A mechanistic model of the northern annular mode

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The northern annular mode (NAM) is the leading mode of variability in the Northern Hemisphere. The NAM consists of a meridional dipole of the zonal mean zonal wind that extends into the stratosphere during the winter season. During the high-index state of the NAM, high-latitude westerly winds are stronger than normal, leading to a stronger and colder polar vortex. Over the past few decades, observations show a trend toward higher NAM index values. To investigate whether this trend is a result of natural variability or is related to greenhouse warming, a mechanistic model is developed that provides a simple analogue of the NAM. The model represents the interaction between the zonally symmetric flow and planetary waves for two meridional modes of variation. Greenhouse warming is specified in the model by varying the radiative equilibrium meridional temperature gradient. Variations in the second meridional mode qualitatively resemble those of the observed NAM. Model runs that correspond to larger greenhouse gas concentrations yield on average higher NAM index values.

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1. Introduction

The North Atlantic Oscillation (NAO), first identified by Walker and Bliss [1932], is a well known mode of climate variability. The NAO is a measure of the meridional sea level pressure (SLP) gradient in the North Atlantic; thus, it is related to the strength of the westerly winds blowing across the North Atlantic basin. Temperature and precipitation patterns correlated with the NAO throughout Europe are a result of these oscillating westerly winds [Hurrell, 1995]. van Loon and Rogers [1978] showed a significant relationship in SLP between the NAO and the North Pacific suggesting that the effects of the NAO were farther reaching than previously thought. Additional studies by Baldwin et al. [1994] and Perlwitz and Graf [1995] have demonstrated a correlation between geopotential heights at 50 hPa, which are a measure of the strength of the polar vortex, and a 500 hPa pattern resembling the NAO.

Thompson and Wallace [1998] performed an empirical orthogonal function (EOF) analysis of Northern Hemisphere (poleward of 20°) wintertime (November–April) monthly mean SLP anomalies and found the first EOF, i.e., the leading mode of variability, to be an annular mode. They named the annular mode, which consists of a seesaw of mass between the Arctic region and surrounding latitudes, the Arctic Oscillation (AO). The AO is also know as the northern annular mode (NAM) and will be referred to as such henceforward. (The associated principle component time series is called the NAM index.) The NAM is very similar to the NAO over the North Atlantic region, and the NAO can be thought of as a regionalization of the NAM [Wallace, 2000]. The main differences of the NAM include its more zonally symmetric spatial pattern and its strong coherence between the troposphere and stratosphere. Regression maps of geopotential data onto the NAM reveal that the mode extends from the surface into the lower stratosphere [Thompson and Wallace, 1998].

When there is anomalous low SLP over the pole and anomalous high SLP values in the surrounding zonal ring, the NAM is said to be in a high-index state. The high-index state corresponds to a stronger polar vortex, stronger westerly winds near 60° in the troposphere, warmer temperatures at the surface over northern Eurasia, and an increase in precipitation over northern Europe. Surface temperature increases are due to the stronger westerly winds blowing warm oceanic air over the continent [Thompson and Wallace, 2000], while the increased precipitation is a result of a northerly shifted storm track [Hurrell, 1995]. The reverse conditions are applicable for low values of the NAM index.

While additional research has shown that the NAM exists throughout the year [Baldwin and Dunkerton, 1999], the amplitude is largest during the winter months when coupling between the troposphere and stratosphere is most pronounced. In addition, Baldwin and Dunkerton [1999] found a very robust correlation between stratospheric sudden warming events and low NAM index values. Every major warming from 1958 to 1997 corresponded to a large negative deviation of the NAM index. These low NAM index values first appear in the stratosphere and then propagate down into the troposphere over a timescale of a few weeks.
The coupling between the troposphere and stratosphere of the NAM suggests a connection via planetary wave propagation. Limpasuvan and Hartmann [2000] reported such a link in their analysis of the NAM. They found that during the high NAM phase, planetary waves were refracted more equatorward than during the low phase. The anomalous poleward fluxes of zonal momentum associated with equatorward propagating stationary waves help maintain the stronger westerly winds in the upper troposphere during the high-index phase of the NAM. In addition, the stronger winds in the upper troposphere allow fewer waves to propagate into the stratosphere; thus, the polar vortex is less disturbed and remains strong.

