nevertheless, the difference between the “Lagrangian” and “Eulerian” anomalies is very small except that the Lagrangian anomaly amplitude is slightly smaller than the Eulerian anomaly because the amplitude of the Lagrangian annual cycle is stronger than that calculated in the Eulerian sense.

3. The polar vortex oscillation index

Because of conservation properties of both potential temperature and PV, the advective tendencies of PV are naturally absent in the θ-PVLAT coordinate. As a result, the temporal variability of PV anomalies in the θ-PVLAT coordinate has a very rich low-frequency variability ranging from intraseasonal to interannual time scale with little signal at synoptic time scales. The first EOF mode of the daily PV anomalies in the θ-PVLAT coordinate explains about 69% of the total variance of daily PV anomalies over the entire NH, depicting a seesaw pattern between the midlatitude PV anomalies and the tropical/polar PV anomalies (Fig. 1). Moreover, in the polar and tropical regions, the stratospheric PV anomalies are negatively correlated with the tropospheric anomalies. The positive phase of the leading EOF mode corresponds to a stronger polar vortex and the negative a weaker polar vortex. For an easy reference, we refer to the daily time series of this mode.
(Fig. 2) as the polar vortex oscillation (PVO) index. It is seen that the daily PVO index exhibits only one–two PVO events in each winter season with little variability at synoptic time scale.

The power spectral analysis indicates that the daily PVO index exhibits a distinct peak spectrum at the time scale of about 107 days (Fig. 3). The peak amplitude at the time scales of about 5 months (9131/60 days) reflects the fact that the PVO mode is active only from late fall to early spring each year and are relatively quiescent in the rest of the year (Fig. 2). On top of the intraseasonal and seasonal time scales, the PVO index also has three peaks at periods of about 8.7, 16.9, and 33.8 months (9131/35, 9131/18, and 9131/9 days). Obviously, these three periods follow a double-periodicity sequence. These three periods correspond to the three-peak extratropical quasi-biennial oscillation (QBO) signal found by Tung and Yang (1994) from the 13-yr (1978–91) column ozone data (the three periods reported in Tung and Yang are 8.6, 20, and 30 months, respectively).

Figure 4 shows the lead/lag correlation between the PVO index and the daily NAM indices (http://www.nwra.com/resumes/baldwin/nam.php) at different vertical pressure levels from 1000 to 20 hPa. It is seen that as the height decreases, the axis of the maximum correlation tilts toward a shorter lag time of the PVO index with respect to the corresponding daily NAM index. Because the PVO index is associated with the leading EOF mode of the entire NH PV anomalies from the Tropics to the North Pole and from the stratosphere to the troposphere, such a change of the maximum correlation with the altitude merely reflects the downward propagation nature of the NAM. The maximum correlation is about 0.91 when the NAM index at 20 hPa leads the PVO index by about 11 days. The maximum instantaneous correlations of the PVO index with the NAM indices at 50, 70, and 100 hPa are 0.85, 0.79, and 0.7, respectively. The general decrease of the maximum correlation from the stratosphere to the troposphere is partly due to the high-frequency fluctuation of the daily NAM index at lower elevations. However, it is of importance to point out that the minimum of the temporal correlation between the daily PVO and NAM indices is found in the middle of the troposphere (at about 500 hPa) instead of at the surface. The maximum correlation between the PVO index and the daily NAM index at 1000 hPa is about 0.28, larger than that at 500 hPa despite the fact that the daily NAM index at 1000 hPa is somewhat noisier than that at 500 hPa. This seems to suggest that there could be an interruption of the downward propagation of anomalies from the stratosphere to the troposphere. We will present more concrete evidence in the next four sections showing the interruption of the downward propagation of circulation anomalies into the lower troposphere from the stratosphere.

4. Propagation of composite anomalies

a. The composite PVO event

To depict the temporal and spatial evolution of the circulation anomalies associated with the PVO index, we have constructed a “relative-intensity-based” composite PVO event based on the daily PVO index. Specifically, we define positive (negative) PVO events when the daily value of the PVO index is above 0.7 (below 0.7) standard deviation of the winter season PVO index. There are totally 31 positive and 25 negative PVO events during the 25-yr span (the color portions of the curves shown in Fig. 2). Each of the PVO events is then normalized by its own peak amplitude. The evolution of positive PVO events is binned into 40 intervals according to the relative amplitude of the normalized PVO index and its temporal tendency from 0 to 1 and then back to 0. The negative PVO events are binned into 40 intervals in a similar fashion reflecting an evolution from 0 to −1 then back to 0. Joining the composite evolution of the positive and negative PVO events forms the composite PVO event. Because of the smoothness of the raw daily PVO index, the temporal evolution of PVO events can be faithfully captured by

1 The slightly smaller correlation with the NAM indices above 20 hPa merely reflects the fact that isentropic analysis used in this study consists of data on and below the 650-K surface, which is roughly at the level of about 20 hPa, rather than suggesting that the PVO is more related to the NAM in lower stratosphere.
Fig. 2. Daily time series of the first EOF mode (or the PVO index) shown in Fig. 1. The red portions of the time series are indicative of the PVO events used in the composite analysis.
the composite PVO cycle. The average “residential” time between two adjacent bins is used to define the timeline of the composite PVO event. As shown in Fig. 5, the composite PVO event (the curve with solid dots), starting at day 0, reaches the strongest polar vortex phase at day 32 and the weakest phase at day 76, and ends at day 116. The notable difference between the composite PVO event and a pure sine function is indicative of a strong asymmetry between the two extreme phases of the composite PVO event. The time scale of 116 days is consistent with the power spectral density function of the daily PVO index that shows a significant peak at the time scale of about 107 days.2

The temporal and spatial evolution of the composite anomalies exhibits essentially the same characteristics as the regressed anomalies in terms of the propagation direction/speed and the temporal/spatial phase relations among different variables (Ren and Cai 2006). The main difference is that the composite anomalies capture the asymmetry between positive and negative PVO events. Also the regressed anomalies show a smoother spatial and temporal pattern but with a smaller amplitude than the composite anomalies.

b. Poleward and downward propagation in the stratosphere

Figure 6 shows the stratospheric anomalies of temperature, zonal wind, and the static stability parameter.

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2 The extra 9 days in the composite PVO event are mainly due to the difficulty in defining precisely the beginning/ending time of some PVO events that evolve slowly away from/toward the “zero amplitude” stage by our FORTRAN program.
averaged between 550 and 650 K as a function of the PV latitudes and the timeline of the composite PVO event. It is evident that the stratospheric anomalies tend to propagate poleward. The poleward propagating zonal wind anomalies tend to follow the temperature anomalies of the opposite sign whereas the poleward propagation of static stability anomalies is ahead of the temperature anomalies of the same sign. The Montgomery potential and PV anomalies exhibit a nearly in-phase relation with the temperature anomalies (not shown).

