

# Atmospheric Processes Governing the Northern Hemisphere Annular Mode/ North Atlantic Oscillation

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The North Atlantic Oscillation, referred to herein as the Northern Hemisphere annular mode (NAM), owes its existence entirely to atmospheric processes. In this chapter, we review the structure of the NAM in the atmospheric general circulation, discuss opposing perspectives regarding its physical identity, examine tropospheric processes thought to give rise to NAM-like variability, and review the role of the stratosphere in driving variability in the NAM. The NAM is characterized by a deep, nearly barotropic structure, with zonal wind perturbations of opposing sign along  $\sim 55$  and  $\sim 35^\circ$  latitude. It has a pronounced zonally symmetric component, but exhibits largest variance in the North Atlantic sector. During the Northern Hemisphere (NH) winter, the NAM is strongly coupled to the circulation of the NH stratosphere. The NAM also affects tropical regions, where it perturbs the temperature and wind fields of both the tropical troposphere and stratosphere. The structure of the NAM is remarkably similar to the structure of the leading mode of variability in the Southern Hemisphere circulation. The processes that give rise to annular variability are discussed. In the troposphere, the NAM fluctuates on timescales of  $\sim 10$  days and is associated with anomalous fluxes of zonal momentum of baroclinic waves across  $\sim 45^\circ$  N. It is argued that the tropospheric component of the NAM exhibits largest variance in the Atlantic sector where the relatively weak thermally driven subtropical flow and the relatively warm lower boundary conditions at subpolar latitudes permit marked meridional excursions by baroclinic waves. In the stratosphere, fluctuations in the NAM evolve on timescales of several weeks. Evidence is presented that long-lived anomalies in the stratospheric NAM frequently precede similarly persistent anomalies in the tropospheric NAM. It is argued that variability in the lower stratospheric polar vortex yields a useful level of predictive skill for NH wintertime weather on both intraseasonal and seasonal timescales. The possible dynamics of these linkages are outlined. The recasting of the North Atlantic Oscillation as an expression of an annular mode has generated a debate over the physical identity of the mode in question. This debate attests to the absence of a unique theory for the existence of annular modes in the first place. Our current understanding of the fundamental processes to which the NAM owes its existence is discussed.

## 1. INTRODUCTION

The North Atlantic Oscillation (NAO) owes its existence, not to coupled ocean-atmosphere interactions, but to dynamics intrinsic to the extratropical troposphere. Numerical simulations of the climate system have repeatedly demonstrated that atmospheric processes alone are sufficient to generate an NAO-like pattern of variability [Hurrell *et al.*, this volume]. Results of both theoretical and modeling studies suggest that midlatitude sea-surface temperature anomalies have only a weak impact on the NAO, at least on month-to-month and year-to-year timescales [Kushnir *et al.*, 2002; Czaja *et al.*, this volume].

In this chapter, we review ongoing research efforts aimed at improving our understanding of the atmospheric processes thought to give rise to NAO-like variability. In Section 2, we document the structure of the NAO in the atmospheric circulation. In Section 3, we review the debate over the physical interpretation of the NAO. Section 4 explores the tropospheric dynamics of the NAO in greater detail, with emphasis on two current research questions: 1) what tropospheric processes give rise to NAO-like variability? and 2) why is variability in the NAO largest over the North Atlantic sector? In Section 5 we outline our current understanding of troposphere/stratosphere coupling in the context of the NAO. Section 6 offers a summary of the chapter and discusses priorities for future research.

## 2. THE STRUCTURE OF THE NAO IN THE ATMOSPHERIC CIRCULATION

Since its discovery, the NAO has been widely viewed as a zonally asymmetric pattern restricted primarily to the North Atlantic sector [see Stephenson *et al.*, this volume, for a history of NAO research]. Its structure has been defined on the basis of one point correlation maps [Wallace and Gutzler, 1981] or rotated empirical orthogonal functions [Barnston and Livezey, 1987]. Its temporal variability has been characterized by differences in sea level pressure between stations located near Iceland and the Azores/Portugal [Rogers, 1984; Hurrell, 1995a; Hurrell and van Loon, 1997; Jones *et al.*, this volume]. Not surprisingly, variability in the NAO has often been interpreted as a reflection of coupled ocean-atmosphere processes [e.g., Grotzner *et al.*, 1998; Rodwell *et al.* 1999].

Recently, it has been proposed that the dynamics of the NAO are analogous to those that drive the leading mode of variability in the Southern Hemisphere (SH) extratropical circulation, which is characterized by a zonally-symmetric seesaw in geopotential height between the polar cap and the surrounding zonal ring along  $\sim 45^{\circ}\text{S}$  [Rogers and van Loon,

1982; Szeredi and Karoly, 1987; Yoden *et al.*, 1987; Kidson, 1988a, b; Karoly, 1990; Hartmann and Lo, 1998; Gong and Wang, 1999] (Note that throughout this chapter, the term “mode” is used to describe a dominant pattern of variability. It does not necessarily refer to the normal modes of a physical system). The analogy between the leading mode of variability in the SH and the NAO suggests that the NAO reflects internal atmospheric dynamics that transcend the striking differences between the land-sea geometry of the two hemispheres. It has hence been suggested that the structure of the NAO is best captured, not by station based indices located over the North Atlantic, but by the leading empirical orthogonal function (EOF) of the Northern Hemisphere (NH) sea level pressure (SLP) anomaly field [Thompson and Wallace, 1998; 2000]. (The leading EOF is the state vector that explains the largest fraction of the total variance in the data. It is the eigenvector corresponding to the largest eigenvalue of the covariance matrix. Anomalies are defined as departures from climatology.) In contrast to the structure of the NAO that emerges from one-point correlation maps, the leading EOF of NH SLP has a strong zonally symmetric component [see also Kutzbach, 1970; Trenberth and Paolino, 1981; Wallace and Gutzler, 1981]. Its structure is characterized by a meridional dipole in SLP not only between centers of action located near Iceland and the Azores/Portugal, but by fluctuations in atmospheric mass between the entire Arctic basin and the surrounding zonal ring.

In this section, we examine the structure of the NAO in the atmospheric circulation, as defined on the basis of the leading EOF of SLP and its associated expansion coefficient (referred to as the principal component, or PC) time series (as noted later in this chapter, the analysis is not sensitive to the choice of SLP as a base level). Following Thompson and Wallace [2000], we present evidence of the similarity between the resulting structure and the leading mode of variability in the SH circulation, defined here as the leading EOF of the SH ( $90^{\circ}\text{S}$ - $20^{\circ}\text{S}$ ) 850-hPa geopotential height field. We use 850- hPa geopotential height in the SH to partially alleviate the ambiguities introduced by the reduction to sea-level over the Antarctic Plateau. In both hemispheres, the analysis is based on data from the NCEP/ NCAR Reanalysis [Kalnay *et al.*, 1996].

Through the rest of this chapter, we will refer to structures derived from the leading EOFs of the NH and SH geopotential height fields as the NH and SH annular modes (NAM and SAM), respectively, and the corresponding expansion coefficient time series of these patterns as the NAM and SAM indices (the NH annular mode is also known as the Arctic Oscillation, e.g., Thompson and Wallace 1998). As noted later in this chapter, the NAM and SAM indices can

be found as the projection of geopotential height anomalies at various levels onto the structure of the corresponding leading EOF. The annular mode nomenclature is used for three reasons: 1) it suggests that the NAO reflects dynamics that would occur in the absence of the North Atlantic ocean; 2) it emphasizes the analogy between the two hemispheres; and 3) it acknowledges the fact that the polar center of action of the NAO has a high degree of zonal symmetry. In practice, the NAM-index time series used in this section is highly correlated with the leading PC time series calculated for SLP data over the North Atlantic half of the hemisphere (the leading PC of NH SLP is correlated with the leading PC of SLP over the North Atlantic half of the hemisphere at a level of  $r=0.95$ ; *Deser, 2000; Hurrell et al., this volume*). Hence, the results presented here are not strongly dependent on the specific definition of the NAO/NAM index. The implications of recasting the NAO as the NAM are discussed further in Section 3.

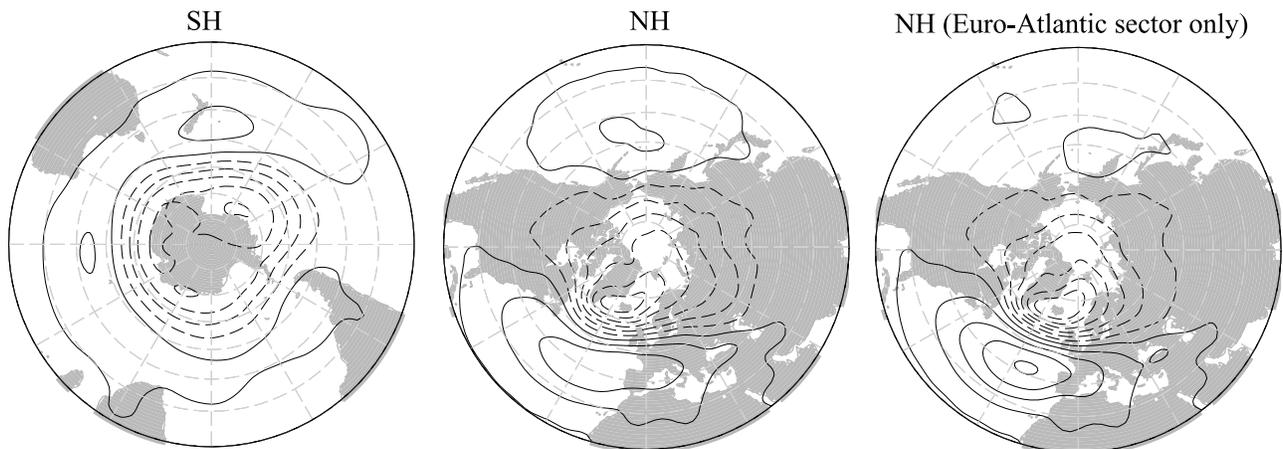
While the results presented in this section are based on monthly-mean data, it should be born in mind that the NAM and the SAM also exhibit variability on submonthly timescales: the  $e$ -folding timescale of the annular modes in the troposphere is  $\sim 10$  days [*Hartmann and Lo, 1998; Feldstein, 2000*], and somewhat longer in the NH during winter. The intramonthly variability in the NAM does not impact the structures derived in this section, but it is further evidence that the annular modes would exist in the absence

of coupled-ocean atmosphere interactions. The processes that give rise to the  $\sim 10$  day timescale of the NAM are explored in Section 4.

### 2.1. Structure

Following *Thompson and Wallace [2000]*, the structure of the NAM in the SLP field is found by regressing monthly mean SLP anomalies onto standardized values of the NAM index (positive values of the annular mode indices are defined as low SLP over the polar regions, and vice versa). As evidenced in Figure 1 (middle), the resulting pattern is characterized not only by a meridional seesaw in SLP over the North Atlantic sector, but by a hemispheric scale seesaw of atmospheric mass between polar latitudes and centers of action located over both the North Pacific and the North Atlantic. The structure of the NAM in lower tropospheric geopotential height bears a strong resemblance to that of the SAM, which is found by regressing lower tropospheric geopotential height in the SH onto the SAM-index time series (Figure 1, left). The patterns in the left and middle panels of Figure 1 are both predominantly zonally symmetric (the SH pattern somewhat more so), and both exhibit similar amplitudes and meridional scales. As noted earlier, the pattern shown in the middle panel of Figure 1 is virtually identical to the leading EOF of the Euro-Atlantic sector (Figure 1, right panel; the EOF of the Euro-Atlantic sector is

Leading EOFs of the lower tropospheric geopotential height field



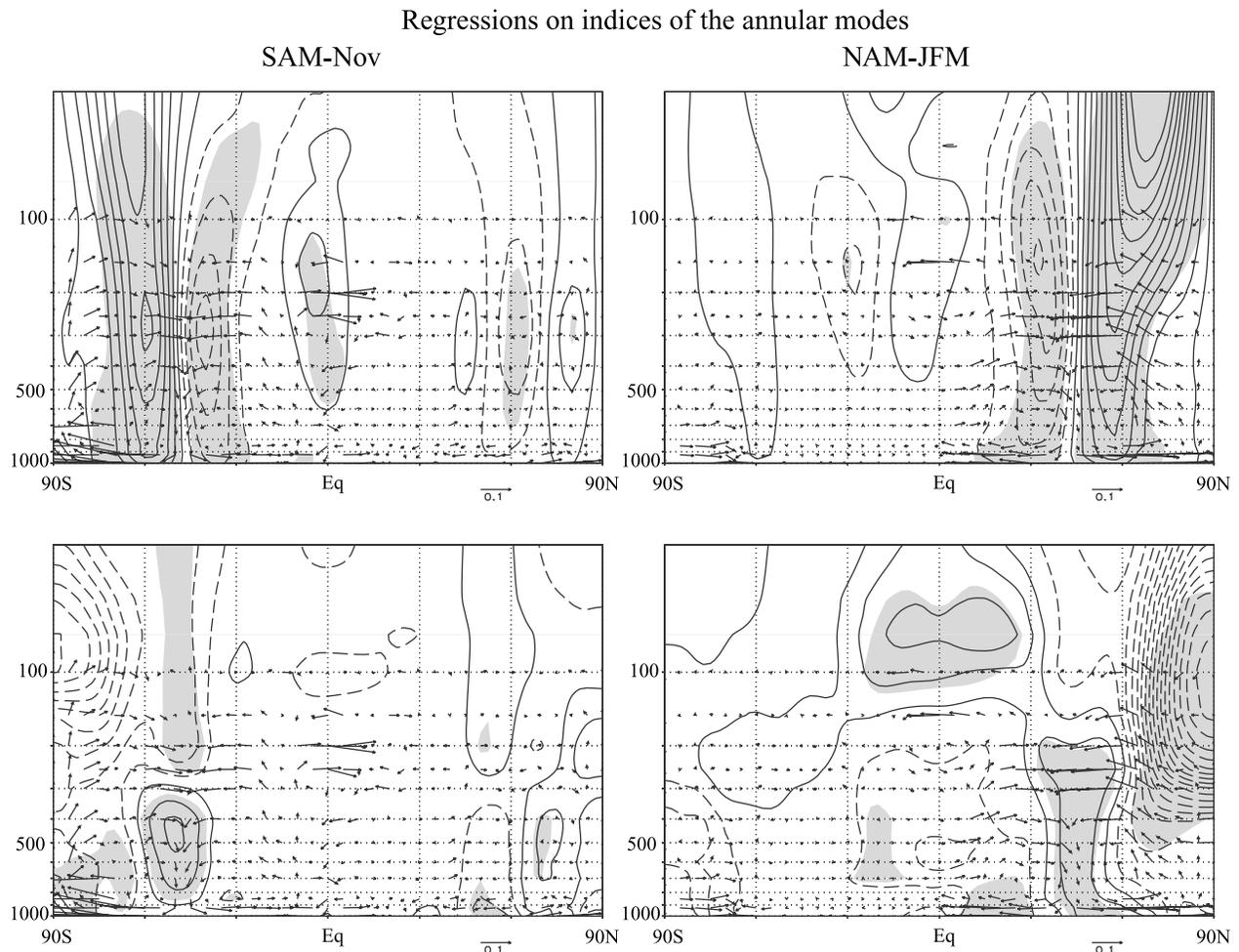
**Figure 1.** The leading empirical orthogonal function (EOF 1) of the Southern Hemisphere (SH;  $90^{\circ}\text{S}$ - $20^{\circ}\text{S}$ ) monthly-mean 850-hPa height field (*left panel*); the Northern Hemisphere (NH;  $20^{\circ}\text{N}$ - $90^{\circ}\text{N}$ ) monthly-mean SLP field (*middle panel*); and the monthly-mean SLP field in the Euro-Atlantic sector ( $20^{\circ}\text{N}$ - $90^{\circ}\text{N}$ ,  $60^{\circ}\text{W}$ - $30^{\circ}\text{E}$ ) (*right panel*). The pattern in the right panel has been extended to include the entire hemisphere by regressing the monthly-mean SLP field upon the corresponding principal component time series. SLP is displayed in units of geopotential height at 1000-hPa. Contour interval 10 m (-5, 5, 15...); negative contours are dashed. Results are based on monthly-mean fields of the NCAR/NCEP Reanalyses (January- March for the NH; all months for the SH) for the period 1958-99. Figures duplicated from *Thompson and Wallace [2000]* and *Wallace and Thompson [2002]*.

extended to the hemispheric domain by regressing hemispheric SLP anomalies onto the respective PC time series). Both patterns have a strong zonally symmetric component, but the hemispheric EOF has larger amplitude over the Pacific sector.