Over the past few decades a trend toward higher values of the NAM index has been observed [Hurrell, 1995; Thompson and Wallace, 1998; Thompson et al., 2000]. This trend is reflected in the lower stratosphere through decreasing polar temperatures and lower geopotential heights which corresponds to a strengthening of the polar vortex. While at the surface, SLP values have been falling over the Arctic region, temperatures have been increasing over most high-latitude continents, and precipitation rates have increased region, temperatures have been increasing over most high-latitude continents, and precipitation rates have increased over northern Europe. These conditions are consistent with latitude continents, and precipitation rates have increased over northern Europe. These conditions are consistent with

Eurasia has been associated with global warming, it is beneficial to see if this trend is anthropogenic in origin or just a result of innate climate variability. In an attempt to answer this question, several studies have tried to reproduce the observed trend by forcing a global circulation model (GCM) with increased greenhouse gas (GHG) concentrations [Graf et al., 1995; Fyfe et al., 1999; Shindell et al., 2001]. All three studies were able to reproduce a positive trend in the NAM. Shindell et al. [2001] attributed their success to the use of a GCM with a fully resolved stratosphere, but Fyfe et al. [1999] found that the inclusion of a detailed stratosphere was unnecessary. Despite these promising model results attributing the recent trend of the NAM to GHG forcing, causal relationships are difficult to obtain from GCM simulations due to their inherent complexity.

In an effort to better clarify the effects of greenhouse gas forcing on the planetary waves that drive the NAM and to further investigate the role of the stratosphere, a simple mechanistic model is developed. The studies cited above indicate that a minimal model of the NAM must include vertical propagation of planetary waves and allow for the meridional transport of zonal momentum. These characteristics are incorporated into the current model, and results reveal that the model is able to produce an oscillation of the zonal mean zonal wind that is qualitatively similar to observations.

Last it is noted that an analogous structure to the NAM has been identified in the Southern Hemisphere [Hartmann and Lo, 1998]. Referred to as the Southern Annular Mode (SAM), it represents the leading mode of variability in the Southern Hemisphere. Although the NAM and SAM are quite similar, the eddy forcing of the annular modes is partitioned differently between the hemispheres. The eddy forcing of the NAM is principally due to stationary waves; however, transient waves primarily contribute to the SAM [Hartmann and Lo, 1998; Limpasuvan and Hartmann, 2000]. Since only stationary waves are incorporated in the numerical model, results are only applicable to the NAM. A detailed description of the numerical model is provided in section 2. Results and conclusions are presented in section 3.

2. Model Description

The numerical model developed and used in this study is an extension of the quasi-geostrophic β-plane channel model of Holton and Mass [1976], hereafter HM76. The model domain is centered at 45°N with a channel width of 60° and extends from the surface to 90 km. Fourier expansions are used in the zonal and meridional directions. Finite differencing is applied in the z-direction with 2-km vertical grid spacing. The model is integrated using a third-order Adams-Bashforth method with a 1-hour time step.

2.1. Governing Equations

The governing equation for the zonal-mean zonal wind written using standard notation [e.g., Andrews et al., 1987] in log-pressure coordinates is

\[
\frac{\partial}{\partial \tau} \left( \frac{\partial^2 u}{\partial y^2} + \frac{f_0^2}{\rho N^2} \frac{\partial^2 \psi}{\partial z^2} \right) - \frac{\partial^2 \psi}{\partial y^2} \cdot \left( \frac{\partial \psi}{\partial \tau} \right) = \frac{\partial^2 \psi}{\partial y^2} \cdot \left( \frac{\partial \psi}{\partial \tau} \right)
\]

where \( q' = \nabla^2 \psi' + \frac{f_0^2}{\rho N^2} \frac{\partial^2 \psi'}{\partial z^2} \)