The poleward propagation speed varies greatly with latitudes. Anomalies advance poleward relatively very fast in the Tropics (<25°N) and within the polar cap (>60°N) and slowly from 25° to 60°N. The temporal lag of wind anomalies with respect to the temperature anomalies of the opposite sign whereas the poleward propagation of static stability anomalies is ahead of the temperature anomalies of the same sign. The Montgomery potential and PV anomalies exhibit a nearly in-phase relation with the temperature anomalies (not shown).

Another way to describe the simultaneous poleward and downward propagation of temperature anomalies shown in Fig. 7 is that at a given latitude, the arrival of temperature anomalies at a higher elevation leads that at a lower elevation. It follows that the static stability anomalies lead the temperature anomalies of the same sign. The temporal phase relation between the simultaneous poleward and downward propagating isentropic temperature and zonal wind anomalies can be explained from the PV dynamics. Cold (warm) temperature anomalies in the θ-PVLAT coordinate imply an upwelling (downwelling) of the isentropic surface due to diabatic cooling (heating) anomalies. As discussed in Hoskins et al. (1985), a cooling anomaly at upper levels implies a positive PV or negative Montgomery potential anomaly. Conversely, a heating anomaly at upper levels implies a negative PV or positive Montgomery potential anomaly. Indeed we found that the poleward and downward propagating temperature anomalies are negatively correlated with PV and positively correlated with Montgomery potential anomalies (not shown). For simplicity, let us use a simple plane wave to represent the simultaneous poleward and downward propagation.

Displayed in Fig. 7 are a series of vertical and time cross section diagrams of the stratospheric temperature, zonal wind, and static stability anomalies at different latitude bands from the equator to the Pole (from the bottom to top panels). The poleward propagation of the stratospheric anomalies is vividly apparent in Fig. 7. The temporal phase relations among temperature, static stability, and zonal wind anomalies shown in Fig. 6 remain unchanged throughout the stratosphere. Accompanying the poleward propagation there exists a simultaneous downward propagation of the stratospheric anomalies. The downward propagation takes place not only in the extratropics but also in the sub-tropics and Tropics. In general, the downward propagation of anomalies is faster by a few days in low latitudes than in high latitudes. In high latitudes, the downward propagation of isentropic temperature anomalies during the transition from the cold to warm phase is faster (about 10 days) than that in the transition from the warm to cold (about 20 days). The downward propagation of isentropic temperature anomalies in low latitudes is faster during the transition from the warm to cold phase.

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Fig. 7. A series of vertical-time cross section diagrams at different latitude bands from the equator to the Pole (bottom to top) of (a) the 15-day running mean composite isentropic temperature anomalies (shading; K) and (b) static stability parameter anomalies (shading; $10^{-5}$ s$^{-2}$). Each panel also has contours for the zonal wind anomalies (m s$^{-1}$); the dotted black, solid black, and white contours are for positive, zero, and negative values, respectively. The abscissa is the timeline of the composite PVO event (day). The anomalies are averaged over the PV-latitude band indicated by the text to the right of each row. The ordinate of each panel is the isentropic surface level ranging from 400 to 650 K.
of the signal in the stratospheric temperature anomalies:

$$T' = A \cos(ly - nz - ur),$$  \hspace{1cm} (2)

where $A$ denotes the perturbation amplitude, $l$ and $n$ are the meridional and vertical wavenumber of the signal, $u$ is the frequency of the signal propagation, and $y$ and $z$ are the distance along the meridian and vertical directions. For a simultaneous poleward and downward propagation, we have that $l > 0, n > 0$, and $u > 0$. The zonal wind anomaly $U'$ is related to the Montgomery potential anomaly $M'$ by the geostrophy, and $M'$ is positively correlated with $T'$ in the stratosphere as discussed above. Therefore, we have

$$fU' = -\frac{\partial M'}{\partial y} - \frac{\partial T'}{\partial y} = A l \sin(ly - nz - ur),$$  \hspace{1cm} (3)

where $f$ is the Coriolis parameter. According to (2) and (3), zonal wind anomalies would lag the Montgomery potential (or temperature) anomalies of the opposite sign by a quarter of period equal to $(2\pi/v)$ for both poleward and downward propagation.\(^3\)

Equation (3) essentially is the thermal wind balance between the zonal wind and temperature anomalies. The strengthening of the meridional temperature gradient follows the poleward propagating cold temperature anomalies. Ahead of the weakening of the meridional temperature gradient are the poleward propagating warm temperature anomalies. By the thermal wind balance, the westerly anomalies propagate with the strengthening of the temperature gradient and the easterly anomalies with the weakening of the temperature gradient. Therefore, the westerly anomalies would be behind the poleward propagating cold anomalies and the easterly anomalies behind the poleward propagating warm anomalies by a quarter of a period.

As indicated in Figs. 6 and 7, the poleward propagation speed varies greatly with latitudes. As a result, the lag of the zonal wind anomalies behind the temperature anomalies of the opposite sign also varies with latitude. Particularly, the poleward propagation is much faster in the Tropics and polar region, resulting in a shorter delay of zonal wind anomalies. In the midlatitudes, the lag of the zonal wind anomalies with respect to the temperature anomalies of the opposite sign is longer because the poleward propagation is slower.

c. Equatorward propagation in the lower troposphere in the extratropics

As will be shown shortly (Fig. 6 versus Fig. 8 and Figs. 10–14), anomalies of most variables have an opposite polarity between the stratosphere and lower troposphere (below 300 K). This seems to suggest that the temporal and spatial variation of the anomalies in the troposphere would exhibit different characteristics from their counterparts in the stratosphere. The low-latitude portion of an isentropic surface in the lower troposphere ($\theta < 300$ K) is below the ground (we refer to the intersection of an isentropic surface with the ground as the “edge” of the isentropic surface). The edges of these isentropic surfaces are highly nonzonal and vary drastically from day to day. As a result, the annual cycle of a variable on an isentropic surface in the lower troposphere is not well defined along the edges, attributing to a noisier anomaly field (in terms of spatial patterns) along the edges. The annual cycle in the $\theta$-PVLAT coordinate for the points near the edges of the isentropic surfaces is even more difficult to define because the meridional coordinate also varies with time. This leads to an even noisier anomaly field in the PVLAT coordinate along the edges of the lower-tropospheric isentropic surfaces. Although the data are noisier only in a very narrow band at the edge of each isentropic surface, there are several “noise stripes” when averaging anomalies vertically across several isentropic surfaces in the lower troposphere. This is because the edges of different isentropic surfaces are located at different latitudes. These noisy stripes make it difficult to visualize the temporal and spatial variation pattern in the lower troposphere. For this reason, unless specified otherwise, we only show the latitude–time cross section diagrams obtained by averaging along the regular latitude for the isentropic surfaces below 300 K because the results obtained by averaging along PV latitudes are similar but with more noticeable noise stripes.