The similarity between the NAM and SAM is even more apparent when one compares their vertical structures. The top panels in Figure 2 show the zonal-mean zonal wind and mean meridional circulations regressed onto “active” season segments of the NAM and SAM indices for the domains extending from pole to pole and from 1000-hPa to 30-hPa. The active seasons are defined as times of year when troposphere/stratosphere coupling is most vigorous, and corre-

spond to periods when the zonal flow in the lower stratosphere is disturbed by waves dispersing upwards from the troposphere. Theory predicts that these interactions should occur during seasons when the zonal flow in the lower stratosphere is westerly, but less than a threshold value [Charney and Drazin, 1961]. Observations reveal that these interactions occur throughout the winter in the NH, but only during late spring (November) in the SH [Thompson and Wallace, 2000].

During the active seasons, both the NAM and the SAM are characterized by equivalent barotropic, meridional dipoles in the extratropical circulation of their respective hemispheres. The annular modes are also marked by weak zonal wind anomalies in tropical latitudes (the anomalies



**Figure 2.** Top: Zonal-mean zonal flow (contours) and mean meridional circulation (vectors) regressed onto the standardized monthly time series of the annular modes for the “active seasons” of stratosphere/troposphere coupling based on monthly data, January-March (NH) and November (SH), from 1979-1999. Bottom: As in the top panel but contours are for zonal-mean temperature. Contour intervals are  $0.5 \text{ m s}^{-1}$  ( $-0.75, -0.25, 0.25$ ) for zonal wind and  $0.2 \text{ K}$  ( $-0.3, -0.1, 0.1$ ) for temperature. Vectors are in units of  $\text{m s}^{-1}$  for the meridional wind component;  $\text{cm s}^{-1}$  for the vertical component (scale at bottom). Shading indicates correlations of  $r > 0.4$ . The top of the diagram corresponds to 50-hPa. Figure adapted from Thompson and Wallace [2002].

are somewhat larger in association with the SAM), which are consistent with global profiles of SLP observed in association with the annular modes [Baldwin, 2001]. The anomalies in the poleward center amplify with height from the surface upward into the lower stratosphere, the axis tilting slightly poleward and the meridional scale broadening from the troposphere to the lower stratosphere [see also Black, 2002]. The zonal wind anomalies located at  $\sim 35^\circ$  N extend equatorward into the subtropics near the surface (particularly in the NH), but exhibit a narrow maximum at the 200-hPa level. A more detailed treatment of the linkage between the NAM and the tropical zonal wind anomalies is presented in the next subsection.

The patterns derived by regressing the mean meridional circulation onto the NAM and SAM indices are shown by the vectors in Figure 2. Both the NAM and the SAM are marked by paired subtropical and high latitude circulation cells that rotate in the opposite sense. The high index polarity of the annular modes is accompanied by anomalous rising motion over subpolar latitudes and subsidence between  $\sim 40$  and  $50^\circ$  latitude. The high latitude cells extend into the lower stratosphere, while the subtropical cells are confined to the troposphere. As noted in Thompson and Wallace, [2000], the Coriolis force acting on the upper branch of the circulation cells acts to weaken the corresponding zonal wind anomalies. Hence, consistent with findings reported in Yoden *et al.* [1987], Shiotani [1990], Karoly [1990], Kidson and Sinclair [1995], Hartmann and Lo [1998], and Limpasuvan and Hartmann [1999; 2000], the wind anomalies associated with the high index polarity of the annular modes must be maintained by anomalous poleward eddy fluxes of westerly momentum in the upper troposphere centered near  $45^\circ$  latitude. The role of eddy fluxes in driving the NAM is discussed further in Section 4. The mean meridional circulation anomalies in the tropics are noisy and difficult to interpret.

The lower panels in Figure 2 show zonal-mean temperature regressed onto the same indices used as a basis for the analysis in the top panels. The positive polarity of the annular modes is marked by anomalously low temperatures over the polar cap that amplify with height, and by anomalously high temperatures in a band centered near  $45^\circ$ . That the temperature anomalies are adiabatically driven is suggested by the fact that cool anomalies tend to coincide with anomalous rising motion, and vice versa. Note that rising motion in regions of cooling is consistent with mechanical driving by anomalous eddy fluxes of zonal momentum. The shallow warm anomalies near the surface between  $55^\circ$  N and  $75^\circ$  N are consistent with anomalous temperature advection over the NH landmasses during the positive polarity of the NH annular mode [Hurrell, 1995a; 1996; Thompson and

Wallace, 1998; 2000; Xie *et al.*, 1999].

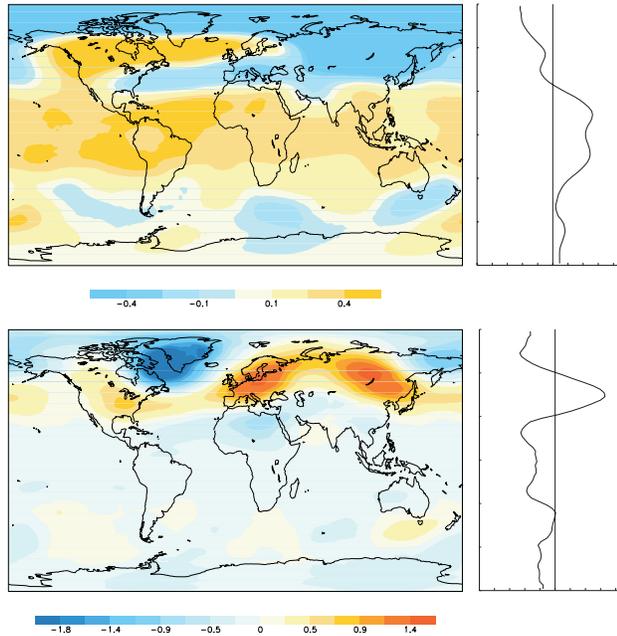
The positive polarity of the NAM is also characterized by positive temperature anomalies at the tropical tropopause. Since water vapor is sparse in the lower stratosphere, the out-of-phase relationship between lower stratospheric temperature anomalies at tropical and polar latitudes presumably reflects adiabatic temperature changes induced by a weakening of the mean-meridional circulation of the lower stratosphere during the high index polarity of the NAM. The SAM has a much weaker signature in tropical tropopause temperature, which is consistent with the fact that stratosphere/troposphere coupling occurs during a much shorter season in the SH. As discussed in the next section, the NAM is also associated with weak negative temperature anomalies throughout the tropical troposphere.

## 2.2. Global-scale Features of the NAM

Several studies have noted a linkage between the NAM and variability in the tropical Atlantic region. For example: Meehl and van Loon [1979] have shown that the NAO is significantly correlated with the strength of the trade-winds over the North Atlantic and with the position of the ITCZ over Africa; Lamb and Pepler [1987] have noted that the NAO has a significant impact on Moroccan rainfall; Moulin *et al.* [1997] have demonstrated that the NAO impacts dust transport from the Sahara desert; Malmgren *et al.* [1998] have shown that the NAO is reflected in Puerto Rican rainfall; and McHugh and Rogers [2001] have suggested that the NAO impacts eastern African rainfall. Baldwin [2001] has shown that fluctuations in the NAM are characterized by significant interhemispheric exchanges of atmospheric mass. Here we offer evidence that the NAM has a distinct signature in the temperature and zonal wind field of the tropics that transcends the tropical Atlantic sector.

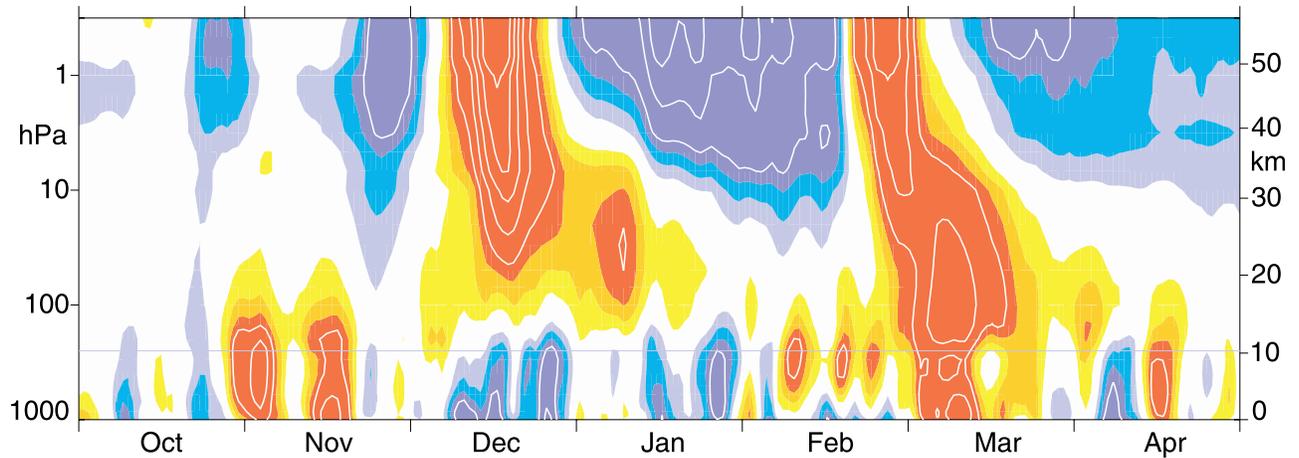
Plate 1 shows tropospheric temperature anomalies from the Microwave Sounding Unit Channel 2LT data [Spencer *et al.*, 1990; Spencer and Christy, 1992; the weighting function for MSU2LT temperature data is centered near 600 hPa] and lower stratospheric temperature anomalies from the MSU4 data [Spencer and Christy, 1993; the weighting function for MSU4 data is centered near 70 hPa] regressed onto JFM monthly values of the NAM index from 1979-1999. Consistent with results presented in the previous section, the high index polarity of the NAM (low SLP over the pole) is characterized by pronounced cooling over the NH polar cap that amplifies with height from the troposphere to the lower stratosphere, and warming in the midlatitude troposphere. However, the signature of the NAM is clearly not restricted to the extratropics of the NH. The regressions also reveal a distinct pattern of temperature anomalies through-

MSU2LT and MSU4 temperature regressed on the NH annular mode

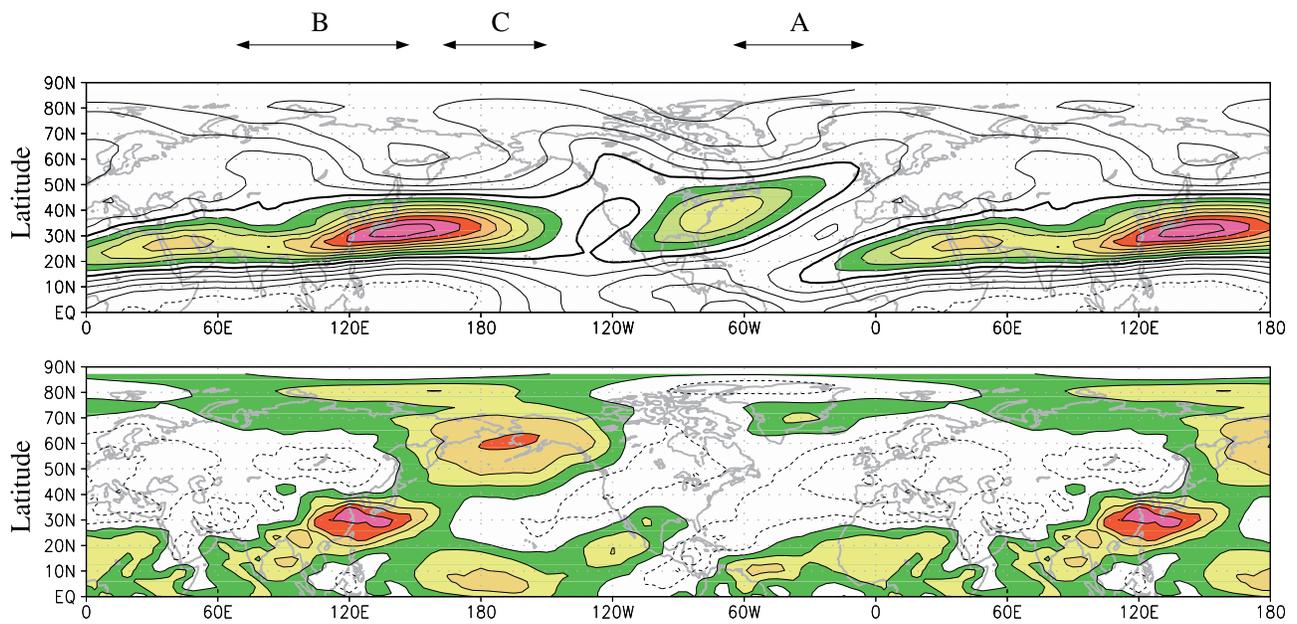


**Plate 1.** Left panels: MSU4 (top) and MSU2LT (bottom) temperature anomalies (K) regressed upon January–March (JFM) standardized monthly values of the NAM-index. Right panels: The correlation coefficients ( $r$ ) between JFM values of the NAM index and zonal mean MSU4 (top) and MSU2LT (bottom) temperature anomalies. Tickmarks are at  $r = 0.2$ . Figure from research in progress in collaboration with D. J. Lorenz (University of Washington).