is the quasi-geostrophic perturbation potential vorticity with \( \psi' = \Phi \psi' \) being the geostrophic stream function. \( \gamma = \gamma(z) \) is the Rayleigh friction rate, and \( \alpha = \alpha(z) \) is the Newtonian cooling coefficient; \( u_R \) is the radiative equilibrium zonal wind, which is related to the radiative equilibrium temperature, \( T_R \), through the thermal wind relation

\[
f_0 \frac{\partial u_R}{\partial z} = - \frac{R}{H} \frac{\partial T_R}{\partial \phi}.
\]

Examining equation (1), it is apparent that the mean flow is driven by three different processes: Rayleigh friction, which damps the mean flow in the mesosphere and represents a simple parameterization of gravity wave drag; radiative forcing, which relaxes the mean flow to a radiative equilibrium state; and the eddy flux of potential vorticity, which consists of the eddy momentum and eddy heat fluxes that drive the flow away from radiative equilibrium. As shown by Andrews et al. [1987] the eddy potential vorticity flux is related to the divergence of the Eliassen-Palm (EP) flux

\[
\left( \nabla q' \right) = - \frac{1}{\rho} \frac{\partial}{\partial \phi} \left( u_R \right) + \frac{f_0^2}{\rho N^2} \frac{\partial}{\partial z} \left( \rho \frac{\partial \psi'}{\partial z} \right) = \frac{1}{\rho} \nabla \cdot F.
\]
The wave fields in the model are governed by the quasi-geostrophic perturbation potential vorticity equation

$$
\left( \frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x} \right) \psi + \frac{\partial \bar{q}}{\partial y} \frac{\partial \psi'}{\partial x} + \frac{1}{\rho} \frac{\partial^2 \psi}{\partial z^2} \frac{\partial}{\partial z} \left( \rho \frac{\partial \psi'}{\partial z} \right) = 0,
$$

(2)

where

$$
\bar{q} = \beta - \frac{\partial^2 \bar{u}}{\partial y^2} - \frac{1}{\rho} \frac{\partial^2 \bar{u}}{\partial z^2} \left( \frac{\partial \bar{u}}{\partial z} \right)
$$

is the meridional gradient of the zonal mean quasi-geostrophic potential vorticity.

HM76 represented the meridional dependence of both the zonal wind and wave fields in terms of a Fourier sine series. In their study only the first mode was retained. We use the same approach, but include the first two modes of the Fourier expansion. As in HM76, only a single zonal harmonic wave mode is retained;

\begin{align}
\bar{u}(y,z,t) &= U(z,t) \sin ly + UC(z,t) \sin 2ly, \\
u_R(y,z) &= U_{R(z)} \sin ly + U_{R(z)} \sin 2ly,
\end{align}

(3a)

(3b)

\begin{equation}
\psi'(x,y,z,t) = \Re \left[ \Psi_R(z,t) e^{i ky} \right] e^{i/2H} \sin ly + \Re \left[ \Psi_M(z,t) e^{i ky} \right] e^{i/2H} \sin 2ly.
\end{equation}

(3c)

Retention of both the first symmetric and asymmetric mode is required for incorporating meridional transport of zonal momentum in the model. Substituting the above solutions into equations (1) and (2) yields quadratic terms in \(y\) which are then expanded in a Fourier sine series as

\begin{align}
sin^2 ly &\approx \epsilon_1 \sin ly + \ldots \\
2 \sin ly \sin ly &\approx \epsilon_2 \sin 2ly + \ldots \\
5 \sin 2ly &\approx \epsilon_2 \sin ly + \ldots
\end{align}

(4a)

(4b)

(4c)

with \(\epsilon_1 = 8/3 \pi\) and \(\epsilon_2 = 32/15 \pi\). Since only projections onto \(\sin ly\) and \(\sin 2ly\) are kept, this substitution results in separate prognostic equations for \(U_A\), \(U_C\), \(\Psi_R\), and \(\Psi_M\).