Displayed in Fig. 8 are the composite temperature and zonal wind anomalies averaged between 270 and 290 K. In the extratropics, the tropospheric temperature anomalies lag the stratospheric temperature anomalies of the opposite sign (shadings in Fig. 6a versus Fig. 8a) whereas the tropospheric zonal wind anomalies are positively correlated with their stratospheric counterparts (contours in Fig. 6 versus Fig. 8b). In contrast to the stratosphere, the tropospheric temperature anomalies appear to propagate equatorward in the extratropics. The meridional propagation becomes less evident when the temperature anomalies reach the subtropics. However, the tropospheric zonal...
wind anomalies mainly exhibit a see-saw pattern between high latitudes and midlatitudes/subtropics. Other tropospheric anomalies (such as Montgomery potential and PV) also display a seesaw oscillating pattern between the midlatitudes and polar region with a weak propagation signal (not shown). The results obtained in the PVLAT coordinate are similar, namely, that only the tropospheric temperature anomalies exhibit a clear equatorward propagation signal in the extratropics.

5. Temporal and spatial variations of atmospheric thermal structure

As discussed in 4b, the simultaneous poleward and downward propagating temperature anomalies are responsible for a systematic change in the meridional temperature gradient and atmospheric static stability. By definition, on an isentropic surface \( \theta = \text{constant} \), we have

\[
\frac{T'}{T} = \frac{R \ p'}{C_p \ p},
\]

where the overbar denotes the mean value, and the primes denote the departure from the mean; \( R \) and \( C_p \) are the gas constant and specific heat of dry air at constant pressure, respectively. Therefore, temperature anomalies in the \( \theta \)-PVLAT coordinate reflect a local elevation change of the isentropic surface due to diabatic heating/cooling. It follows that the simultaneous poleward and downward propagating isentropic temperature anomalies would have to result in a simultaneous poleward and downward propagation of the change in the isentropic surface slope, as conjectured in Cai and Ren (2006).

To reveal the existence of such a coherent change in isentropic surface slope, we have calculated the meridional slope of isentropic surfaces (denoted as “\( \theta \_\text{slope} \)”) using the daily data according to

\[
\theta \_\text{slope} = -\frac{\partial \theta}{\partial y} \frac{\partial y}{\partial z}.
\]

The same procedures described in section 2 are used to obtain the daily 2-D \( \theta \_\text{slope} \) field in both \( \theta \)-PVLAT and regular \( \theta \)-latitude coordinates. The daily anomaly field of \( \theta \_\text{slope} \) is obtained by removing the annual cycle. A positive slope anomaly of an isentropic surface corresponds to the “steepening” case in which the isentropic surface becomes more steeply sloped toward north whereas a negative slope anomaly is indicative of the “leveling” of the isentropic surface.

It is very evident that there exists a simultaneous poleward and downward propagation of the slope anomalies from low to high latitudes throughout the stratosphere (Fig. 9a). The maximum slope variability is collocated with the maximum zonal wind variability (Fig. 10a versus Fig. 6a), at the latitude where the polar jet core is located (about 62°N). Comparisons of Fig. 9a with Fig. 7a and Fig. 10a with Fig. 6a also reveal that the slope anomalies of isentropic surfaces in the stratosphere in general lag the temperature anomalies of the opposite sign but are slightly ahead of zonal wind anomalies of the same sign. Another interesting feature is that the poleward propagation of the leveling anomalies is faster than the steepening anomalies of isentropic surfaces, consisting with the results shown in Figs. 6a and 7a. It is of importance to point out that both leveling and steepening anomalies of isentropic surfaces propagate slowest at the southern edge of the jet core (50°–60°N). To the north of the jet core (＞65°N), the leveling/steepening anomalies appear to take place nearly simultaneously with that at the southern edge of the jet core, resulting in the fastest expansion of anomalies into the polar circle. The similar situation is also very noticeable in temperature, zonal wind, and static stability anomalies (Fig. 6).
Similar to the relation between the tropospheric and stratospheric temperature anomalies, the isentropic slope anomalies in the troposphere (Fig. 10b) appear to be negatively (positively) correlated with their counterparts in the stratosphere in high and low latitudes (mid-latitudes). It is also evident that the tropospheric slope anomalies tend to propagate equatorward. Again, as the temperature anomalies, the timing of the equatorward propagating isentropic slope anomalies in the troposphere coincides with the arrival of the poleward propagating isentropic slope anomalies of the opposite sign over the polar stratosphere. In other word, the leveling (steepening) of tropospheric isentropic surfaces in high latitudes is synchronized with the steepening (leveling) of the isentropic slope in the polar stratosphere.

Shown in Fig. 11 are the $\theta$-PVLAT cross section diagrams of the composite isentropic slope and zonal wind anomalies at days 0, 30, 49, and 68 of the composite PVO event. In the stratosphere, the maximum centers of zonal wind and slope anomalies of the same sign coincide with one another. For the positive phase (day
30), the polar jet is the strongest (the maximum composite westerly wind anomaly is $10 \text{ m s}^{-1}$) with steeply sloped isentropic surfaces over the polar stratosphere whereas easterly wind anomalies with a less-sloped isentropic surface prevail from the Tropics to midlatitudes. The opposite situation is observed at the negative phase (day 68).

The poleward propagation signal in the stratosphere is very noticeable from these “instantaneous” cross-section diagrams, as indicated by the poleward shifting of the meridional see-saw pattern from the bottom to top panel. The relatively slower change at the jet core discussed earlier can be vividly seen from the vertical–latitude cross section diagram at the transition period between the two extreme phases. At day 49, several weeks before the extreme negative phase, the slope anomalies at the jet core are still positive despite the fact that the negative slope anomalies already appear at both sides. At the same time, the maximum center of westerly anomalies remains intact although it has been weakened significantly. Just a few days later, the easterly wind anomalies begin to prevail over the entire polar stratosphere accompanied with the collapsing of the sloped isentropic surfaces due to both enhanced static stability and reduced meridional temperature gradient.

In the extratropics, the upper-stratosphere and lower-tropospheric slope anomalies have opposite polarity at the two extreme phases but the same polarity at the transition periods ("day 0" and "day 49"). This implies that the reversing of the polarity of the slope anomalies in the lower troposphere lags that in the upper stratosphere, consistent with the results shown in

**Fig. 10.** The latitude–time cross section diagrams of the 15-day running mean composite anomalies of isentropic surface slope (units of $10^{-6}$) averaged between (a) 550 and 650 K and (b) 270 and 290 K. The gray shading is indicative of positive anomalies. The abscissa is the timeline of the composite PVO event (day) and the ordinate is PV latitude in (a) and regular latitude in (b). The thick solid black line in (b) marks the mean latitude at which the isentropic surface of 290 K intersects with the ground.