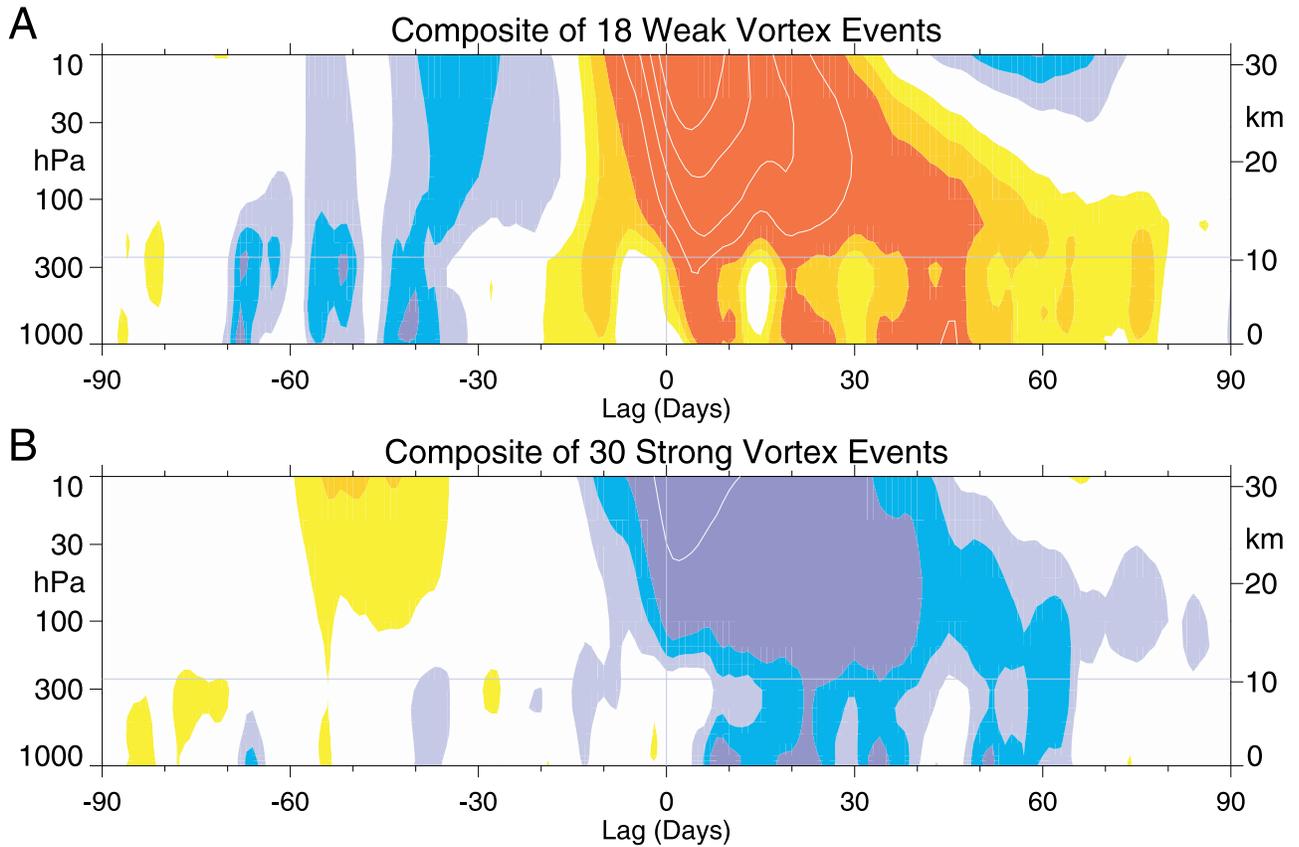
1998 - 1999 Northern Annular Mode



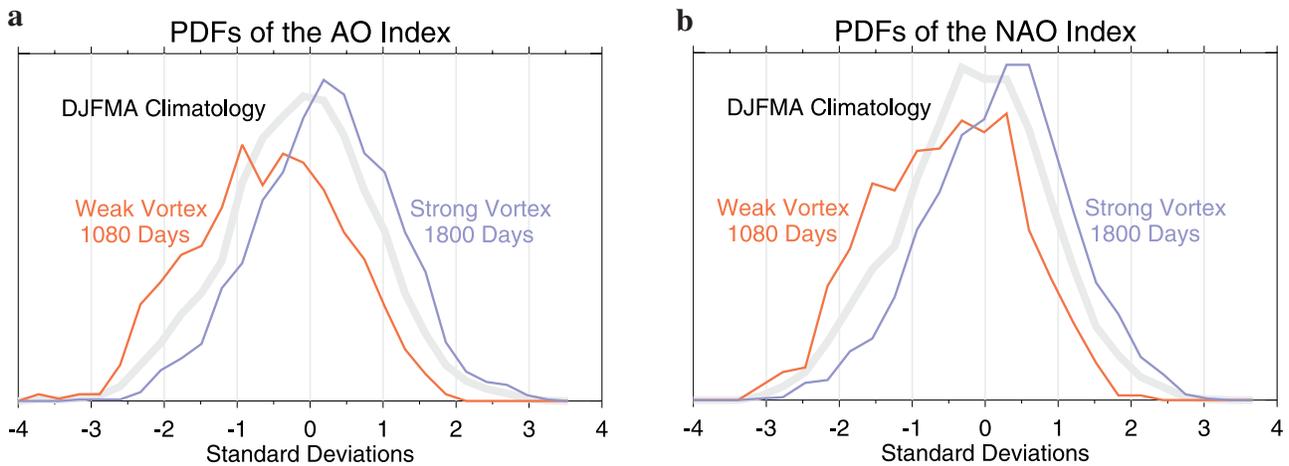
**Plate 2.** Time-height development of the NAM during the winter of 1998–1999. The indices have daily resolution and are nondimensional. Blue corresponds to positive values (strong polar vortex) and red corresponds to negative values (weak polar vortex). The contour interval is 0.5, with values between -0.5 and 0.5 unshaded. The thin horizontal line indicates the approximate boundary between the troposphere and the stratosphere. From Baldwin and Dunkerton [2002].



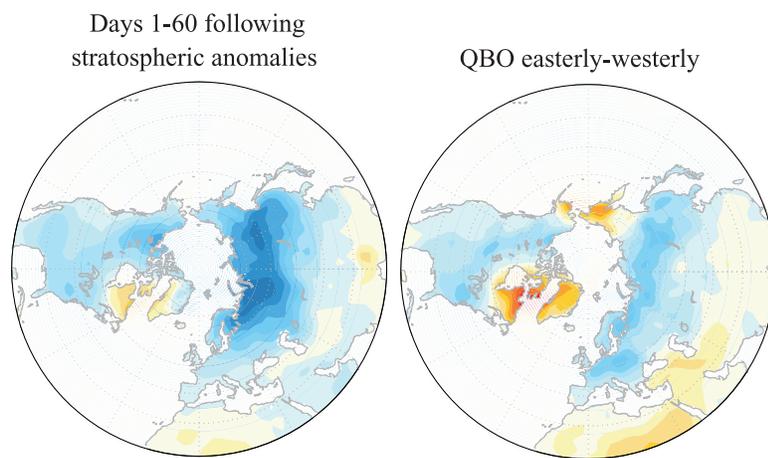
**Plate 3.** The NH 300-mb (a) zonal and (b) meridional winds averaged over December-January-February, 1958-1997. Contour interval is 5 m s<sup>-1</sup> for (a) and 2 m s<sup>-1</sup> for (b). Shading is above 20 m s<sup>-1</sup> for (a) and 0 m s<sup>-1</sup> for (b). Reproduction from *Lee and Kim* [2002].



**Plate 4.** Composites of the time-height development of the NAM for (a) 18 weak vortex events and (b) 30 strong vortex events. The events are determined by the dates on which the NAM at 10-hPa crosses  $-3.0$  and  $+1.5$ , respectively. The indices are nondimensional; the contour interval for the color shading is 0.25, and 0.5 for the white contours. Values between  $-0.25$  and  $0.25$  are unshaded. The thin horizontal lines indicate the approximate boundary between the troposphere and the stratosphere. From *Baldwin and Dunkerton* [2002].



**Plate 5.** (a) Probability density function for the normalized daily NAM index during December–April (gray curve), the 1080 days during weak vortex conditions (red curve), and the 1800 days during strong vortex conditions (blue curve). (b) As in (a), but for the index of the NAM based solely on North Atlantic data. From *Baldwin and Dunkerton* [2002].



**Plate 6.** The difference in daily mean surface temperature anomalies between the 60-day interval following the onset of weak and strong vortex conditions at 10-hPa (left) and between Januarys when the QBO is easterly and westerly (right panel). The samples used in the analysis are documented in *Thompson et al.* [2002]. Contour levels are at 0.5 C. From *Thompson et al.* [2002].

out the tropics, with warm anomalies found at the tropical tropopause region, and cool anomalies found throughout much of the tropical troposphere. Tropical-mean ( $20^{\circ}\text{S}$ - $20^{\circ}\text{N}$ ) tropopause temperatures are 0.36 K warmer and tropical-mean tropospheric temperatures are 0.12 K cooler per one standard deviation increase in the NAM index (the correlation coefficient between the NAM index and tropical-mean tropopause temperature is  $r=+0.49$ , and the correlation between the NAM index and tropical mean tropospheric temperature is  $r=-0.32$ . Both correlations are significant at the 95% level based on the  $t$ -statistic). In both the troposphere and stratosphere, the largest temperature anomalies in the tropics are found, not over the equator, but over the subtropical regions of both hemispheres, as clearly evidenced in the correlations between zonal mean temperature and the annular mode indices (Plate 1, right panels).

The cooling of the tropical troposphere observed in the MSU2LT data is more pronounced than that observed in surface temperature data (not shown), which hints that the mid-tropospheric anomalies are dynamically induced through eddy-driven vertical motions. For example, tropical mean surface temperature decreases by only 0.06 K per standard deviation increase in the NAM index [based on data described in Jones, 1994] in contrast to the corresponding 0.12 K decrease observed in the MSU2LT data. To what extent the NAM has contributed to the widely publicized discrepancy between trends in temperatures at the surface and in the free troposphere remains to be assessed [see the National Research Council Report on Reconciling Temperature Observations for an overview of the discrepancies between recent tropospheric and surface temperature trends].

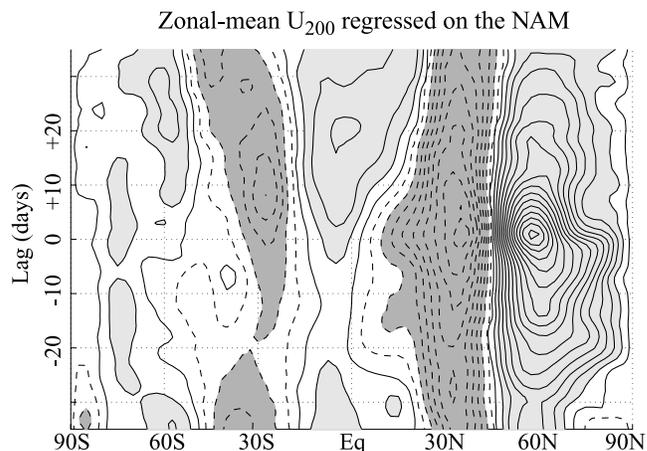
Regressions based on daily data reveal that the linkage between the NAM and the circulation of the tropics is most pronounced when the tropical circulation lags variability in the NAM by several weeks (the daily NAM index used in the regressions was generated by projecting daily-mean fields of SLP from the NCEP/NCAR Reanalysis onto the structure in the middle panel of Figure 1). Figure 3 shows regression coefficients between the daily-mean zonal-mean zonal wind anomalies at 200-hPa and daily values of the NAM index for lags ranging from -35 to +35 days, where positive lags indicate the zonal flow lags the NAM index, and vice versa. The zonal wind anomalies in the tropics and in the SH are weak in the simultaneous regressions (i.e., at lag 0), but they exhibit a marked structure roughly ~2-3 weeks following the development of the largest wind anomalies in the extratropical NH. Hence, the NAM does not project strongly onto the tropics when the daily-mean zonal-mean circulation is regressed on contemporaneous daily values of the NAM index (Figure 4, top). However, when the zonal-mean circulation in Figure 4 is regressed onto

daily values of the NAM index lagged by two weeks (bottom), the resulting patterns exhibit pronounced features in both the tropical troposphere and subtropical SH. The zonal wind anomalies in the bottom panel of Figure 4 reflect a banded structure that extends all the way from the polar regions of the NH deep into the subtropics of the SH. The distinct off-equatorial cooling maxima evident in the monthly regressions shown in Plate 1 are also clearly evident in these lagged daily regressions.

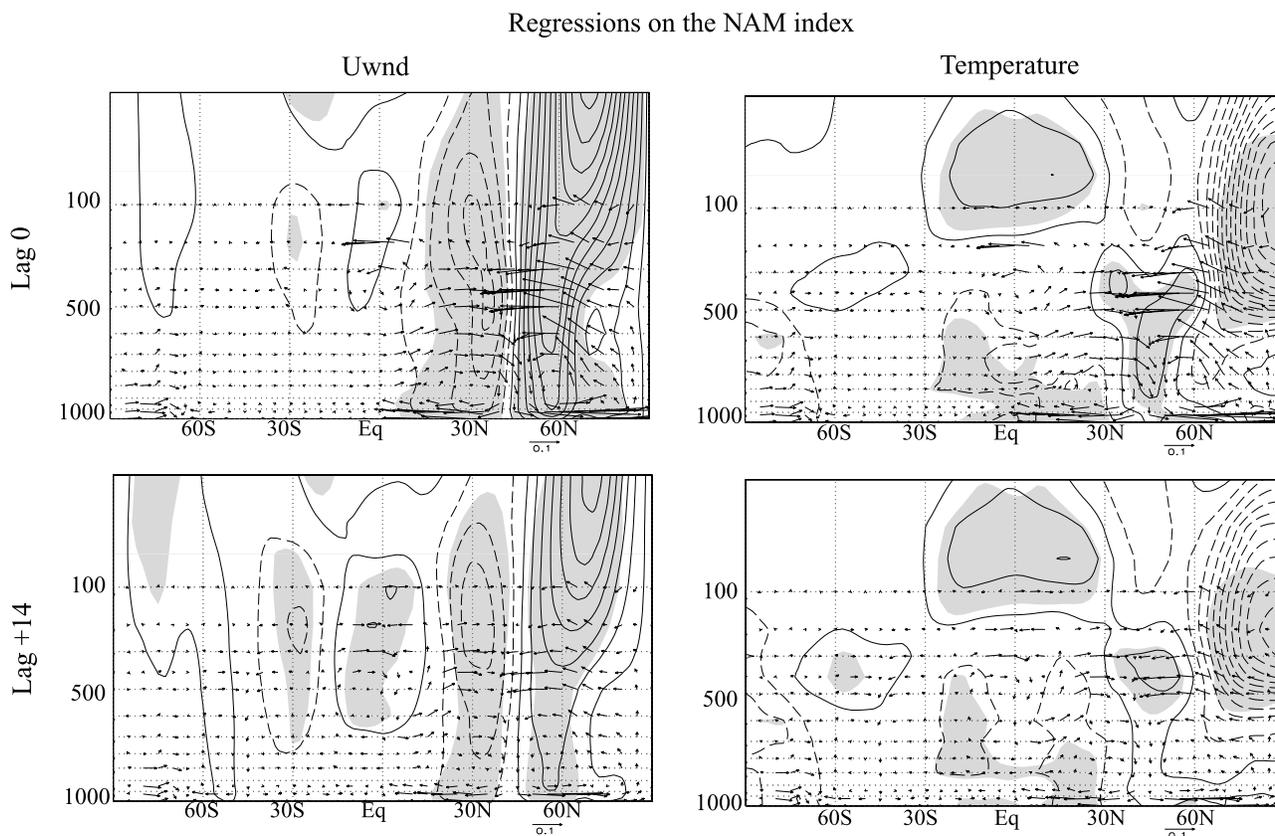
As discussed earlier, the warming of the tropical tropopause during the high index phase of the NAM is consistent with a weakening of the wave-driven Brewer-Dobson circulation in the lower stratosphere [Thompson and Wallace, 2000]. The processes that drive the tropical tropospheric features evident in Plate 1 and Figs. 3-4 remain to be determined, but the lag between the extratropics and tropics suggest that they may reflect dynamics associated with waves originating in the extratropical NH.

### 3. THE NAO AS AN ANNULAR MODE

As noted at the beginning of the previous section and in Stephenson *et al.* [this volume], the NAM has been widely viewed as a fundamentally zonally asymmetric structure, i.e., a pattern that owes its existence to regional sources of heat and momentum. Month-to-month variability in the NAM has been interpreted as a reflection of dynamics unique to the North Atlantic stormtrack. Decadal variability in the NAM has been interpreted as a manifestation of cou-



**Figure 3.** Zonal-mean zonal flow at 200-hPa regressed onto the standardized daily time series of the NAM for lags as indicated. Results based on JFM data from 1979-1999. Positive lags are for the zonal flow lagging the NAM index, and vice versa. Contour intervals are  $0.2 \text{ m s}^{-1}$  ( $-0.3, -0.1, 0.1$ ). Values exceeding  $\pm 0.2 \text{ m s}^{-1}$  are shaded. Figure from research in progress in collaboration with D. J. Lorenz (University of Washington).