A very similar modification of the HM76 model was used by Kodera and Kuroda [2000] to investigate coupled stratosphere-troposphere variability. The coupled mode of variability found in their model is related to the variability associated with the NAM, but direct comparisons between their results and the NAM are difficult. The primary reason for this is that their model is centered at 60°N. The current model is centered at 45°N, and the first meridional mode, which has a \(\sin ly\) latitudinal dependence, has a maximum at 45°N. Therefore, the first mode zonal mean zonal wind, \(U_A\), represents the strength of the polar vortex. The second meridional mode in the model has a \(\sin 2ly\) dependence in latitude, and thus depicts a meridional dipole with peaks of opposite sign at 60°N and 30°N, and a node at 45°N. The meridional structure of the second mode qualitatively resembles the observed dipole of zonal mean zonal wind correlated with the NAM [Baldwin and Dunkerton, 1999; Thompson and Wallace, 2000]; consequently, the time evolution of the second mode zonal mean zonal wind, \(UC\), represents the time evolution of the NAM in this model. Therefore, positive (negative) NAM events are associated with positive (negative) \(UC\) values at 60°N in the model. Such associations with the NAM are not possible in Kodera and Kuroda [2000]. Other limitations of their model are discussed below.

2.2. Boundary Conditions

Vanishing normal flow is assumed at the lateral boundaries of the channel; thus, \(\psi' = 0\) and \(\bar{v} = 0\) at the southern and northern boundaries, \(y = 0, L\). At the top boundary we assume that the zonal mean zonal wind shear is zero, i.e.,

$$
\frac{\partial U_A}{\partial z} = \frac{\partial U_C}{\partial z} = 0 \quad \text{at} \quad z = z_f,
$$

and apply a radiation condition on the waves

$$
\Psi_R \propto e^{imz} \quad \text{and} \quad \Psi_M \propto e^{imz} \quad \text{at} \quad z = z_f.
$$

\(m_R\) and \(m_M\) are determined by solving the prognostic equations for \(\Psi_R\) and \(\Psi_M\), respectively, at steady state conditions with \(\alpha = 0\) and \(UC = 0\).

HM76 set the lower boundary at 10 km. At the lower boundary the zonal mean zonal wind was fixed at a particular value, and the wave forcing was specified with a geopotential height perturbation. Using such a boundary condition eliminates the possibility of any troposphere-stratosphere interaction. Kodera and Kuroda [2000] extended the lower boundary to the surface, but the wave forcing was still specified through a geopotential height perturbation. The mean wind was constrained at the lower boundary by relaxing the flow toward its radiative equilibrium level, \(UR\). This type of lower boundary allows for a troposphere, but still limits the interaction between the mean wind and waves at the lower boundary. In an effort to include a more realistic troposphere, we incorporate a lower boundary at the surface where both the mean wind and wave components are internally determined. The boundary condition we use was first developed by Tung [1983] and has been used by Robinson [1985], Wakata and Uryu [1987], and Yoden [1987] in their quasi-geostrophic \(\beta\)-plane model studies.

To obtain the lower boundary condition we first assume the vertical velocity induced by flow over topography is given by Tung [1983], i.e.,

$$
w = u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} + D_e \left( \nabla^2 \psi' - \frac{\partial \bar{u}}{\partial y} \right) \quad \text{at} \quad z = 0,
$$

(5)

where \(h\) is the height of the topography, and \(D_e\) is the coefficient of eddy diffusion. Here we have assumed that the geometric lower boundary is at \(z = 0\) in large-pressure coordinates. Although Robinson [1985] did not make this approximation, Tung [1983] showed that the induced error of this assumption is small for small-wave
number, stationary waves. The vertical velocity given in equation (5) is now separated into mean and perturbation components:

\[ w = u' \frac{\partial h}{\partial x} + v' \frac{\partial h}{\partial y} - D_v \frac{\partial}{\partial y} (\bar{u} - u_b), \tag{6a} \]

\[ w' = \bar{\frac{\partial h}{\partial x}} + D_e \nabla^2 \psi'. \tag{6b} \]