**Fig. 11.** The $\theta$-PVLAT cross section diagrams of the 15-day running mean composite slope (shading; units of $10^{-6}$) and zonal wind (contours; m s$^{-1}$) anomalies at days 0, 30, 49, and 68 (bottom to top) of the composite PVO event. The dotted black, solid black, and white contours are for positive, zero, and negative values, respectively. The abscissa is PVLAT and the ordinate of each panel is the isentropic surface level ranging from 270 to 650 K. The vertical scale between 270 and 300 K is enlarged slightly as marked by the solid black horizontal line. To highlight the latitudinal variation, the hemispheric mean value of the slope anomalies above 300 K has been removed.
Fig. 10. The polarity of zonal wind anomalies, on the other hand, remains unchanged from the stratosphere to troposphere at all of the four phases. This is another way of saying that the zonal wind and slope anomalies are negatively correlated in the extratropical troposphere.

6. Propagation of mass anomalies

To relate the mass circulation variability to the annular mode, we need to calculate the global mass circulation on daily basis based on

$$\frac{\partial M_a}{\partial t} + \nabla_x \cdot (V M_a) + \frac{\partial \theta M_a}{\partial \theta} = 0,$$  \hspace{2cm} (6)

where $M_a = -1/(g \partial p/\partial \theta)$ is the mass between isentropic surfaces, $\nabla_x \cdot (V M_a)$ is the horizontal divergence of the mass flux vector $V M_a$ along isentropic surface, $(\partial \theta M_a/\partial \theta)$ is the mass flux across isentropic surfaces due to the diabatic heating $\theta$. Unlike the time mean mass circulation which only involves mass streamfunction (e.g., Johnson 1989; Schneider 2006), the daily mass circulation also need to account for the local change of the mass between isentropic surfaces due to the unbalance between divergence of mass fluxes and diabatic heating. The standard NCEP–NCAR reanalysis resolution (both vertical and horizontal) seems to be somewhat too coarse to have an accurate representation of the daily mass circulation. Instead of diagnosing the mass circulation on daily basis, which is beyond the scope of this paper, we examine the temporal and spatial pattern of mass anomalies between isentropic surfaces to infer the mass circulation variability.

The mass field between two adjacent isentropic surfaces is calculated by

$$M_a(x, y, t) = (p_a - p_{a+\Delta\theta})/g,$$  \hspace{2cm} (7)

where $p_a$ is the pressure field along the isentropic surface $\theta$ and $\Delta\theta > 0$. Similarly, we have also calculated the mass field below the lowest isentropic surface (denoted as $M_{surf}$),

$$M_{surf}(x, y, t) = (p_{surf} - p_{a\text{, lowest}})/g,$$  \hspace{2cm} (8)

where $p_{surf}$ and $p_{a\text{, lowest}}$ are surface pressure and pressure field on the lowest isentropic surface (in the dataset used in this paper, the lowest isentropic surfaces varies from 270 K from high latitudes to 300 K in low latitudes in winter). Following the same procedures described in section 2, we obtain the daily 2D mass fields in both $\theta$-PVLAT and regular $\theta$-latitude coordinates. The daily mass anomaly fields are obtained by removing their annual cycles.

The results shown in Fig. 9b clearly indicate that the isentropic mass anomalies in the stratosphere propagate poleward and downward simultaneously. Comparison of Fig. 9b with Fig. 7a suggests that the mass anomalies lag the temperature anomalies of the same sign by few days. Therefore, the poleward propagating positive (negative) isentropic mass anomalies reflect mainly the diabatic dynamic heating (cooling) anomalies associated with an enhanced (weakened) meridional mass circulation.

In contrast to the stratosphere where the mass anomalies propagate poleward (Fig. 9b and Fig. 12a), the near surface mass anomalies ($M_{surf}$; Fig. 12c) appear to propagate equatorward from the pole to midlatitudes whereas the upper-tropospheric mass anomalies (270–300 K) exhibit a sea-saw pattern between high and midlatitudes (Fig. 12b). In the extratropics, the near-surface mass anomalies are negatively correlated with the upper tropospheric mass anomalies, but positively correlated with the stratospheric mass anomalies. The total column air mass anomalies above the ground (i.e., the surface pressure anomalies divided by the gravity parameter; Fig. 12d) are nearly in phase with the mass anomalies in the stratosphere and near the surface, but are negatively correlated with the upper-tropospheric mass anomalies. Another important feature is that the mass anomalies in high latitudes are much larger than other latitudes throughout the entire column of the atmosphere.

These seemingly complicated features can be qualitatively explained by the global mass circulation variability. Because of the earth’s rotation, a Lagrangian-type single overturning circulation cell from the equator to the Pole cannot prevail. Outside the Tropics, the time mean meridional mass circulation is largely carried out by the geostrophic flow rather than the ageostrophic flow although the former has no contribution to the zonal mean meridional wind (e.g., Johnson 1989, and the references therein). The existence of the apparent single overturning time mean mass circulation from the equator to the Pole in the isentropic coordinate merely reflects the long-time mean balance between the thermodynamic heating and the poleward heat transport (or dynamic heating) by the Hadley circulation in the Tropics and baroclinic eddies in the extratropics. However, at any given instance, the eddy-driven meridional mass circulation would not be able to cover the entire hemisphere because the meridional span of the dynamic heating itself is limited by the meridional scales of baroclinic eddies. As time progresses, these individual circulation systems would collectively cover the entire hemisphere, as reflected in the time mean mass circulation.
Next let us first discuss the warm air branch in the stratosphere. The dynamic heating anomaly associated with an enhanced warm air branch in the stratosphere implies a lowering of the tropopause and therefore \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) > 0 \] in the stratosphere. Meantime, the heating anomaly would cause an enhanced poleward warm air advancement into higher latitudes, resulting in a pair of negative and positive convergence anomalies of mass fluxes, namely, \[ -\nabla_o \cdot (\mathbf{V} M_o) > 0 \] for partially balancing \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) > 0 \] locally and \[ -\nabla_o \cdot (\mathbf{V} M_o) > 0 \] on the farther poleward side. The local imbalance between \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) > 0 \] and \[ -\nabla_o \cdot (\mathbf{V} M_o) < 0 \] causes a positive mass anomaly locally whereas the pattern \[ -\nabla_o \cdot (\mathbf{V} M_o) > 0 \] would contribute to a new dynamical heating anomaly in the latitude band poleward. This explains the systematic poleward propagating mass anomalies (Fig. 12a) that are positively correlated with heating anomalies (Fig. 6a). Such a pattern goes on continuously poleward till it approaches the Pole where, in addition to \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) > 0 \], \[ -\nabla_o \cdot (\mathbf{V} M_o) \] is also positive because the warm air branch “ends” at the Pole. The same polarity of \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) \] and \[ -\nabla_o \cdot (\mathbf{V} M_o) \] is responsible for the largest mass anomaly over the polar stratosphere (Fig. 12a). The positive mass anomalies in the polar stratosphere mean the falling of the isentropic surfaces with a lowered tropopause, corresponding to the leveling of isentropic surfaces in the extratropical stratosphere (Fig. 10a or the top panel of Fig. 11). The negative mass anomalies over the polar stratosphere, on the other hand, are associated with a weaker warm air branch. The weaker dynamic heating (or negative heating anomalies) implies a rising of the tropopause because \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) < 0 \] and a reduction of the warm air supply over the polar stratosphere because \[ -\nabla_o \cdot (\mathbf{V} M_o) < 0 \]. Therefore, as the warm phase, these two terms reinforce each other instead of a partial cancellation in the polar stratosphere, resulting in the largest negative mass anomaly in the cold phase (Fig. 12a).