**Figure 4.** Top panels. Left: Zonal-mean zonal flow (contours) and mean meridional circulation (vectors) regressed onto the standardized daily time series of the NAM based on JFM data from 1979- 1999. Right: As in the left panel but contours are for zonal-mean temperature. Bottom panels. As in the top panels but the data are lagged by +14 days with respect to the NAM index. Contour intervals are  $0.5 \text{ m s}^{-1}$  ( $-0.75, -0.25, 0.25$ ) for zonal wind and  $0.2\text{K}$  ( $-0.3, -0.1, 0.1$ ) for temperature. Vectors are in units of  $\text{m s}^{-1}$  for the meridional wind component;  $\text{cm s}^{-1}$  for the vertical component (scale at bottom). Shading indicates correlations that exceed the 95% confidence level based on the t-statistic. Figure from research in progress in collaboration with D. J. Lorenz (University of Washington).

pled atmosphere/ocean interactions over the North Atlantic sector.

Recently, it has been suggested that NAM reflects dynamical processes that transcend the land-sea geometry and orography of a particular hemisphere, and hence can be viewed as an annular mode analogous to the leading mode of variability in the SH circulation [Thompson and Wallace, 1998; 2000]. In this case, the NAM favors in-phase fluctuations in the atmospheric circulation along latitude circles not only over the North Atlantic sector, but throughout much of the Northern Hemisphere.

The “annular mode” perspective is not unanimously embraced by the research community [e.g., see Kerr, 1999]. As such, it has generated a debate over our fundamental understanding of what is arguably the most important pattern of NH climate variability. The debate is centered on three different perspectives, each of which defines a unique way of viewing the identity of the NAM:

- 1) the “historical perspective”, in which the NAM reflects dynamics unique to the North Atlantic sector;
- 2) the “annular mode perspective”, in which the NAM is viewed as the NH analogue of the SH annular mode, and both annular modes are viewed as organizing extratropical variability on a hemispheric scale;
- 3) the “regional perspective”, in which the analogy between the NAM and the SAM is acknowledged, but the annular modes are viewed as statistical artifacts of locally occurring stormtrack dynamics.

In this section, we outline the principal arguments that support the annular mode perspective over the historical perspective, summarize the principal arguments that favor the regional perspective, and discuss why the choice of perspective matters. We begin with a brief review of the history of zonally symmetric patterns of variability.

### 3.1. A Brief History of Zonally Symmetric Modes of Variability

Interest in zonally symmetric patterns of variability in the NH circulation can be traced back to at least the late 1930s, when C. G. Rossby and collaborators at the Massachusetts Institute of Technology postulated that the positions of the NH wintertime semipermanent centers of action are dependent on the intensity of the zonally symmetric circulation [Rossby, 1939]. Rossby noted that the longitude of the Aleutian Low was highly correlated with the mean pressure difference between 35°N and 55°N (i.e., the strength of the zonal wind along 45°N) during the winter of 1938-39. That the position of the Aleutian Low lagged the zonal wind at 45°N was cited as evidence that the strength of the zonal circulation impacted the displacement of longitudinally dependent centers of action.

Willett [1948] and Rossby and Willett [1948] ascribed this variability to a distinctive physical phenomenon, whereby low frequency climate variability of the NH circulation is marked by fluctuations between two discrete climate states, defined on the basis of the meridional pressure gradient between 35°N and 55°N. The low index state (weak zonal flow along 45°N) was defined by an extensive polar vortex and pronounced baroclinicity at middle latitudes; the high index state by a contracted polar vortex and pronounced baroclinicity at high latitudes.

A slightly different interpretation of zonally symmetric variability was suggested by Lorenz [1951]. Upon examining correlation statistics between zonal mean SLP at latitudes throughout the hemisphere, Lorenz concluded that the dominant mode of variability of the zonally symmetric circulation corresponded, not to the zonal mean zonal wind along 45°N, but to the zonal mean wind along 55°N. A similar conclusion was reached by Namias [1950], who proposed that the leading mode of variability in the zonal mean circulation was, in fact, characterized by meridional shifts in the strength of the zonal flow between ~35°N and ~55°N. However, unlike Lorenz, Namias argued that fluctuations in the zonal circulation were indicative of a physical phenomenon with cyclical behavior: the so-called “zonal index cycle”.

Owing to its purported cyclical behavior, the “zonal index cycle” in the zonal-mean flow was presumed to be of practical use for long range weather forecasting. Namias theorized that the transition from the high index state (contracted polar vortex) to the low index state (extended polar vortex) would occur when cyclogenesis broke out simultaneously within sectors throughout the hemisphere. Namias reasoned that the anomalous equatorward vorticity fluxes and advection of cold air would displace the baroclinic zone

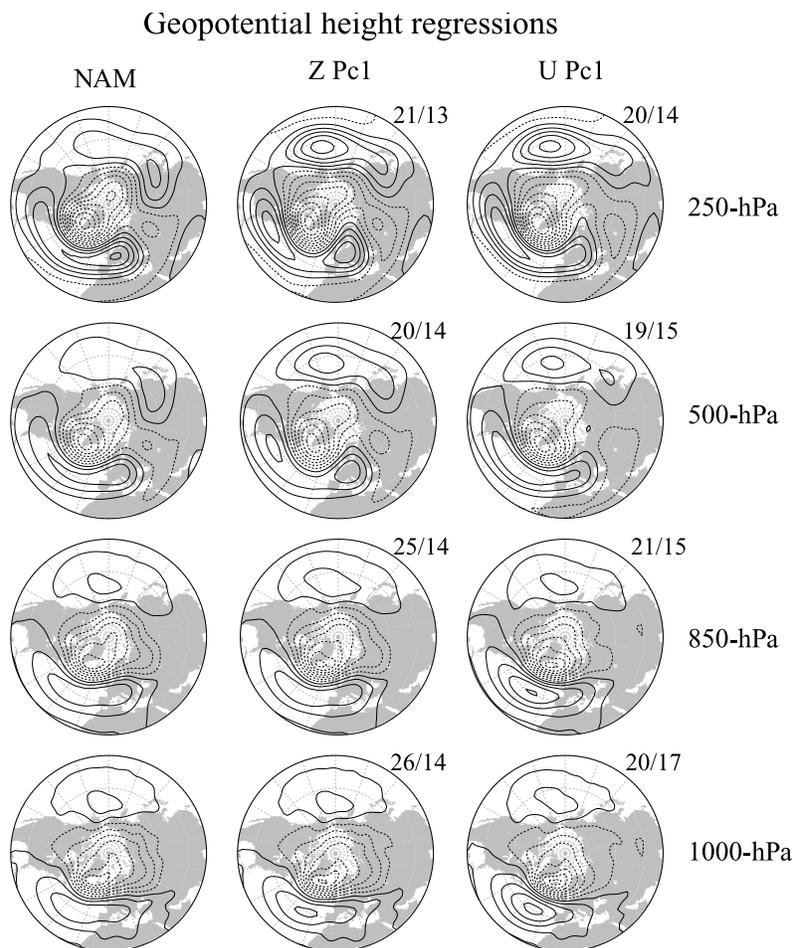
and zonal wind maximum equatorward, leading to the low index state. The transition from the low index state back to the high index state was subsequently accomplished through the re-establishment of the high latitude baroclinicity through diabatic heating, and the weakening of the low latitude zonal wind anomalies through frictional dissipation. However, when this cyclical behavior and the attendant coordinated behavior of NH weather systems at widely separated longitudes failed to materialize in observational data, the zonal-index paradigm was largely abandoned.

With the demise of interest in the NH zonal index cycle, both observational and theoretical research of large-scale climate variability in the NH focused increasingly on wavelike structures in the middle tropospheric geopotential height field, referred to as “teleconnection” patterns [e.g., van Loon and Rogers, 1978; Hoskins and Karoly, 1981; Wallace and Gutzler, 1981; Barnston and Livezey, 1987]. Unlike the zonal-index, these patterns were presumed to owe their existence to fundamentally zonally asymmetric sources of heat and/or momentum. Of the myriad teleconnection patterns that have been documented in the literature [see Wallace and Gutzler, 1981; Barnston and Livezey, 1987 for a survey of teleconnection patterns], two have been widely viewed as the principal patterns of variability in the NH extratropical circulation [e.g., Kushnir and Wallace, 1987]: the Pacific North-America pattern, and the NAO.

### 3.2. The Reemergence of the Zonal Index Paradigm

The NH annular mode perspective suggests that the NAO is not a regional “teleconnection” pattern, but that it is an expression of a zonally symmetric pattern reminiscent of the structure examined by Rossby, Willett, Lorenz, and Namias in the NH [Thompson and Wallace, 1998; 2000; Wallace, 2000]. The principal arguments in favor of the annular mode perspective over the view that the NAO reflects dynamics unique to the North Atlantic stormtrack are:

- The structure of the NAM is virtually identical to that of the widely recognized SH annular mode (as documented in the previous section).
- The zonally symmetric component of the NAM is evident in the leading EOFs of the zonally varying geopotential height and zonal wind fields at levels throughout the depth of the troposphere and stratosphere. Figure 5 shows geopotential height anomalies at various levels of the troposphere regressed on JFM values of the NAM-index (left panels), the leading PC time series calculated for the corresponding two-dimensional zonally varying geopotential height field (middle panels), and the leading PC time series calculated for the corresponding two-dimensional zonally varying

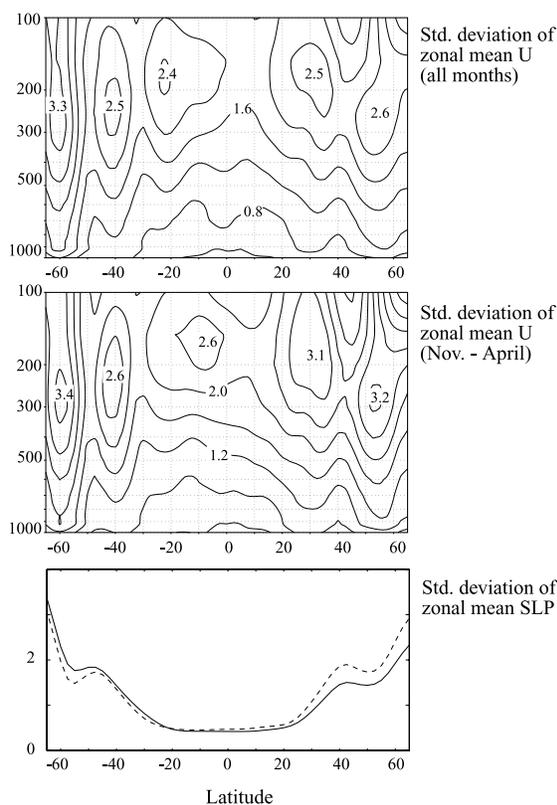


**Figure 5.** (Left) Monthly-mean values of geopotential height regressed on JFM values of the NAM-index at levels indicated. (Middle and right) As in the left panel, but regressions are based on the standardized leading principal component time series of the zonally varying geopotential height field (middle) and zonal wind field (right) calculated for the levels indicated. The variance explained by each PC is noted in the top right corner of the respective panel (mode 1/mode 2). Contour intervals are 10 m (1000Z; 850Z), and 15 m (500Z; 250Z). Figure adapted from *Thompson* [2000].

zonal wind field (right panels). The middle tropospheric patterns in Figure 5 are more wavelike than their lower tropospheric counterparts. Nevertheless, at each level the leading mode of variability of the two-dimensional zonally varying circulation clearly bears a striking resemblance to the corresponding signature of the annular mode. *Baldwin and Dunkerton* [2001] have shown that the NAM is evident in the leading EOFs of the zonally varying circulation up to 0.3-hPa in the lower mesosphere, and *Thompson and Wallace* [2000] have noted that the NAM emerges as the leading EOF of the zonal-mean circulation from the surface to the lower stratosphere.

- The signatures of the annular modes are reflected in the meridional profiles of the month-to-month variance of zonally averaged SLP and in the zonal component of the geostrophic wind from the surface to the upper troposphere.

In the absence of a mode of variability that organizes the zonally symmetric flow, the rms variance of monthly-mean, zonal-mean SLP field would be expected to increase monotonically with latitude as one moves away from the equator. This is because: 1) the meridional gradient in pressure is directly related to the Coriolis parameter,  $f$ , which increases with latitude; 2) the temporal variance of an area-averaged quantity should increase as the averaging area decreases, as is the case in zonal-means at increasingly high latitudes. From Figure 6, it is clear that the rms variance of zonal-mean SLP does not increase monotonically with latitude, but exhibits well defined shoulders near  $45^\circ$  latitude in both hemispheres [bottom panel; a similar “shoulder” was also noted in *Lorenz*, 1951]. Consistent with this feature, the rms variance of the zonal-mean zonal wind field exhibits distinct maxima throughout the depth of the troposphere  $\sim 30\text{-}40^\circ$



**Figure 6.** Standard deviation of zonal-mean zonal wind (top and middle) and zonal-mean SLP (bottom). Calculations are based on monthly-mean anomalies, 1979-1997, for all months (top panel and solid line in bottom panel); and for November-April (middle panel and dashed line in bottom panel). Tickmarks on the vertical axis are at intervals of 1 hPa for SLP. Contour intervals are  $0.4 \text{ m s}^{-1}$  for zonal-mean zonal wind. Figure adapted from *Thompson* [2000].

and  $\sim 55\text{-}60^\circ$  latitude in both hemispheres (top panels). During the NH cold season November-April (middle panel), the NH variance maxima are more pronounced and there is a distinctive variance maximum in the tropics centered  $\sim 10^\circ\text{S}$ .

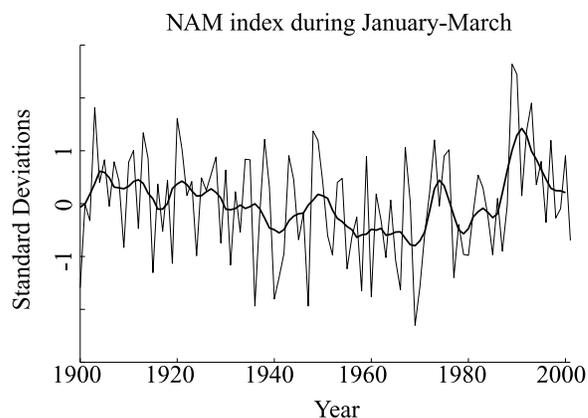
- There has been a trend in the mode in question over the past few decades that is reflected in both the leading PC of NH SLP (i.e., the NAM index; see Figure 7) and in indices based on station data over the North Atlantic [i.e., the NAO index as used in *Hurrell*, 1995a]. However, the hemispheric scale of recent climate trends is better captured by regressions based on the hemispheric NAM- index than regressions based on North Atlantic station data [e.g., see *Thompson and Wallace*, 1998]; Eurasian surface air temperature trends extend as far east as Lake Baikal [*IPCC*, 2001], and sea-level pressure has fallen throughout the Arctic basin [*Walsh et al.*, 1996]. It is unclear why decadal variability in the NAM would have such pronounced amplitude throughout the hemisphere if it is, in fact, a pattern of variability confined to the North Atlantic sector.

- The leading mode of variability of the zonally varying stratospheric circulation more closely resembles a zonally symmetric structure at the surface than a pattern restricted to the North Atlantic sector, albeit the surface signature of this coupling has largest amplitude over the North Atlantic half of the hemisphere [*Thompson et al.*, 2000; *Baldwin and Dunkerton*, 1999].
- Variations in the NAM are reflected in the frequency of occurrence of blocking and cold air outbreaks not only over the Atlantic half of the hemisphere, but over the Pacific half as well [*Thompson and Wallace*, 2001].

### 3.3. The Case for the Regional Perspective

The arguments outlined above suggest that the NAM reflects dynamics that transcend the North Atlantic sector, that would occur in the absence of the North Atlantic ocean, and that are analogous to those that drive the leading mode of variability in the SH. However, the evidence outlined above does not necessarily prove that the NAM (or the SAM) orchestrates variability on a hemispheric scale. The absence of such proof motivates the regional perspective.