The last term in equation (6a) has been modified by replacing \( \bar{u} \) with \( \bar{u} - u_b \). As explained by Yoden [1987], this provides an ad hoc method for maintaining westerslies at the surface and is “crucial for obtaining a forced-wave component.” To express equations (6) in terms of \( U_A, U_C, \Psi_K, \) and \( \Psi_M \) we expand \( h \) in a Fourier sine series as

\[ h(x, y) = \Re[H_k e^{ikx}] \sin ky + \Re[H_m e^{ikx}] \sin 2ky, \tag{7} \]

substitute equations (3), and express quadratic terms as in equations (4), to yield the following lower boundary equations at \( z = 0 \) for \( U_A, U_C, \Psi_K, \) and \( \Psi_M \):

\[ \frac{\partial}{\partial t} \left( \frac{\partial U_A}{\partial z} \right) = \frac{k^2 N^2}{f_o} \left[ \varepsilon_1 H_k \frac{\partial \Psi_K}{\partial z} + \varepsilon_2 H_M \frac{\partial \Psi_M}{\partial z} \right] \]

\[ + D_f \frac{N^2}{f_o} (U_A - U_{R_L}) - \alpha \frac{\partial}{\partial z} (U_A - U_{R_L}) \]

\[ + \frac{k^2}{2} \left[ \varepsilon_1 \left( \frac{\partial \Psi_K}{\partial z} \right)^2 + \varepsilon_2 \left( \frac{\partial \Psi_M}{\partial z} \right)^2 \right] \tag{8} \]

\[ \frac{\partial}{\partial t} \left( \frac{\partial U_C}{\partial z} \right) = 2 \varepsilon_2 k^2 \frac{N^2}{f_o} \left[ H_k \frac{\partial \Psi_K}{\partial z} + H_M \frac{\partial \Psi_M}{\partial z} \right] \]

\[ + 4 D_f \frac{N^2}{f_o} (U_C - U_{R_L}) - \alpha \frac{\partial}{\partial z} (U_C - U_{R_L}) \]

\[ + 2 \varepsilon_2 k^2 \left[ \frac{\partial \Psi^*_K}{\partial z} \frac{\partial \Psi^*_M}{\partial z} + \frac{\partial \Psi^*_M}{\partial z} \frac{\partial \Psi^*_K}{\partial z} \right] \tag{9} \]

\[ \left( \frac{\partial}{\partial t} + ik \varepsilon_1 U_A \right) \left( \frac{\partial \Psi_K}{\partial z} \right) = \frac{\varepsilon_1 \Psi_K}{2H} \frac{\partial \Psi_M}{\partial z} + \frac{\varepsilon_1 \Psi_M}{2H} \frac{\partial \Psi_K}{\partial z} \]

\[ + i k \varepsilon_2 \frac{U_C}{2H} \frac{\partial \Psi_M}{\partial z} + i \varepsilon_2 U_C \frac{\partial \Psi_M}{\partial z} + \frac{N^2}{f^2} \left[ \varepsilon_1 U_A \Psi_K + \varepsilon_2 U_C \Psi_M \right] \]

\[ + D_f \frac{N^2}{f_o} \left( k^2 + \ell^2 \right) \Psi_K \tag{10} \]

\[ \left( \frac{\partial}{\partial t} + ik \varepsilon_2 U_A \right) \left( \frac{\partial \Psi_M}{\partial z} \right) = \frac{\varepsilon_2 \Psi_M}{2H} \frac{\partial \Psi_K}{\partial z} + \frac{\varepsilon_2 \Psi_K}{2H} \frac{\partial \Psi_M}{\partial z} \]

\[ + i k \varepsilon_1 \frac{U_C}{2H} \frac{\partial \Psi_K}{\partial z} + i \varepsilon_1 U_C \frac{\partial \Psi_K}{\partial z} + \frac{N^2}{f^2} \left[ \varepsilon_2 U_A \Psi_K + \varepsilon_1 U_C \Psi_M \right] \]

\[ + D_f \frac{N^2}{f_o} \left( k^2 + \ell^2 \right) \Psi_M \tag{11} \]