A stronger (weaker) warm air branch over the polar stratosphere coincides with a stronger (weaker) cold air branch in the lower troposphere, as indicated by the negatively correlated dynamic heating anomaly between the stratosphere and the lower troposphere (Fig. 6a versus Fig. 8a). A dynamic cooling anomaly associated with a stronger cold air branch in the lower troposphere implies a rising of isentropic surfaces near the ground (Fig. 10b), contributing to a positive mass anomaly below 270 K. In terms of (6), we have \[ -\left( \frac{\partial \theta M_g}{\partial \theta} \right) > 0 \] for a cooling anomaly in the lower troposphere. Meantime, the polar area is the beginning point of the cold air branch. The cooling anomaly would imply an enhanced mass flux (of cold air) away...
from the polar area toward lower latitudes, namely,
\[- \nabla_a \cdot (VM_a) < 0, \text{ which counters } \left(- \frac{\partial M_a}{\partial u}\right) > 0.\]
The imbalance between \(\left(- \frac{\partial M_a}{\partial u}\right) > 0\) and
\[- \nabla_a \cdot (VM_a) < 0\] results in a positive mass anomaly
in the lower troposphere (Fig. 12c). The opposite situation is observed for a weak mass circulation.

Such a change in the atmospheric thermostructure in response to the heating anomalies helps to explain why
the mass anomalies in the stratosphere and near the ground are positively correlated. The positively correlated change in\(\left(- \frac{\partial M_a}{\partial u}\right) > 0\) between the stratosphere and lower troposphere due to a nearly simultaneous lowering of the tropopause and rising of the lower tropospheric isentropic surfaces implies that the mass anomalies in the stratosphere and lower troposphere are sandwiched by a mass anomaly of the opposite sign in the upper troposphere. Because of the opposite propagation between the stratospheric and near-surface mass anomalies, the mass anomalies in between (Fig. 12b) would show little propagation, but mainly a sea-saw pattern between high and midlatitudes.

We here wish to add that there is an asymmetry between low and high latitudes from the thermodynamics point of view. On top of the heat gain/loss due to the irreversible nonlinear dynamic mixing associated with the mass circulation, each of the two branches would also gain/lose heat via thermodynamic processes. The warm air branch aloft would lose its steam mainly via radiative cooling, which is a much slower thermodynamic process. Because of vigorous heat exchange with the surface below, the cold air branch would lose its strength and “identity” very quickly before it reaches its final destination. This is consistent with the fact that the dynamic heating anomalies in the upper-level atmosphere in high latitudes due to the variability of the warm air branch is much larger than the lower-level heating anomalies in low latitudes associated with the cold branch variability. This also helps to explain why the equatorward propagation of heating anomalies in the lower atmosphere is evident mainly in the extratropics.

7. Relation between the mass circulation and polar surface pressure anomaly

We next apply the same reasoning for explaining the surface pressure anomalies associated with the annular mode from the perspective of the mass circulation variability. Figure 13a shows that the mass anomalies have the same polarity throughout the column above the 350-K surface (about 180 hPa) other than a short temporal phase difference due to the downward propagation. Again as seen before, the upper-tropospheric and lower-stratospheric mass anomalies (between 270 and 350 K) in high latitudes are negatively correlated with mass anomalies both above 350 K and below 270 K (or about 800 hPa). The total column mass anomalies (or the surface pressure anomalies) lag the stratospheric mass anomalies by a few days (Fig. 13b versus Fig. 13d).
The variation of the total mass anomalies above 350 K (thick curve in Fig. 13b) reflects a rising of the 350-K surface by about 9 hPa (−90 kg m$^{-2}$ in terms of mass) during the positive phase of the PVO and a falling by about 11 hPa (110 kg m$^{-2}$) in the negative phase.

Although the mass anomaly near the surface (below 270 K in Fig. 13a) is also positively correlated with the surface pressure anomaly, its amplitude (about 40 kg m$^{-2}$) is less than half of that above the 350-K surface and it lags the surface pressure anomaly by several days. It is also of importance to note that the mass anomaly above the 550-K surface (about 35 hPa) is already nearly as large as that below 270 K. In terms of percentage of the mean mass, the mass anomaly in the stratosphere is about 6%–9% of the mean mass there (30/350 for above 550 K and 100/180 for above 350 K), whereas the mass anomaly near the surface only accounts for 2% of the mean mass. Therefore, the information of the temporal phase relation and the amplitude of the mass anomaly seem to suggest that the surface pressure anomaly in high latitudes is largely determined by the convergence anomaly of the warm air branch.

As discussed in section 6, the convergence anomalies of the upper and lower branches over the polar region would be negatively correlated. We could estimate the net convergence anomaly associated with each branch of the mass circulation by finding a layer in the middle at which the horizontal mass flux is zero. This level should be somewhere in the middle of 270–350 K where the mass anomalies are negatively correlated with mass anomalies above and below. In other word, the mass above this nondivergent level involves mass exchange with the stratosphere (above 350) and below with the lower troposphere (below 270 K). The total airmass anomaly above (below) this nondivergent layer reflects the convergence anomaly of the warm (cold) air branch. The thick and thin solid curves in Fig. 13c are the total mass anomalies above 300-K (about 350 hPa or the tropopause level) and 290-K (about 500 hPa) surfaces, respectively. Both curves meet the “necessary” conditions for contributing to the surface pressure anomaly: their amplitudes are larger, and they lead and are positively correlated with the surface pressure anomaly. The elevation change of the 300-K surface, which approximately corresponds to tropopause level in high latitudes, is about 14 hPa between the two extreme phases of a NAM event, very close to the 10-hPa change estimated by Ambaum and Hoskins (2002).