The regional perspective acknowledges that the NAM is the NH analogue of the SAM, and it also acknowledges that the NAM reflects dynamics that would occur in the absence of the NH land-sea geometry. However, the regional perspective does not acknowledge that the annular modes are accompanied by in-phase fluctuations along latitude circles, as purported by the annular mode perspective. In essence, the regional perspective argues that the annular modes are statistical artifacts of locally occurring dynamics.



**Figure 7.** Standardized January-March (JFM) mean values of the NAM-index based on sea-level pressure data described in *Trenberth and Paolino* [1980]. Light lines indicate JFM seasonal means; heavy lines indicate 5-year running means. The interval between tick marks on the vertical axis is one standard deviation. Figure adapted from *Thompson et al.* [2000].

The principal observation in favor of the regional perspective is the absence of significant correlations between variability at widely separated longitudes in the NH middle latitudes. For example, *Deser* [2000] found that the Arctic center of action of the NAM is significantly correlated with both the Atlantic and Pacific centers of action, but that the Atlantic center is not significantly correlated with the Pacific center. It follows that the complete NAM pattern is not evident in maps of teleconnectivity (maps based on one-point correlations) based on SLP data for points located over the North Atlantic and North Pacific sectors along  $\sim 45^\circ\text{N}$ .

The regional perspective also questions the reliance of the annular mode perspective on EOF analysis. EOFs are mathematical constructs designed to maximize the amount of variance captured by a single spatial pattern; they do not necessarily correspond to physically meaningful structures. *Ambaum et al.* [2001] and *Dommenget and Latif* [2002] examined idealized statistical models of the NH circulation in which variability in the model Pacific and Atlantic sectors was constrained to be linearly independent. Both found that the leading EOFs in these idealized scenarios corresponded to annular structures that suggest in phase fluctuations between the idealized Atlantic and Pacific sectors, despite the fact that these sectors were uncorrelated by construction. *Ambaum et al.* [2001] also noted that the climatology upon which NAM variability is superposed is strongly zonally asymmetric: positive anomalies in the NAM correspond to a strengthening of the North Atlantic eddy driven jet, but a weakening of the subtropical jet in the Pacific sector.

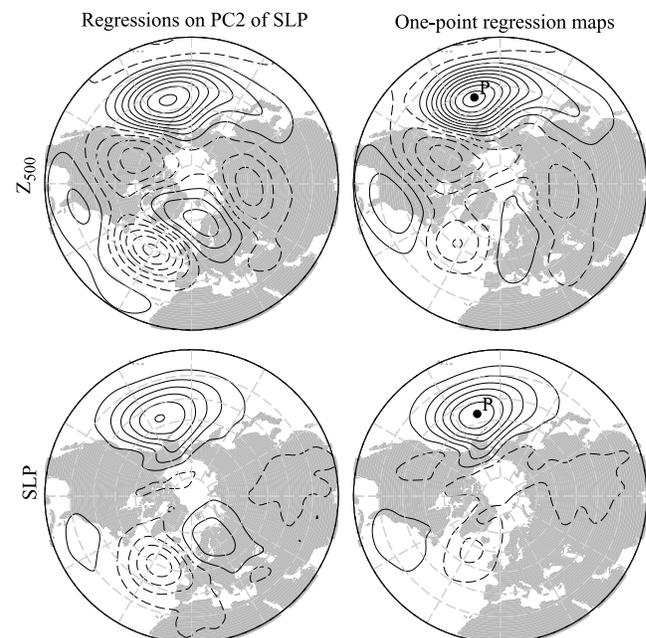
There are two principal observations that oppose the regional perspective's assertion that the annular modes do not orchestrate variability on a hemispheric scale. First, while correlations between geopotential height anomalies at widely separated longitudes are weak in middle latitudes, they are strong at polar latitudes. For example, one-point correlation maps based on geopotential height anomalies at grid points poleward of  $\sim 60^\circ$  latitude yield similarly signed correlations throughout the polar cap [e.g., see *Ambaum et al.*, 2001 for the one-point correlation map based on a reference point near Iceland]. Hence, the annular mode's high latitude center of action varies in phase at all longitudes on month-to-month timescales.

Second, as noted in *Wallace and Thompson* [2002], the observed lack of correlations between the Pacific and Atlantic sectors is entirely consistent with the presence of a zonally symmetric mode *if* a second mode is present that favors out-of-phase correlations between these two sectors. They suggested that the Pacific-North America (PNA) pattern favors such negative correlations. Specifically, they noted that the PNA pattern defined on the basis of the second PC of SLP favors geopotential height anomalies of

opposing sign over the North Pacific and North Atlantic sectors (Figure 8, left panels). As proof of the physical basis of this second PC, they demonstrated that a similar structure emerges in one-point regression maps based on North Pacific geopotential height (Figure 8, right panels). After linearly regressing the PNA pattern out of the SLP data, *Wallace and Thompson* [2002] noted that the correlations between the Atlantic and Pacific sectors are positive and highly significant ( $r > 0.6$ ). They reasoned that a similar argument applies to the idealized statistical model proposed by *Ambaum et al.* [2001].

### 3.4. Why Does the Choice of Perspective Matter?

While the debate over which perspective most accurately describes the NAM remains unresolved, it is worth bearing in mind why the choice of perspective matters [see *Wallace, 2000* for a detailed discussion of this topic]. For example, the choice of perspective has implications for the underlying dynamical processes thought to give rise to NAM-like variability. The historical perspective suggests that NH climate variability occurs on regional, not hemispheric scales, and that the dynamics of the NAM are unique to the North Atlantic stormtrack. In this case, the NAM has often been interpreted as the hallmark of coupled ocean/atmosphere



**Figure 8.** The 500-hPa height field (top) and SLP field (bottom) regressed on standardized time series corresponding to: (left) PC 2 of the NH monthly-mean SLP field; (right) standardized values of SLP at point P. Contour intervals are 10 m (-5, 5, 15...). Negative contours are dashed. Adapted from *Wallace and Thompson* [2002].

interactions over the North Atlantic sector [e.g., *Grotzner et al.*, 1998; *Rodwell et al.*, 1999].

Both the annular and regional perspectives suggest that the NAM is driven by local wave-mean flow interactions reminiscent of the dynamics that dominate extratropical variability in the SH atmosphere. However, while the regional perspective implies that locally occurring storm-track dynamics are primarily coupled with the local circulation [e.g., as in the relationships documented in *Lau*, 1988], the annular mode perspective suggests that the hemispheric scale circulation plays an important role in organizing local eddy activity, presumably through positive feedbacks between the zonal-mean flow and transient wave disturbances [e.g., *Lorenz and Hartmann*, 2001; 2002].

The choice of perspective also has implications for the coupling between the extratropical stratospheric and tropospheric circulations. The regional perspective suggests that the observed coupling between the stratosphere and troposphere is accomplished via a distinctive wavelike pattern in the middle troposphere. In contrast, the annular mode perspective suggests that the coupling is intrinsic to the dynamics of the zonally symmetric polar vortex itself, i.e., that dynamical processes at stratospheric levels can impact the strength of the polar vortex throughout the depth of the troposphere via the combined effects of an induced, thermally indirect mean meridional circulation analogous to the Ferrell cell and induced changes in the poleward eddy fluxes of zonal momentum at intermediate levels.

The contrasting mechanisms implied by the choice of perspective underscores the fact that the fundamental physical processes that underlie the NAM are poorly understood. In the next section, we explore these fundamental processes in greater detail. The discussion explores research questions that are relevant regardless of which perspective one subscribes to.

#### 4. TROPOSPHERIC DYNAMICS

As of this writing, there is no clear consensus regarding the processes that govern the existence of annular modes. The ubiquity of annular modes in not only the extratropical SH [e.g., *Karoly*, 1990; *Hartmann*, 1995; *Hartmann and Lo*, 1998; *Kidson and Watterson*, 1999; *Limpasuvan and Hartmann*, 1999; 2000; *Lorenz and Hartmann*, 2001], but also simple two-layer models [*Robinson*, 1991; 1994; 1996; *Lee and Feldstein*, 1996], multi-layer numerical simulations with zonally symmetric boundary conditions [*James and James*, 1992; *Yu and Hartmann*, 1993; *Feldstein and Lee*, 1996], and general circulation models with realistic land-sea geometry [*Fyfe et al.*, 1999; *von Storch*, 1999; *Shindell et al.*, 1999; 2001; *Yamazaki and Shinya*, 2000] attests to the

robustness of the dynamical process to which they owe their existence. Nevertheless, exactly what fundamental processes govern the amplitude, meridional scale, and zonal structure of annular modes remains open to debate.

What is clear is that annular-mode like variability requires wave-induced momentum transport. For example, consider the quasi-geostrophic zonal-mean zonal momentum equation on a beta plane [e.g., see *Holton*, 1992]:

$$\frac{\delta}{\delta t}[u] = \frac{\delta}{\delta y}[u'v'] + f_0[v] + friction \quad (1)$$

where  $u$  and  $v$  denote the zonal and meridional wind, respectively,  $f$  the Coriolis parameter, square brackets the zonal mean, and primes the departures from the zonal mean. The individual terms in (1) can be understood as follows: the lhs corresponds to the acceleration of the zonal-mean wind; the first term on the rhs the meridional convergence of zonal wave momentum; the second term on the rhs the Coriolis force acting on the meridional component of the flow; and the last term on the rhs the effect of friction in the planetary boundary layer. When integrated vertically, the Coriolis term equals zero. Hence, neglecting surface friction, the vertically averaged extratropical zonal-mean zonal wind can only be accelerated through the convergence of the eddy momentum flux. It follows that the barotropic zonal wind anomalies associated with the NH and SH annular modes, as shown in Section 2, must be driven by the first term on the rhs in (1).

The dynamics that give rise to this term are reflected in the temporal variability of the zonal flow and hence the corresponding annular mode index. In the case of the NH, temporal variability in the annular mode varies greatly between the troposphere and stratosphere. In the troposphere, the temporal variability of the NAM resembles a first-order Markov process with a decorrelation timescale of  $\sim 10$  days [*Feldstein*, 2000; 2002]. The short intrinsic timescale of the NAM in the troposphere is consistent with idealized modeling studies in which NAM-like variability is driven entirely by momentum fluxes associated with high-frequency (i.e., synoptic-scale) eddies [*Robinson*, 1991; 1994; 1996; *Yu and Hartmann*, 1993; *Lee and Feldstein*, 1996]. Since tropospheric indices of the NAM resemble a red-noise process, the prevalence of the NAM on month- to-month and season-to-season timescales may be interpreted as “climate noise” [e.g., *Leith*, 1973; *Madden*, 1976]. In other words, interannual variability in the NAM may simply reflect the aggregated effect of short-term fluctuations in the tropospheric circulation [e.g., see *Stephenson et al.*, 2000].

The intrinsic timescale of NAM-like variability in the stratosphere is considerably longer than that observed in the troposphere, on the order of several weeks. The longer

timescale of variability in the stratosphere reflects the differing processes that perturb the zonal flow about its mean state. Whereas tropospheric variability in the NAM is driven by interactions between the zonal flow and rapidly evolving baroclinic waves, stratospheric variability in the NAM is driven by relatively slower interactions between the zonal flow and planetary-scale waves.

The differing timescales of NAM-like variability in the troposphere and stratosphere are evidenced in Plate 2, which shows daily values of the NAM index calculated at levels throughout the depth of the troposphere and lower stratosphere during the northern winter of 1998/1999 (the details of the calculation are discussed in Section 5). In the troposphere, the NAM clearly exhibits variability on submonthly timescales, with large amplitude changes in the NAM evident from one week to the next. In the stratosphere, large amplitude anomalies in the NAM tend to persist for several weeks.

The above discussion suggests that the NAM may be viewed as comprising two phenomena - a tropospheric component and a stratospheric component. That the tropospheric NAM would exist in the absence of the stratospheric NAM (and the ocean) is supported by the fact that vacillations in the zonal flow reminiscent of the NAM are clearly evident in a hierarchy of models for which the stratosphere is absent and for which there are no prescribed ocean dynamics [e.g., *Robinson*, 1991; 1992; 1994; 1996; *Yu and Hartmann*, 1993; *Lee and Feldstein*, 1996; *Feldstein and Lee*, 1996]. That the stratospheric NAM would exist in the absence of the tropospheric NAM is supported by the fact that annular-mode like vacillations are evident in numerical simulations of the stratosphere with no prescribed troposphere [e.g., *Holton and Mass*, 1976]. Hence, while the tropospheric and stratospheric components of the NAM are at times coupled during the NH winter (as discussed in Section 5), it is evident that the tropospheric and stratospheric NAM would exist in the absence of each other.

In this section, we focus on the dynamics that give rise to variability in the tropospheric component of the NAM on its intrinsic timescale of  $\sim 10$  days; the dynamics of the coupling between the tropospheric and stratospheric components of the NAM are discussed in Section 5. In the following discussion, we focus on two principal questions related to the tropospheric NAM: 1) what tropospheric processes give rise to NAM-like variability during the NH winter? and 2) why is variability in the tropospheric component of the NAM largest over the North Atlantic sector?

#### 4.1. Tropospheric Processes on Week-to-Week Timescales

Since the tropospheric component of the NAM is characterized by out-of-phase fluctuations of the strength of the

zonal flow along  $\sim 35^\circ$  and  $\sim 55^\circ\text{N}$  [Section 2], it follows that an understanding of the tropospheric NAM requires an understanding of the processes that drive variability in the extratropical zonal flow. In Eq. (1), we considered the relationship between the zonal flow and the eddy momentum flux convergence in the zonally averaged circulation. In the following, we consider the relationship between local wind-speed maxima (referred to as “jets”) and zonally asymmetric departures from the time mean flow (referred to as “transient eddies”).

For the purpose of this discussion, we assume that variability in the extratropical zonal flow can be divided into two broad classes [e.g., *Lee and Feldstein*, 1996; *Lorenz and Hartmann*, 2001]: jet meandering, in which the latitude of the local zonal wind maximum exhibits pronounced north-south shifts; and jet pulsation, in which the local windspeed maximum exhibits marked strengthening and weakening without shifting latitude. Both types of variability are observed in the NH, but the amplitude of each type varies strongly with longitude: in the North Atlantic sector, variability in the zonal flow is characterized by both jet pulsation and jet meandering of the North Atlantic polar-front jet; in the North Pacific sector, variability in the zonal flow is dominated by pulsation of the downstream end of the Asia-Pacific jet. Since the NAM spans both sectors, it presumably reflects both of these processes.

Whether jet meanders or jet pulsation dominates the variability of the extratropical zonal flow depends largely on the nature of the local jet. For example, thermally driven jets found at subtropical latitudes are characterized by strong zonal flow and pronounced vertical wind shear. Since the shear is greatest along the subtropical jet, one may expect the baroclinic eddies that are excited and organized by the jet to enhance its variability. However, the strength and sharpness of the jet is also accompanied by a large, but latitudinally confined meridional gradient of potential vorticity (PV). If the PV gradient is characterized by a sharp, strong peak in the meridional direction, Rossby waves propagating along the gradient are trapped meridionally [*Swanson et al.*, 1997]. As such, the pronounced PV gradient of the subtropical jet acts to hold eddies that grow along the jet, limiting their movement in the meridional direction. Hence, it follows that strong thermally driven subtropical jets, although turbulent, are characterized by relatively small meridional meanders in the zonal flow.

In contrast to thermally driven jets, eddy-driven jets are driven entirely by eddy fluxes of momentum. That eddies are capable of maintaining a jet in the absence of thermal forcing is evidenced in Eq. 1, and has been extensively documented in the literature. For example, *Panetta and Held* [1988], *Panetta* [1993], and *Lee* [1997] examined jets that

spontaneously organize in a baroclinically unstable flow. In all three studies, the models neither support a subtropical jet nor force a prescribed jet. Instead, the model flow is driven by a broad zone of uniform baroclinicity that is maintained against mixing by the eddies either by a fixed meridional temperature gradient of the basic flow [Panetta and Held, 1988; Panetta, 1993] or by relaxation of the model temperature towards radiative equilibrium [Lee, 1997]. Hence, the jets in these models are driven entirely by eddy momentum flux convergences: when the flow is perturbed, baroclinic waves spontaneously grow and drive a westerly jet through the associated meridional convergence of the westerly eddy momentum flux.

Eddy-driven jets are generally accompanied by marked jet meanders because their associated meridional PV gradients are not externally imposed: eddy-driven jets can be interpreted as the signature of decaying eddies. However, the extent to which jet meanders dominate the variability of eddy-driven jets also depends on the meridional scale of the midlatitude eddies and their associated baroclinic zone. Since the role of the baroclinic eddies (i.e., the synoptic-scale waves in the atmosphere) is to stabilize the flow, jet pulsation and jet meanders can be viewed as a competition between the removal of the baroclinicity by the eddy fluxes and the restoration of the baroclinicity by radiative relaxation. Results from an idealized two-layer beta-plane channel model suggest that jet meanders occur when the meridional size of the eddies is insufficiently large to fully remove the negative lower layer PV gradient at any given time. In this case, the eddies and their attendant momentum fluxes meander towards regions where the lower layer PV gradient is large and negative in their perpetual attempt to stabilize the flow [Lee and Feldstein, 1996]. The same set of calculations suggest that jet pulsation occurs when the baroclinic zone is comparable to the meridional scale of the eddies. Except for very limited values of the baroclinic zone width, the most dominant form of variability of the eddy-driven jet in this case is found to be the north-south meander of the zonal flow about the latitude corresponding to its climatological mean [Lee and Feldstein, 1996].

The above discussion suggests that the variability of eddy-driven jets is generally larger than that of subtropical jets. It also suggests that the variability of eddy-driven jets is largest when the size of the baroclinic zone exceeds the size of the eddies, hence permitting meanders in the extratropical zonal flow. Subsequently, to the extent that the NAM is driven by variability in eddy momentum fluxes, one expects the amplitude of the NAM pattern to be largest in regions where the eddy-driven jet is most prominent and the subtropical jet is weakest, and vice versa. In the following section, we draw on this argument to provide an explanation for the zonally varying structure of the NAM.

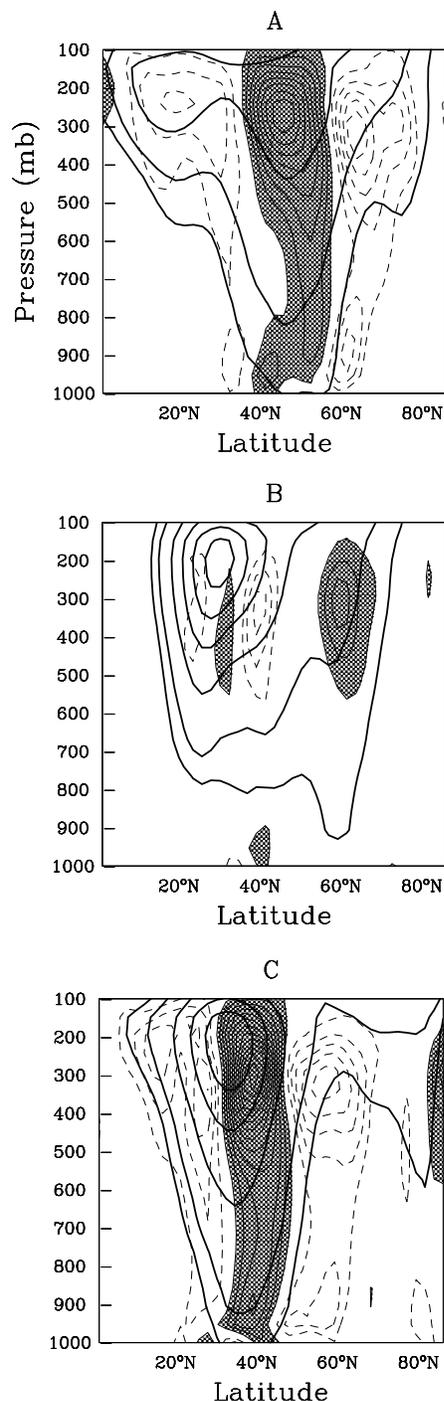
#### 4.2. Why is Variability in the NAM Most Pronounced Over the North Atlantic Sector?

Idealized model experiments suggest that, for a given sector of the Northern Hemisphere, the strength and variability of the local eddy-driven jet is intrinsically linked to the strength and latitude of the corresponding subtropical jet [Lee and Kim, 2002]. Lee and Kim's results suggest that when the subtropical jet is sufficiently strong, it tends to organize a large fraction of the baroclinic eddies, hence inhibiting the growth of eddy-driven flow at higher latitudes. Conversely, when the subtropical jet is weak, the baroclinic eddies tend to organize themselves in the poleward baroclinic zone, hence driving a purely eddy-driven jet.

While the results of such idealized model experiments should be interpreted with caution, the findings in Lee and Kim [2002] are consistent with several key features of the climatological flow. Plate 3 shows the climatological December-March mean 300-mb zonal flow (top) and the divergent component of the 300-hPa meridional wind (bottom) averaged over the 1958-97 period, and Figure 9 shows the zonal wind and high-pass (periods less than 10 days) eddy momentum flux convergence (i.e., analogous to the first term on the rhs of Eq. 1) averaged over longitude bands corresponding to the Atlantic, Asian, and Pacific sectors. The climatology of the Atlantic sector is clearly dominated by two jets (Plate 3 (top); Figure 9a). The jet along 20°N corresponds to the thermally driven subtropical jet: it is driven by a thermally direct meridional overturning cell (as revealed by the poleward flow on its equatorward side (Plate 3 (bottom)), and is largely restricted to the upper troposphere (Figure 9a). In contrast, the jet along 55°N corresponds to the eddy-driven jet: it is driven by eddy momentum flux convergence (Figure 9a) and extends throughout the depth of the troposphere. (That eddy driven jets extend throughout the depth of the troposphere is illustrated in Eq. 1. Since the lhs is zero in the steady-state and the second term on the rhs is zero when integrated vertically, it follows that the vertically integrated eddy momentum forcing can only be balanced by frictional dissipation at the surface.)

A much weaker eddy-driven jet is evident in the NH winter over the Asian continent, where the subtropical jet is very strong (Figure 9b). The Asian sector eddy-driven jet is not discernible in the upper tropospheric zonally varying zonal wind (Plate 3 (top)), but it is weakly evident in the zonally averaged wind and the eddy momentum flux convergence in this region (Figure 9b). The secondary zonal wind maximum at 60°N is much weaker than the subtropical jet at 30°N, but it exhibits a much deeper vertical structure.

The jet centered at ~40°N in the Pacific sector (Figure 9c) is also an eddy-driven jet, as evidenced by the correspon-



**Figure 9.** The zonal winds (thick lines) and high-pass eddy momentum flux convergence (thin lines) averaged over December-January-February, 1958-1997, between (a) 60°W and the Greenwich meridian, (b) 60°E-140°E, and (c) 170°E-150°W. Contour interval is 10 m s<sup>-1</sup> for the zonal wind. Contour interval for the eddy momentum flux convergence is 10-6 m s<sup>-2</sup>, shading denotes values exceeding 10-6 m s<sup>-2</sup>. Adapted from Lee and Kim [2002].

ding eddy momentum flux convergence along this latitude. Since the central Pacific is located immediately downstream of the strong Africa-Asia subtropical jet, the baroclinic eddies in this region are likely to be organized by the upstream subtropical jet [Branstator, 2002]. Studies of atmospheric midlatitude stormtracks have shown strong evidence that this is indeed the case [e.g., Chang *et al.*, 2002]. Presumably this explains the fact that this particular eddy-driven jet is characterized not by meridional meanders, but by a pulsation and lengthening/shortening of the jet [Schubert and Park, 1991].

The results in Plate 3 and Figure 9 serve to underscore the relationship between the NAM and the background climatology. In the Atlantic sector, where the climatological subtropical jet is relatively weak, the NAM corresponds to a mixture of strengthening-weakening and north-south movement of the eddy-driven midlatitude jet. In the Asian-Pacific sector, where the climatological meridional PV gradient across the upstream subtropical jet is large and sharp, the NAM corresponds to pulsation of both the subtropical jet (Figure 9b; Plate 3, longitude band B) and the downstream eddy-driven jet (Figure 9c; Plate 3, longitude band C). A close inspection of the signature of the NAM in the Asian sector (see the left-most panel in Figure 5) indicates that the pulsation is much weaker for the Asian sector subtropical jet, despite the fact that this jet is stronger than the downstream eddy-driven jet.

The above discussion suggests that the distortion of the NAM towards the Atlantic sector is intrinsically linked to the zonally varying climatological strength of the subtropical and eddy-driven jets. Idealized numerical model experiments and observations collectively suggest that the North Atlantic sector dominates the NAM because the subtropical jet is weaker in this sector, hence allowing relatively strong eddy activity to form in the North Atlantic. That eddy activity in the North Atlantic sector spans a broad range of latitudes is consistent with the fact that this sector is marked by warm lower boundary conditions that extend to relatively high latitudes.

The results presented above suggest that an understanding of the time-mean state is needed for studying fluctuations about the time-mean state. They also suggest that the convergence of wave momentum flux occurs locally, and that the frequency and/or strength of the wave driving is likely to be greatest over the North Atlantic where the subtropical jet is weak. The results raise two possible explanations for the structure of both the SH and NH annular modes. It is possible that the structure of the annular modes is a statistical artifact in that it represents an “average” of local dipoles that arise from randomly occurring, locally confined, wave-mean flow interactions. This hypothesis is consistent with

the fact that baroclinic waves tend to be organized into localized wave packets rather than a zonally uniform wave-train [Lee and Held, 1993; Chang, 1993]. On the other hand, it is also possible that a positive feedback between the zonal flow and the eddies [see Section 4.4] or zonal wave dispersion (e.g., Branstator, 2002; see also Section 6.2) acts to organize localized eddy forcing into hemispheric scales. Simultaneous wave-mean flow interaction on the hemispheric scale has been observed in the analyzed SH data [Randel and Stanford, 1985] and in the analyzed NH data [Randel, 1990].

#### 4.3 Comments on the Role of Stationary Waves

The results presented so far have focused exclusively on the role of transient eddy fluxes in driving variability in the NAM. We have yet to address recent findings that suggest stationary eddies also play an important role. For example, Feldstein and Lee [1998] and DeWeaver and Nigam [2000a, b] both found that the eddy momentum flux due to the interaction between climatological stationary eddies and anomalous transient eddies contributes toward the driving of the anomalous zonal mean flow in the NH winter. In a separate analysis, Limpasuvan and Hartmann [2000] found that interactions between the eddy and zonal mean component of the monthly-mean flow contribute to variability in the NAM. Hence, while the argument in this section suggests that transient ( $\sim 10$  day) eddy fluxes alone drive variability in the NAM, the results of these previous studies suggest that the zonally asymmetric climatological mean and monthly mean flow also play a key role.

There is at least one possible explanation for this apparent inconsistency. The time-mean midlatitude jet over the Atlantic is primarily driven by transient eddy fluxes (Figure 9a). Hence, by construction, the transient eddies must play an important role in forcing the stationary waves in the Atlantic sector [Hurrell, 1995b]. It follows that the forcing of the NAM by stationary waves may ultimately reflect the forcing due to the transients. We speculate that the role of stationary waves in driving the NAM may simply reflect the mathematical constraints that result from the decomposition of the flow into a zonal mean, a time mean, and deviation from these averages.

#### 4.4 Comments on eddy-zonal flow feedbacks

Throughout this section, we have focused on the role of eddy momentum fluxes in driving zonal flow anomalies associated with the NAM. A number of studies suggest that such zonal flow anomalies, in turn, act to organize eddy activity so that it reinforces anomalies in the zonal flow,

hence prolonging variability in the annular modes against frictional dissipation. For examples: Karoly [1990], Hartmann [1995], and Hartmann and Lo [1998] have argued that a positive feedback between the eddies and the zonal flow is evident in association with the SH annular mode; Robinson [1991] and Yu and Hartmann [1993] have argued that a similar mechanism is evident in idealized simulations of the zonal index; and Lorenz and Hartmann [2001, 2002] have noted that the positive feedback between the zonal flow and baroclinic waves acts to enhance the variance of both the SH and NH annular modes by roughly  $\sim 70\%$  on month-to-month timescales. Lorenz and Hartmann [2001, 2002] further suggest that positive feedbacks between the zonal flow and the eddies play a key role in the selection of the annular modes as the leading modes of variability of their respective midlatitude circulations.

The existence of a positive feedback between the zonal flow and the eddies is not supported by all studies of annular-mode like behavior [e.g., Lee and Feldstein 1996; Feldstein and Lee 1996]. Recent studies that do support the existence of such a feedback suggest that it is primarily restricted to the interactions between the zonal flow and high frequency transient eddies [Feldstein and Lee 1998; Lorenz and Hartmann 2001, 2002], with possible contributions from the stationary waves in the NH [Lorenz and Hartmann 2002]. The mechanism whereby high frequency transient eddies provide a positive feedback on the zonal flow anomalies is envisioned by Robinson [2000] and Lorenz and Hartmann [2001, 2002] as follows: 1) westerly anomalies in the upper tropospheric zonal flow induced by the anomalous convergence of eddy momentum flux overlies an enhanced meridional temperature gradient, as required by thermal wind; 2) the enhanced meridional temperature gradient, in turn, leads to increased baroclinic wave activity along the latitude of the strongest westerly anomalies; 3) the propagation of baroclinic wave activity away from the latitude of the strongest westerly anomalies acts to flux momentum into the core of the westerlies, hence acting to maintain them against frictional dissipation.

There are at least two caveats to the mechanism proposed by Robinson [2000] and Lorenz and Hartmann [2001, 2002]. First, the convergence of eddy momentum flux in the upper troposphere drives thermally indirect mean meridional motions that act to weaken the vertical shear. Hence, it is unclear whether the convergence of eddy momentum flux in the upper troposphere yields a meridional temperature gradient that is sufficiently large to give rise to the feedback envisioned in step 2). Second, step 3) assumes that the enhancement of the meridional temperature gradient by the induced mean meridional motions overwhelms the effect of the baroclinic wave heat flux in the lower troposphere.

Hence, the momentum flux convergence must be relatively strong, and may be expected to drive a sharper jet. In that case, the barotropic governor effect [James 1987; James and Gray 1986], which acts to suppress baroclinic growth, may not be negligible. Further research is needed to resolve these issues.

## 5. STRATOSPHERE/TROPOSPHERE COUPLING

In the previous section, we focused on the tropospheric dynamics of the NAM. In this section, we consider the coupling between tropospheric variability in the NAM and the circulation of the lower stratosphere. We also present evidence that the coupling between the tropospheric and stratospheric components of the NAM can be exploited for improved weather prediction on subseasonal and seasonal timescales. We first review the dynamics of stratosphere/troposphere coupling.

### 5.1. Stratosphere/Troposphere Coupling

Dynamic coupling between the stratospheric and tropospheric circulations occurs largely through planetary-scale Rossby waves dispersing upwards from the troposphere. These waves can propagate upwards and interact with the zonal flow in the stratosphere when the stratospheric winds are westerly, but less than a critical value [Charney and Drazin, 1961]. In the NH, these conditions are met during winter.