Note that the 290-K surface is well below the tropopause where the mass exchange between the troposphere and stratosphere is the strongest. Therefore, the mass anomalies above the 290-K surface would be mainly due to the net convergence anomalies of the warm air branch in the stratosphere. According to the thin curve in Fig. 13c, the extra mass brought by the enhanced warm air branch peaks on day 66, contributing to about 60 kg m$^{-2}$ (or 6-hPa pressure) of mass over the polar cap. The difference between Fig. 13d and Fig. 13c represents the convergence anomaly of the cold air branch, which is negatively correlated with the surface pressure anomaly. The (cold air) mass brought out of the polar cap by the enhanced cold air branch peaks few days later, moving about 35 kg m$^{-2}$ mass out of the polar cap. The maximum positive surface pressure anomaly over the polar cap (about 2.5 hPa or 25 kg m$^{-2}$) occurs when the polar vortex is weakest. Afterward, the surface pressure anomaly gradually becomes smaller because of the slowdown of the arrival of the warm air aloft and the continuation of the moving out of the cold air mass.

The discussion above provides a physical explanation for why the NAM index in the upper levels is positively correlated with the AO index. During the stronger polar vortex phase, the global meridional mass circulation is weaker. This implies a relatively reduction of the supply of the air mass above the polar cap from the warm air branch (or a divergence anomaly of the warm air branch). Meantime, a “slowdown” of the equatorward advancement of cold air mass implies a convergent mass flux anomaly of the cold air branch in the troposphere [in this sense, the cold air is imprisoned within the polar circle in the high zonal index phase as first suggested by Namias (1950)]. Because the reduction of the stratospheric warm air due to the lack of supply is larger, the total surface column air mass has a negative anomaly over the polar circle. The arrival of the warm air mass in the stratosphere implies an extra amount of air mass above the polar cap surface during the weak polar vortex phase. This leads to a rising of the surface pressure over the polar region till the surface cold air moves out into lower latitudes.

8. Coupling of stratospheric and tropospheric anomalies

The downward propagation of geopotential height and zonal wind anomalies from the stratosphere to the troposphere in the polar region has been recognized as the evidence suggestive of the stratosphere influence onto the tropospheric circulation (e.g., Haynes 2005, and the references therein). As shown in Figs. 7 and 9, the downward propagation of stratospheric anomalies takes place not only in high latitudes but also in other latitudes. However, the continuation of the downward propagation into the troposphere is not observed in all
variables and at all latitudes. Particularly, the mass anomalies do not propagate continuously from the stratosphere to the surface as shown in Fig. 13a. In this section, we wish to explain the seemingly complicated phase relation between stratospheric and tropospheric anomalies at different latitudes and among different variables from the dynamic balance between PV and heating anomalies associated with the mass circulation.

As discussed in the previous section, from the global mass circulation point of view, it is natural to expect a strong coupling between the stratosphere and troposphere in high latitudes where the poleward warm air branch of the global mass circulation is connected to the equatorward cold air branch. As indicated in Fig. 6a and Fig. 8a (also see Fig. 10a versus Fig. 10b, and Fig. 12a versus Fig. 12c), the arrival of the stratospheric temperature anomalies over the polar region coincides with the beginning of the compensating equatorward advancement of tropospheric temperature anomalies of the opposite sign. Therefore, the temperature anomalies of the same sign do not “propagate” downward continuously into the troposphere in high latitudes (Fig. 14a).

According to Hoskins et al. (1985), isentropic temperature anomalies in upper levels correspond to PV anomalies of the opposite sign (Fig. 14b). The relatively shallowness of the isentropic temperature anomalies in the lower troposphere implies that the polarity of the PV anomalies near the surface is mainly determined by the static stability anomalies rather than the vorticity anomalies. Because near the surface, the static stability anomaly has to be negatively correlated with isentropic temperature anomaly (Fig. 14c versus Fig. 14a), the lower-tropospheric PV anomalies, which are positively correlated with the static stability, are also negatively correlated with isentropic temperature anomalies as is the case in the upper atmosphere. It follows that the negative correlation of the stratospheric and tropospheric temperature anomalies (Fig. 14a) immediately implies that PV anomalies in the stratosphere and troposphere are negatively correlated and the downward propagation of PV anomalies in the stratosphere would not extend to the troposphere (Fig. 14b).

In upper levels, the positive (negative) static stability anomalies are due to a warming (cooling) above and/or a cooling (warming) below. The downward propagation of temperature in the stratosphere implies a downward propagation in statistic stability anomaly, which leads the temperature anomaly of the same sign (Fig. 14c versus Fig. 14a). Near the surface, the static stability anomalies are negatively correlated with temperature anomalies. Because the timing of the tropospheric temperature anomalies in the polar region coincides with the arrival of the stratospheric temperature anomalies of the opposite sign, the static stability anomalies appear to continuously propagate into the troposphere from the stratosphere, but at a much slower speed (Fig. 14c).

As discussed in Hoskins et al. (1985), upper-level isentropic temperature anomalies would produce a vor-
ticity anomaly of the opposite sign whereas a low-level isentropic temperature anomaly would produce a vorticity anomaly of the same sign. It follows that the situation shown in Fig. 14a would immediately imply an apparent continuous downward propagation of both Montgomery potential (Fig. 14d) and zonal wind (Fig. 14e) anomalies into the troposphere from the stratosphere. Therefore, the continuation of the downward propagation into the troposphere is determined by the “local” dynamic response to the heating/cooling anomalies and should not be regarded as evidence suggesting the stratospheric origin of the tropospheric anomalies. Also as indicated in Fig. 13a, the negatively correlated heating anomalies between the stratosphere and troposphere are responsible for the positively correlated mass anomalies in the stratosphere and lower troposphere sandwiched by mass anomalies of the opposite sign in between. Based on such vertical profile of mass anomalies plus the condition that the surface pressure anomaly is positively correlated with the stratospheric mass anomalies, we would expect a pair of positively correlated height anomaly centers in the stratosphere and in the lower troposphere sandwiched by a minimum anomaly in between according to the hydrostatic balance. Therefore, the apparent equivalent barotropic structure of the annular mode results mainly from dynamic response to the negatively correlated heating anomalies above and below, rather than the barotropic nature of the underlying dynamics. This is consistent with the observations showing that there is a minimum in both Montgomery potential (Fig. 14d) and zonal wind anomalies (Fig. 14e) between the tropopause and surface. This seems to explain why the correlation between the PVO index and NAM indices has a minimum in the middle of the troposphere rather than at the surface (Fig. 4).