During the NH winter, upward propagating planetary waves disturb the strength of the stratospheric polar vortex by transporting westward angular momentum upward. In turn, the strength of the polar vortex impacts the refraction of upward propagating planetary-scale waves by modifying the background “wave guide”: a stronger polar vortex tends to be shielded from vertically propagating waves, while a weaker vortex is more susceptible to their influence. Hence, vertically propagating waves can initiate a positive feedback in which subsequent waves penetrate to successively lower altitudes as the profile of the zonally averaged wind changes [Holton and Mass, 1976; Dunkerton *et al.*, 1981; Shindell *et al.*, 2001]. Local longitudinal flow anomalies of sufficient scale may also be enough to substantially modulate refraction.

As noted in Hines [1974], the interactions between the zonal flow and upward propagating waves in the stratosphere lead to downward phase propagation of zonally averaged zonal wind anomalies within the stratosphere. The downward phase propagation of zonal wind anomalies does not imply that lower-level anomalies originate at upper levels. Rather it reveals that the stratosphere alters the conditions for planetary-wave propagation in such a way as to draw mean-flow anomalies poleward and downward [Dunkerton, 2000]. These wave-induced “vacillations” in

the zonal flow of the stratosphere are readily simulated in relatively simple models [Holton and Mass, 1976] and in general circulation models [Kodera *et al.*, 1990; Christiansen, 2001]. They are also evident in observations, where they frequently originate at levels above ~50km (at 50km, the atmospheric density is less than 0.1% of that at the surface) [e.g., Kodera *et al.*, 1990].

While the downward propagation of zonal wind anomalies in the stratosphere is clearly of theoretical interest, it was only recently that they were found to impact the circulation of the troposphere. Since this discovery was made in the context of the NAM, we first review the observed relationship between the strength of the stratospheric polar vortex and tropospheric variability in the NAM.

### 5.2. Relationship Between the Stratospheric Polar Vortex and the NAM

During the mid 1990s, a statistical relationship was recognized between month-to-month variability in the strength of the stratospheric polar vortex and a pattern resembling the NAO in the troposphere. Baldwin *et al.* [1994] and Perlwitz and Graf [1995] demonstrated that a stronger stratospheric polar vortex was associated with positive NAO-like anomalies in the troposphere, and vice versa. Baldwin *et al.* [1994] speculated that the wave-1 component of the pattern could act to force the stratosphere from below, and also suggested that part of the tropospheric pattern may represent downward influence from the stratosphere. The analysis of monthly mean data precluded any understanding of cause and effect.

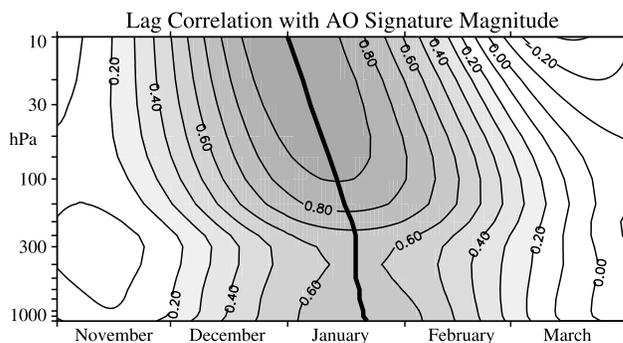
The advent of the NAM paradigm cast the coupling between the stratosphere and the troposphere as intrinsic, not to the dynamics of the wave-like structures emphasized in Perlwitz and Graf [1995] and Baldwin *et al.* [1994], but to the dynamics of the zonally symmetric polar vortex. In an effort to examine the causal linkages between zonally symmetric fluctuations in the polar vortex at stratospheric and tropospheric levels, Baldwin and Dunkerton [1999] examined the time-height development of the NAM using daily NAM-index time series calculated at 17 pressure levels from 1000 to 10-hPa (in practice, correlations between the “multi-level NAM index” and zonal-mean wind and temperature yield structures virtually identical to those presented in Section 2). Figure 10 illustrates the lag correlations between the 90-day low-pass NAM index time series at 10 hPa on January 1 (the key date for the calculation) with the NAM index at all other levels during November–March. The choice of January 1 is not critical; similar results are obtained throughout the winter. The dominant feature in Figure 10 is the downward propagation of the signature of

the NAM through the lower stratosphere into the troposphere, with lag correlations exceeding 0.65 at 1000 hPa ~3 weeks later. The thick line illustrates the peak correlation at each level. The lag correlations in Figure 10 should not be interpreted as a precise downward propagation speed; rather, they indicate an average tendency for downward propagation on a timescale of a few weeks.

The results in Figure 10 suggest that at low frequencies, the annular mode is strongly coupled in the vertical and that the phase of the patterns tends to move downward with time, often reaching Earth's surface. *Baldwin and Dunkerton* [2001] extended the analysis to 26 pressure levels from 1000 to 0.316 hPa. They defined the annular mode separately at each level as the first EOF of 90-day low-pass filtered November–April geopotential anomalies north of 20°N. In this case, time series for each level were found by projecting daily geopotential anomalies onto each level's EOF. In practice, the resulting time series are virtually identical to those used in *Baldwin and Dunkerton* [1999]. In the stratosphere, the local NAM time series are a measure of the strength of the polar vortex, similar to the zonal-mean zonal wind at 60°N; at the surface, they are identical to those used in Section 2 of this Chapter.

Plate 2 illustrates the time-height development of the NAM at daily resolution during the northern winter of 1998–1999. In the stratosphere, the time scale is relatively long: the polar vortex is warm and weak (indicated by the red shading) during middle December and late February, and cold and strong (indicated by the blue shading) through all of January and early February. Consistent with the discussion in Section 5.1, the largest anomalies in the stratosphere originate above 1-hPa (~50km), and descend through the stratosphere over a period of ~1-2 weeks. As stated in Section 4, the relatively high frequency variations in the NAM in the troposphere during 1998-1999 appear for the most part unrelated to the stratosphere. However, most winters over the past several decades (1958–1999) have descending positive and negative anomalies that appear as similarly signed anomalies in the troposphere [*Baldwin and Dunkerton*, 1999]. In general, only the strongest anomalies of either sign appear to connect to the surface, while weaker anomalies typically remain within the stratosphere. In some cases, tropospheric anomalies appear to precede stratospheric anomalies.

Following *Baldwin and Dunkerton* [2001], the average behavior of extreme events in the stratosphere can be seen by forming composites based on large negative and positive anomalies in the NAM index at 10-hPa. Since the NAM index at 10-hPa is highly correlated with the corresponding zonal-mean zonal wind along 60°N ( $r=0.95$ ), it follows that large positive values in the NAM index at 10-hPa corre-



**Figure 10.** Correlations between the 90-day low-pass annular mode values at 10 hPa on January 1 with the annular mode values at all levels during November–March. From *Baldwin and Dunkerton* [1999].

spond to a strong, well-organized vortex, while large negative values correspond to a weak, disorganized vortex (i.e., the most extreme negative values correspond to major stratospheric warmings; *Gillett et al.*, 2001). Weak and strong vortex “events” are defined here as dates on which the 10-hPa annular mode index dips below -3.0 or rises above +1.5 thresholds, respectively. The thresholds yields 18 weak vortex events and 30 strong vortex events over the period 1958-99, with the highest concentration of events occurring during December–February.

Composites based on the extreme events defined above reveal that, on average, large circulation anomalies at 10 hPa descend to the lower stratosphere over a period of ~10 days, where they tend to persist for ~60 days [Plate 4]. That the anomalies in the lower stratosphere persist longer than those at 10-hPa presumably reflects the longer radiative time scale in the lower stratosphere [*Shine*, 1987]. The composites also reveal that the persistent anomalies do not stop at the tropopause level, but that they are reflected as a shift in the mean of the relatively high frequency variability that occurs at tropospheric levels. Hence, the composites in Plate 4 reveal that, on average, large amplitude anomalies at stratospheric levels are generally followed by a bias in the mean of the tropospheric variability that persists for ~60 days.

The average surface circulation anomalies during the 60-day period following the onset of weak and strong vortex events bear a remarkable resemblance to the signature of the NAM, with the largest effect on pressure gradients in the North Atlantic and Northern Europe (Figure 11). Hence, the 60-day periods following the onset of opposite signed anomalies in the stratosphere are marked by shifts in the mean value of the NAM-index (the projection of the pattern in the middle panel of Figure 1) and by definition, the mean value of the NAO-index [the daily NAO-index shown in Plate 5 is defined in *Baldwin and Dunkerton*, 2001]. The probability density functions (PDFs) of these indices for the

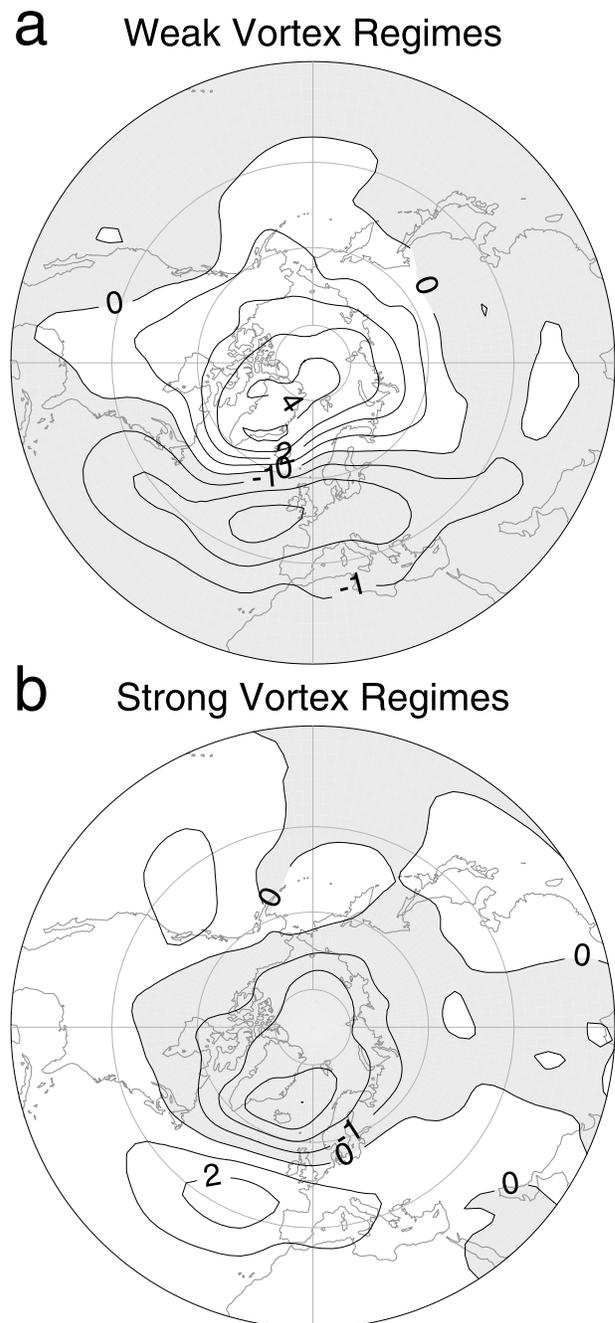
contrasting weak and strong vortex conditions are compared in Plate 5. Large amplitude anomalies in the stratosphere are followed not only by shifts in the mean of the PDFs, but also by substantial changes in the shapes of the PDFs. Values of the NAM (or NAO) indices greater than 1.0 are 3-4 times as likely following strong vortex conditions than they are following weak vortex conditions. Similarly, index values less than  $-1.0$  are 3-4 times more likely following weak vortex conditions than they are following strong vortex conditions. Since large swings in the NAM (and NAO) indices are associated with significant changes in the probabilities of weather extremes such as cold air outbreaks, snow, and high winds across Europe, Asia, and North America [Thompson and Wallace, 2001], it follows that the observed circulation changes following weak and strong vortex conditions in the stratosphere have substantial implications for prediction of Northern Hemisphere wintertime weather up to two months in advance. In the next section, we assess the direct linkage between the stratospheric polar vortex and the climate impacts of the NAM in greater detail.

### 5.3. Connection Between the Stratospheric Polar Vortex and Surface Temperature Anomalies

Following Thompson *et al.* [2002], in this section we present results that examine the direct linkage between stratospheric circulation anomalies and surface weather. The analysis technique is very similar to that used in Baldwin and Dunkerton [2001], but in this case variability in the lower stratospheric polar vortex is defined as the time series of 10-hPa geopotential height anomalies averaged  $60^{\circ}$  -  $90^{\circ}$  N, and the onset dates of weak and strong vortex conditions are defined as days when this index dips below  $-1$  standard deviations and exceeds  $+1$  standard deviations, respectively. Note that the onset dates of weak and strong vortex conditions can be assessed in real-time; they do not depend on how long the stratospheric anomalies persist.

The differences between surface temperature anomalies averaged over the 60-day periods following the onset of weak and strong stratospheric polar vortex conditions are shown in Plate 6 [left panel; see Thompson *et al.*, 2002 for details of the analysis]. The pattern of surface temperature anomalies in the left panel of Plate 6 is largely consistent with the pattern of surface temperature anomalies associated with the surface signature of the NAM [Hurrell, 1995a; Thompson and Wallace, 2000]: most of the mid-high latitude land masses tend to be anomalously cold following the onset of weak stratospheric polar vortex conditions while extreme eastern Canada and North Africa are anomalously warm (the pattern would be entirely consistent with the surface signature of the NAM if eastern Siberia and Alaska

were of the opposite sign). Densely populated regions such as eastern North America, northern Europe, and eastern Asia are  $\sim 1$ - $2$  K colder following the onset of weak vortex conditions. As noted in Thompson *et al.* [2002], the surface



**Figure 11.** Average sea level pressure anomalies (hPa) for (a) the 1080 days during weak vortex conditions and (b) the 1800 days during strong vortex conditions. From Baldwin and Dunkerton [2002].

temperature anomalies shown in the left panel of Plate 6 are roughly comparable to those associated with the contrasting phases of the El-Niño/Southern Oscillation cycle.

The 60-day interval following the onset of weak vortex conditions is also characterized by an enhanced frequency of occurrence of extreme low temperatures in most large cities that lie in NH midlatitudes [Thompson *et al.*, 2002]. Extreme cold events are roughly two times more likely during the 60-day period following weak vortex conditions than they are during the 60-day period following strong vortex conditions throughout much of North America to the east of the Rocky Mountains, northern Europe and Asia [Thompson *et al.*, 2002].

#### 5.4. Connection Between the Quasi-Biennial Oscillation and the NAM

The equatorial quasi-biennial oscillation (QBO) is a downward propagating quasi-periodic reversal in the direction of the zonal flow in the equatorial stratosphere with a mean period of ~27 months [Reed *et al.*, 1961; Baldwin *et al.*, 2001]. While the QBO is a tropical stratospheric phenomenon, it also impacts the strength and stability of the NH wintertime stratospheric polar vortex [Holton and Tan, 1980]: the easterly phase of the QBO favors a weaker stratospheric polar vortex, and vice versa. That the impact of the QBO on the extratropical circulation may extend to the surface is suggested by the fact that anomalies in the NH polar vortex frequently precede similarly signed anomalies at Earth's surface, as demonstrated in the previous section. Hence, the easterly phase of the QBO should not only favor a weaker stratospheric polar vortex, but through the linkages observed in Plates 3-4, the low index polarity of the NAM at the surface as well.

That the QBO does in fact impact the NAM is evidenced in SLP composites calculated for the opposing phases of the QBO [Holton and Tan, 1980; Baldwin *et al.*, 2001]. It is also evidenced by the recent finding that time series of the NAM and the QBO exhibit statistically significant coherence on ~27 month timescales [Coughlin and Tung, 2001]. Hence, the dynamical coupling between the troposphere and stratosphere not only has implications for the predictability of NH wintertime weather on month-to-month timescales, but on seasonal timescales as well.