From the global mass circulation point of view, heating anomalies in low latitudes would have to be a mirror image of their counterparts in high latitudes because the Tropics/subtropics is the region where the remnant cold air branch of the global mass circulation is connected to the poleward warm air branch. This helps to explain why the isentropic temperature anomalies in low latitudes are negatively correlated with those in the polar region (Figs. 7 and 9). We can apply the same reasoning to explain the “continuation” (e.g., zonal wind anomalies) or “interruption” (e.g., PV anomalies) of the downward propagation into the troposphere from the stratosphere in low latitudes (not shown).

9. Summary and concluding remarks

This paper reports a study on the spatial and temporal evolution of the atmospheric circulation anomalies viewed in a semi-Lagrangian $\theta$-PVLAT coordinate generated by following contours of the daily potential vorticity (PV) field on isentropic ($\theta$) surfaces using the NCEP–NCAR reanalysis II dataset from 1979 to 2003. Because of conservation properties of both potential vorticity and potential temperature, the advective tendencies of PV are naturally absent in the $\theta$-PVLAT coordinate. As a result, the temporal evolution of daily PV field in the $\theta$-PVLAT is dominant with time scales much longer than the typical weather time scale. More importantly, because of the semi-Lagrangian property, the grids in the $\theta$-PVLAT coordinate serve as natural boundaries separating air masses of different properties.

The leading EOF mode of the daily Northern Hemisphere PV anomalies in the $\theta$-PVLAT coordinate explains about 69% of the total variance of daily PV anomalies, representing an oscillation between a stronger and weaker polar vortex [referred as the polar vortex oscillation (PVO) mode]. The daily time series of the PVO mode (referred as the PVO index) has little sign of synoptic-scale variability. It has large amplitude primarily in winters and exhibits only 1–2 PVO events within a winter season. The maximum correlation is about 0.91 when the NAM index in the upper stratosphere leads the PVO index by about 11 days. The power spectral density function of the daily PVO index exhibits five distinct peaks. The 107-day peak corresponds to the dominant intraseasonal NAM variability in cold season, the main subject of this paper. The 5-month peak reflects the fact that the NH PVO is active only from late fall to early spring each year and is relatively quiescent in the rest of the year. The periods of 8.7, 16.9, and 33.8 months, which approximately follow a double-periodicity sequence, correspond to the three-peak extratropical QBO signal first reported in Tung and Yang (1994).

Based on the daily PVO index, the composite method is used to depict the temporal and spatial evolution of the circulation anomalies at the 107-day scale. It is found that temperature anomalies in the stratosphere propagate poleward and downward simultaneously. The poleward propagation of the stratospheric anomalies is faster in the Tropics and in the polar region, but is slower from the subtropics to the extratropics. On average, it takes about 40 days for warm anomalies and 70 days for cold anomalies in the stratosphere to travel from the equator to the Pole. The speed of the downward propagation also strongly depends on the anomaly polarity. In high latitudes, the downward propagation during the transition from cold to warm anomalies is faster (about 10 days) than that from warm to cold anomalies (about 20 days). The
anomalies in low latitudes are nearly out-of-phase with the high-latitude anomalies, but their downward propagation is faster than that in high latitudes by a few days.

The simultaneous poleward and downward propagating temperature anomalies coincide with mass anomalies between isentropic surfaces of the same sign. It reflects the temporal and spatial variations of dynamic heating/cooling anomalies associated with the global mass circulation. The upper-level branch of the mass circulation is associated with the poleward advancement of the warm air mass whereas the low-level branch with the equatorward marching of the cold air mass. It follows that the poleward propagation of warm (cold) temperature anomalies reflects a stronger and faster (weaker and slower) meridional mass circulation.

The dynamic heating/cooling anomalies associated with the global mass circulation inevitably lead to a temporal and spatial variation of the atmospheric thermosstructure. In the stratosphere, the simultaneous poleward and downward propagation implies that temperature anomalies at a higher elevation lead those at a lower elevation at all latitudes, accounting for a simultaneous poleward and downward propagation of the static stability anomalies that is ahead of the propagating temperature anomalies of the same sign. The poleward propagation of warm (cold) temperature anomalies implies a weakening (strengthening) of meridional temperature gradient. The negatively correlated change in meridional temperature gradient and static stability implies a coherent change in the isentropic surface slope. We have shown that accompanied with the simultaneous poleward and downward propagation of warm (cold) temperature anomalies are a succession of leveling (steepening) of the isentropic surface slope from the equator to Pole and from the stratosphere to the tropopause.

In the extratropical troposphere, temperature anomalies propagate equatorward. The beginning of the equatorward propagation of cold (warm) tropospheric temperature anomalies coincides with the arrival of the poleward and downward propagating warm (cold) temperature anomalies in the polar stratosphere and therefore corresponds to a stronger (weaker) meridional mass circulation. Accompanied with the equatorward propagating cold (warm) tropospheric temperature anomalies is the strengthening of the meridional temperature gradient and the steepening (leveling) of the isentropic surfaces in the lower troposphere. Unlike the stratosphere, the tropospheric temperature anomalies lose their “identity” by heat exchange with the surface below at a time scale much shorter than radiative process. As a result, the equatorward propagation of tropospheric temperature anomalies becomes less evident from mid- to low latitudes.

The warm air branch of the global mass circulation directly connects the stratospheric anomalies between the Tropics and high latitudes. The union of the warm and cold air branches in high latitudes links the stratospheric and tropospheric anomalies in the extratropics. The nearly simultaneous strengthening/weakening of the warm air branch in the stratosphere and cold air branch in the lower troposphere over the polar region results in the asymmetry of heating anomalies associated with the mass circulation between the stratosphere and lower troposphere. This helps to explain the complicated relations between the stratospheric and tropospheric anomalies in high latitudes, as summarized in Table 1. Particularly, the apparent continuation of the downward propagation of zonal wind and Montgomery potential anomalies in the high latitudes is due to the local dynamic response to the diabatic heating/cooling anomalies associated with the mass circulation, rather than suggesting the stratospheric origin of the anomalies. Or the apparent equivalent barotropic structure of the annular mode mainly results from the dynamic response to the heating anomalies that has an opposite polarity between the stratosphere and lower troposphere, rather than an indicative of the barotropic nature of the underlying dynamics.