The right panel in Plate 6 shows the differences in daily mean temperature between Januarys when the QBO is easterly and westerly [see Thompson *et al.*, 2002 for details of the analysis]. The amplitudes of the SAT anomalies are weaker than those obtained for the 60-day period following stratospheric anomalies [Plate 6 left panel], but the structure of the anomalies clearly bears a striking resemblance to the

signature of the NAM in surface temperature. During the easterly phase of the QBO, midwinter temperatures are lower over much of North America and northern Eurasia, and most large cities that lie in NH midlatitudes experience a greater frequency of occurrence of extreme cold events [Thompson *et al.*, 2002]. Since the phase of the QBO can be predicted several months in advance, the results in Plate 6 strongly suggest that the contrasting phases of the QBO provide a useful level of predictive skill for NH wintertime weather on seasonal timescales.

#### 5.5. Stratospheric Contribution to Predictability of the NAM

The results reviewed in this section suggest that stratospheric processes yield a useful level of predictability for the climate impacts of the NAM on timescales longer than the ~10 day limit of deterministic weather prediction. This predictability derives from three key observations: 1) NAM anomalies tend to propagate downward, as evidenced in the ~10 day time lag between stratospheric and tropospheric anomalies; 2) the timescale of the attendant surface anomalies is ~60 days, considerably longer than the timescale of internal tropospheric dynamics; and 3) the QBO impacts the strength of the extratropical zonal flow, not only in the stratosphere, but in the troposphere as well. The results imply that high frequency variability in the NAM in the troposphere is sometimes "nudged" by low frequency variability in the lower stratosphere.

The dynamics of the apparent impact of the stratosphere on the tropospheric circulation are currently under investigation. The impact could occur directly through momentum forcing of the extratropical circulation: stratospheric anomalies should induce a deep, thermally indirect mean meridional circulation below the level of the forcing that acts to transport momentum downwards [Haynes *et al.*, 1991; Hartley *et al.*, 1998; Black, 2002]. It may also occur indirectly through the effect of stratospheric circulation anomalies on the refraction of planetary waves dispersing upwards from the troposphere: westerly flow in the extratropical stratosphere favors increased equatorward propagation and anomalous poleward flux of westerly momentum in the upper troposphere/lower stratosphere, and vice versa [Hartmann *et al.*, 2000; Shindell *et al.*, 2001].

Despite the skill evidenced in this section, stratosphere/tropospheric coupling has yet to be applied in operational numerical weather prediction. Operational forecast models at the European Centre for Medium-range Weather Forecasts (ECMWF) and at the Meteorological Research Institute in Japan both have well-resolved stratospheres and include adequate representations of the relevant stratospheric dynamics. As such, these models presumably capture the

dynamics of downward-propagating zonal wind anomalies. In principle, an ensemble of forecasts run out to 60-90 days with slightly perturbed initial conditions should yield results similar to those in the composites in Plate 4. If the forecast model is capable of simulating the observed downward propagation of NAM anomalies, the model surface NAM should be nudged towards the same sign as those in the stratosphere.

Forecasts which include stratospheric information have the potential of benefiting society in a manner similar to the benefits derived from ENSO forecasts. However, forecasts based on stratospheric information will differ from those based on ENSO in three principal ways. First, since NH stratosphere/troposphere coupling is most vigorous during the winter months, subseasonal stratospheric forecasts only apply to the NH winter season. Second, while the QBO offers some hope for predictability on seasonal timescales, it only appears to impact the surface during late December-January. Third, since the stratospheric flow changes more rapidly than ENSO, forecasts may be updated daily throughout the winter season.

In light of the research emphasis placed on forecasting the NAM via midlatitude sea-surface temperature anomalies [see *Rodwell*, this volume], we feel that the evidence outlined in this section argues for increased emphasis on the skill that derives from the dynamical coupling between the tropospheric and stratospheric circulations.

## 6. SUMMARY AND CONCLUSIONS

### 6.1. Summary

This chapter provides an overview of the state of the art of our understanding of the atmospheric processes that underlie NAO-like variability.

Section 2 documents the structure of NAO when defined on the basis of the leading EOF of the NH SLP field. The results suggest that the NAO can be interpreted as the NH analogue to the leading mode of variability in the SH circulation: both patterns are characterized by vacillations in the strength of the zonal flow with centers of action located  $\sim 55^\circ$  and  $35^\circ$  latitude, and both are marked by polar centers of action in the geopotential height field with a high degree of zonal symmetry. As such, the observations presented in Section 2 motivate recasting of the NAO as the Northern Hemisphere annular mode.

The most compelling argument in favor of abandoning the perspective that the dynamics of the NAM are restricted to the North Atlantic sector is its striking similarity to the leading mode of variability in the SH, the so-called SAM. Nevertheless, whether the mode in question is viewed as a

statistical artifact of locally occurring stormtrack dynamics (the regional perspective) or as a physical phenomenon that organizes climate variability on a hemispheric scale (the annular mode perspective) remains open to debate.

Regardless of which perspective one subscribes to, the mode in question clearly has a pronounced signature in climate variability throughout much of the NH. For example, fluctuations in the NAM are strongly coupled to variability in the strength of the wintertime stratospheric polar vortex: a colder and stronger stratospheric polar vortex is associated with anomalously strong tropospheric westerlies along  $\sim 55^\circ\text{N}$ , and vice versa. Fluctuations in the NAM are also coupled with the circulation of the tropics: the high index polarity of the NAM is marked by stronger than normal trade winds over both the Atlantic and Pacific sectors, low temperatures throughout the tropical free troposphere, and weak westerly anomalies along the equator centered at  $\sim 200\text{-hPa}$ . The high index polarity of the NAM also favors positive temperature anomalies, and hence anomalously weak upwelling, in the tropical tropopause region.

Section 4 examines the tropospheric dynamics that give rise to NAM-like variability in greater detail. The section focuses on two key questions related to variability in the tropospheric component of the NAM: 1) what are the dynamical processes that determine the structure of the NAM?, and 2) why is variability in the NAM largest over the North Atlantic sector? Clearly, the solutions to these questions require an understanding of the physical processes that drive variability in the extratropical zonal flow.

Variability in localized maxima in the extratropical zonal flow (referred to as jets) can be divided into two general classes: jet meandering, and jet pulsation. Jet meandering occurs when the zonal flow exhibits marked meridional excursions about its climatological mean latitude; jet pulsation occurs when the strength of the jet varies in strength at a fixed latitude. Whether jet pulsation or jet meandering dominates the variability of the extratropical zonal flow depends on both the strength and the meridional scale of the subtropical jet, and the meridional scale of the midlatitude baroclinic zone. In general, a strong jet and/or a small baroclinic zone acts to restrict the meridional excursions of the midlatitude eddies, and hence favors jet pulsation, and vice versa.

Extratropical jets are either thermally driven (e.g., subtropical jets) or driven by the convergence of eddy momentum flux (eddy-driven jets). Eddy-driven jets are generally weaker than their thermally driven counterparts and are found in broader baroclinic zones. Hence, variability in eddy driven jets is generally characterized by meanders in the zonal flow. Since the forcing mechanism that drives eddy-driven jets (transient waves) exhibits more variability than the mechanism that drives subtropical jets (meridional

gradients in heating), one expects that eddy driven jets are also generally marked by greater variability than their subtropical counterparts.

In the troposphere, the NAM is characterized by variability in the eddy-driven jet. Hence, the amplitude of the NAM is largest in regions where the eddy-driven jet is most prominent and the subtropical jet is weakest. Observations and results from idealized model studies suggest that, over a given sector of the hemisphere, the strength of the local eddy-driven zonal flow is an inverse function of the strength of the local subtropical zonal flow. Over the Pacific sector, the thermally driven zonal flow is very strong and the eddy-driven zonal flow is relatively weak; over the Atlantic sector, the subtropical zonal flow is weaker than its Pacific counterpart, while the eddy-driven zonal-flow in the North Atlantic is relatively strong. Hence, the results in Section 4 suggest that the observed distortion of the NAM reflects the zonally varying climatological strength of the subtropical and eddy-driven jets. It is suggested that variability in the NAM is most pronounced over the North Atlantic sector because the subtropical zonal flow is weakest and the eddy-driven zonal flow is strongest in that region. It is also noted that the presence of warm lower boundary conditions at sub-polar latitudes in the North Atlantic sector should permit eddy activity over a relatively broad range of latitudes there.

In the last section in this chapter we examined the relationship between stratosphere/troposphere coupling and temporal variability in the NAM. While most studies of predictability of the NAM emphasize atmosphere/ocean coupling on decadal timescales [e.g., see *Rodwell*, this volume], the results presented in this section suggest that the dynamical coupling with the stratosphere yields a significant level of predictability on both subseasonal and winter-to-winter timescales.

Variability in the circulation of the NH lower stratosphere is driven by waves dispersing upwards from the troposphere. Since only ~25% of the mass of the extratropical atmosphere lies above the tropopause, it has generally been assumed that wave-induced variability in the strength of the stratospheric polar vortex has little impact on the circulation of the troposphere. At least two key pieces of observational evidence outlined in Section 5 suggest otherwise: 1) large amplitude anomalies in the strength of the zonal flow along ~60°N frequently originate in the middle stratosphere and descend into the troposphere. Lag correlations between the circulation at ~10-hPa and the surface reveal that variability in the strength of the lower stratospheric polar vortex leads similar signed variability in the troposphere by ~1-2 weeks; 2) the downward propagating stratospheric circulation anomalies appear to modulate relatively high frequency tropospheric variability for periods up to ~60 days following

the initiation of the stratospheric signal. Since the ~60 day timescale far exceeds the ~10 day timescale of extratropical tropospheric variability, the results in Section 5 may be interpreted as reflecting the impact of the lower stratosphere on the tropospheric circulation.

The coupling between the stratosphere and troposphere also has implications for predictability of the NAM on winter-to-winter timescales. In this case, the predictability derives from the impact of the QBO in the equatorial stratosphere on the strength and stability of the extratropical polar vortex. For example, the easterly phase of the QBO is associated with a weaker and more disturbed extratropical polar vortex and, through the linkages described above, weaker zonal flow in the troposphere consistent with the low index polarity of the NAM.

The observed linkages between long-lived anomalies in the stratospheric circulation and the surface signature of the NAM are not only of theoretical interest, but are of practical interest as well. For example, the 60 day period following weakenings in the strength of the stratospheric polar vortex are marked by lower temperatures and substantial increases in the frequency of extreme cold events throughout much of the NH. The connection between stratospheric and tropospheric circulations yields a level of predictability for NH weather that is comparable to that observed in association with the El-Nino/Southern Oscillation phenomenon.

## 6.2. Theoretical Considerations

The debate over whether the mode in question is more accurately described as a “Northern Hemisphere annular mode” or as the “North Atlantic Oscillation” attests to a key shortcoming in our understanding of the NAM, namely the absence of a unique theory for its existence in the first place. The discussion in this chapter highlights what we do understand in this regard.

In the troposphere, NAM-like variability is driven primarily by the meridional convergence of zonal momentum of baroclinic waves; in the stratosphere, it is driven by the meridional convergence of zonal momentum of waves with zonal wavenumbers 1-2. That the NAM is driven by wave-mean flow interactions is suggested by the fact that its centers of action in zonal wind are located on the poleward and equatorward flanks of the latitude band where the climatological mean eddy fluxes are largest. In Section 4, we argued that this latitude band corresponds to the region where the midlatitude eddies exhibit pronounced meridional meanders. On the basis of results from an idealized model experiment, we further suggested that tropospheric eddies exhibit pronounced meridional meanders in regions where their meridional scale is less than the meridional scale of the

baroclinic zone. Hence, we conclude that the characteristics of annular-mode like variability are determined in large part by the meridional scale of the eddies.

What determines the relevant length scale of the eddies? In the context of linear baroclinic instability, the relevant length scale is the Rossby radius of deformation. However, since the energy associated with individual eddies cascades to larger spatial scales in a quasi-two dimensional circulation [Kraichnan, 1967], a more appropriate length scale is given by Rhines [1975]. This spatial scale, commonly referred to as the Rhines scale, corresponds to the length scale at which the upscale energy cascade is balanced by Rossby wave radiation, and is determined by the meridional gradient of planetary vorticity and the root-mean-square velocity of the flow. But the Rhines scale itself may also not be entirely appropriate, as it only holds for an inviscid flow contained in a sufficiently large domain [Held, 1999]. In the presence of friction, the inverse energy cascade may be halted before it ever reaches the Rhines scale. In the case where the Rhines scale is greater than the size of Earth, the length scale of the energy containing eddies will be that of Earth itself.

While our understanding of the fundamental dynamics of the NAM is incomplete, the above discussion highlights two important conclusions: 1) annular modes are constrained by the eddy length scale and hence are governed by fundamental quantities such as the radius of Earth, its rotation rate, and stratification; and 2) the physical process of the NAM are rooted in the dynamics of large-scale turbulence. Still, the above discussion does not necessarily provide a theory for the possible existence of coordinated variability on a hemispheric scale. That is, the above discussion does not provide proof that the NAM is a coherent physical mode that organizes climate variability throughout the hemisphere rather than a statistical artifact of locally occurring wave-mean flow interactions. It is possible that the NAM reflects the organization of two-dimensional turbulence into zonal jets, which occurs in cases where  $\beta$  is large [Rhines, 1975; Williams, 1978; Pedlosky, 1987]. It is also possible that the NAM reflects the organization of eddy activity by anomalies in the zonal flow [e.g., Lorenz and Hartmann, 2002]. However, whether the radiative forcing that continuously energizes the atmosphere allows enough time for these processes to take place remains unclear.

### 6.3. Concluding Remarks

As noted above, the absence of a unique theory for the existence of the NAM constitutes a key shortcoming in our understanding of extratropical climate variability. Another key shortcoming regards the dynamical coupling between the stratospheric component of the NAM and the circulation

of the troposphere. The results in this chapter clearly suggest that an improved understanding of this coupling is of practical use for weather prediction. Several theories have been proposed to explain how stratospheric anomalies can impact the circulation of the troposphere, and several models are capable of simulating the observed linkages. Nevertheless, our understanding of the dynamics of the coupling remains incomplete.

The NAM has played an important role in recent climate change [Hurrell, 1995a; 1996; Thompson *et al.*, 2001; Gillett *et al.*, this volume], and similar trends have been observed in the Southern Hemisphere [Hurrell and van Loon, 1994; Meehl *et al.*, 1998; Thompson and Solomon, 2002]. Recent research suggests that both annular modes are sensitive to a wide array of forcing mechanisms, including increasing greenhouse gases [Shindell *et al.*, 1999; 2001; Fyfe *et al.*, 1999; Kushner *et al.*, 2001], feedbacks between greenhouse gases and ozone depletion [Hartmann *et al.*, 2000], increases in tropical sea-surface temperatures [Hoerling *et al.*, 2001], and variations in solar forcing [Shindell *et al.*, 2001b]. Nevertheless, it is unlikely that the source(s) of the observed trends in the annular modes can be unequivocally isolated in the absence of a consensus regarding the atmospheric processes that give rise to annular variability in the first place. In our view, establishing a theory for the existence of annular variability is of paramount importance for future research.

*Acknowledgments.* Thanks to W. J. Randel, W. A. Robinson, J. W. Hurrell, and one anonymous reviewer for their helpful comments and suggestions. Thanks also to T. J. Dunkerton, S. Feldstein, H.-K. Kim, D. J. Lorenz, and J. M. Wallace for their assistance and insight at various stages of this research, and to S.-W. Son for generating Plate 3. DWJT is supported by the National Science Foundation under grant CAREER: ATM-0132190. SL is supported by NSF grant ATM-0001473. MPB is supported by NSF grant ATM-0002485, NASA's SR&T Program for Geospace Science, contract NASW-00018, and NOAA's OGP CLIVAR Atlantic grant NOAA-0572.

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