The diagnostics presented in this paper portrays a physical nature of the annular mode variability by relating it to the global mass circulation variability. Because of the Earth’s rotation, the westerly flow associated with the vertically sloped isentropic surfaces that prevails from the troposphere to the stratosphere and from the subtropics to the Pole is a physical barrier for the direct exchange of warm and cold air masses. In the positive phase of the annular mode, the barrier of the westerly flow is the strongest, corresponding to the case of a weaker meridional mass circulation in the extratropics. As a result of the relatively weaker warm (cold) air advancement into higher (lower) latitudes, the isentropic surfaces in the stratosphere (lower troposphere) are steeply (gently) sloped. Because the surface cold air mass is effectively “imprisoned” within the polar region (Namias 1950), there exist prevailing warm surface temperature anomalies in the extratropics. Meantime, the weaker mass circulation due to the stronger barrier of the westerly jet implies a relative reduction of supply of the air mass above the polar cap. This explains the negative surface pressure anomalies over the polar cap when the upper level vortex is stronger. As the barrier of the westerly flow continues to build up, the dynamic
mixing due to the development of baroclinic waves causes the leveling (steepening) of the isentropic surfaces in the stratosphere (lower troposphere). The more gently sloped isentropic surface in the extratropical stratosphere corresponds to the negative phase of the annular mode in which the meridional temperature gradient is reduced and the polar jet is weakened. The equatorward advancement of the surface cold air results in frequent cold air outbreak episodes and overwhelming cold temperature anomalies in the extratropics. Meantime, the surface pressure anomalies become positive as more warm air arrives at the polar cap in the stratosphere till the cold air in the lower troposphere moves out equatorward. This explains why the surface pressure anomalies over the polar cap are positive when the upper-level westerly jet is weaker.

The global mass circulation paradigm for the annular mode appears to be analogous to the frontal circulation system for weather. The pioneer studies by Reed and Sanders (1953), Reed (1955), and Sanders (1955) established the association of the tropopause foldings with upper-level (troposphere) frontogenesis and surface fronts (also Palmén and Newton 1969; Newton and Holopainen 1990; Shapiro and Crónás 1999; particularly, see Fig. 10.5 in Shapiro and Keyser 1990). A frontal zone (or baroclinic zone) is largely parallel to the isentropic surfaces and always tilts toward cold air. Therefore, above the surface cold high pressure center and behind a surface cold front is the zone of upper-level frontogenesis where the tropopause folds or the stratosphere descends (Palmén and Newton 1969). Obviously, the (tropospheric) frontogenesis case would correspond to the strong advancement of the surface cold air into the warmer sector. Farther up above the tropopause, the warm air branch in the stratosphere would also be strong, resulting in a stronger convergence of the warm air mass above and the extra warm air in the stratosphere naturally means the elevation drop of the tropopause by more weight and by the dynamic heating that warms the air near the tropopause. Underneath the upper-level frontal zone where the tropopause folds is the surface high pressure center where the cold air mass flux diverges. The simultaneous relation between the stratospheric warm air and its convergence, the surface cold air and its divergence, the falling of the tropopause above, and the rising of the surface pressure below for the weather system is very much similar to that for the annular mode except the former has a much larger ampli-

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<td>Positive due to both (1) and (2)</td>
<td>Positive; positive due to (1) but negative due to (2)</td>
<td>Mass anomaly in the upper troposphere is negative due to heating above and cooling below</td>
</tr>
<tr>
<td>Surface pressure</td>
<td>(2) brings about 60 kg m⁻² warm air over the polar cap</td>
<td>(2) moves 35 kg m⁻² cold air mass out of the polar cap</td>
<td>Positive (+2.5 hPa)</td>
</tr>
<tr>
<td>Temperature</td>
<td>Warmer</td>
<td>Colder</td>
<td>Arrival (beginning) of a stronger warm (cold) branch</td>
</tr>
<tr>
<td>Static stability⁷</td>
<td>Positive</td>
<td>Positive</td>
<td>Heating above and cooling below</td>
</tr>
<tr>
<td>Temperature</td>
<td>Negative</td>
<td>Positive</td>
<td>Arrival (beginning) of a stronger warm (cold) branch</td>
</tr>
<tr>
<td>gradient</td>
<td>Negative</td>
<td>Positive</td>
<td>Near the surface, PV and static stability anomalies are positively correlated¹</td>
</tr>
<tr>
<td>PV</td>
<td>Negative</td>
<td>Positive</td>
<td>Heating above and cooling below¹</td>
</tr>
<tr>
<td>Vorticity⁷</td>
<td>Negative</td>
<td>Negative</td>
<td>Follow the vorticity anomaly</td>
</tr>
<tr>
<td>Zonal wind⁷</td>
<td>Easterly</td>
<td>Easterly</td>
<td>“Tropopause folding” is a severe way to have stratospheric isentropic surfaces lowered</td>
</tr>
<tr>
<td>Height³</td>
<td>Positive</td>
<td>Positive</td>
<td>Follow the vorticity anomaly</td>
</tr>
<tr>
<td>Weather</td>
<td>More tropopause folding</td>
<td>Cold air outbreaks</td>
<td></td>
</tr>
</tbody>
</table>

* Mass anomaly between isentropic surface for interior or between the ground and the 270-K surface for the lower troposphere.
† Apparent continuation of the downward propagation into the troposphere.
 Rohing the dynamic constraints between PV and heating anomalies discussed in Hoskins et al. (1985).
tude with a much shorter scale in time and space. Therefore, the annular mode variability can be regarded as a low-frequency component of the global mass circulation that reflects a collection of the mass circulation systems carried out by individual events during a time span of weeks/months and covers the entire hemisphere.

The results presented here qualitatively relate the temporal and spatial evolution of heating anomalies and the resultant changes in the atmospheric thermodynamic heating to the annular mode variability. The changes in the momentum field are explained mainly from the dynamic balance between PV and heating anomalies. A more quantitative understanding about the amplitude, timing, and temporal and spatial scales of individual annular mode events requires a comprehensive diagnostics of the instantaneous two-way coupling between heating and momentum fields. In the θ-PV/LAT coordinate view, field structure alternation by the dynamic/thermodynamic heating inevitably changes the dynamic form drag applied on the isentropic surfaces (Andrews 1983) and PV surfaces (Stan and Randall 2007) or the convergence of the Eliassen–Palm (E–P) fluxes. The change in the momentum field leads to a change in the dynamic (through mixing/eddy-driven mass circulation) and thermodynamic (via radiation/exchange with the surface) heating. The strength of this instantaneous and endless two-way coupling between motion and its induced change in heat ultimately determines the amplitude and time scales of the annular mode variability. The results presented here echo the optimistic view on the climate prediction in winter seasons that has been put forward since the discovery of the downward propagation signal of the annular mode (e.g., Baldwin et al. 2003) because at the time scales of 40–70 days, the signal over the polar stratosphere is intimately connected to the Tropics via the warm air branch of the global mass circulation, providing that there are more and accurate observations above the tropopause in low/midlatitudes to be included in the model initial state and the global models have an accurate representation in the stratosphere.

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REFERENCES