2.3. Parameters

Rayleigh friction is incorporated into the model to mimic the zonal drag effect of gravity wave breaking in the mesosphere. \( \gamma \), the Rayleigh friction rate coefficient, is defined as

\[ \gamma(z) = \left[ 1.0 + \tanh \left( \frac{z - 70}{15} \right) \right] \times 5.0 \times 10^{-6} \text{s}^{-1}; \]

thus, values increases from near zero at 50 km to 0.8 days\(^{-1}\) at the upper boundary. The Newtonian cooling coefficient applied in the model is

\[ \alpha(z) = \left[ 2.0 + \tanh \left( \frac{z - 35}{7} \right) \right] \times 10^{-6} \text{s}^{-1}. \]

The radiative damping times associated with \( \alpha \) are approximately 12 days in the troposphere and lower stratosphere and decrease to 4 days in the mesosphere.

[20] The buoyancy frequency squared, \( N^2 \), is assumed constant throughout the depth of the model. Thus, there is no change in static stability between the stratosphere and troposphere. Chen and Robinson [1992] have investigated the effects of this assumption using a linear, primitive equation model. They found using constant \( N \) led to a decrease in the amount wave activity and an enhancement of vertical wave propagation when compared to using a realistic profile. Despite this, the structure of the amplitude and phase of the wave response was similar between the two runs. Thus, using constant \( N \) should not inhibit us from obtaining a realistic response in the model. Nevertheless, results from a model with a realistic profile of the buoyancy frequency would be beneficial for confirming our results.

[21] The radiative equilibrium zonal wind, \( u_{R_L} \), is composed of two modes: \( U_{R_L} \) and \( U_{R_R} \). The first mode is defined to have a constant zonal mean wind shear of 3.0 m s\(^{-1}\) km\(^{-1}\), which corresponds to a meridional temperature difference of \( \sim 40 \text{ K} \) across the channel. This is representative of Northern Hemisphere winter conditions. The second mode is used to move the stratospheric jet to a more northerly and realistic position. Figure 1 shows the resultant two-mode radiative equilibrium zonal wind. A subtropical jet structure is not incorporated.
into the radiative equilibrium zonal wind. Limpasuvan and Hartmann [2000] provides justification for this simplification, as their results suggest that the NAM is not sensitive to the subtropical jet structure. Furthermore, Kodera and Kuroda [2000] found that the strength of the subtropical jet had little impact when wave forcing was greater than 100 m.

3. Results and Conclusions

For all model results, forcing is applied only through the first mode with a zonal planetary wave number two structure; i.e., $k = \frac{2}{\pi \sin \phi}$ and $H_k = 0$ in equation (7). $H_k$ is asymptotically increased over the first 30 days, whereafter it is held constant. The radiative equilibrium shear and all other model parameters are independent of time; thus, all observed variability in the model is internally generated.

### 3.1. Control Case

The time evolution of the zonal mean zonal wind for both modes at 60°N with 125 m lower boundary forcing is displayed in Figure 2. The zonal mean zonal wind oscillates between easterly and westerly values with a period of about 60 days in the stratosphere in Figure 2. The magnitude of the oscillation is largest in the mesosphere and stratosphere with values ranging between $-40$ m s$^{-1}$ and $80$ m s$^{-1}$ in the first mode and between $-10$ m s$^{-1}$ and $15$ m s$^{-1}$ in the second mode. Little time variability exists in the first mode below 20 km, but a signal is evident in the second mode near the bottom boundary in the troposphere. Low index events propagating downward from the mesosphere into the stratosphere are quite obvious. Propagation through the lower stratosphere into the troposphere is less apparent, though. Kodera and Kuroda [2000] obtained a similar oscillation of the zonal mean zonal wind in their model.

To further investigate the downward propagation in the model, EP flux vectors and divergence for the total flow are calculated. For the quasi-geostrophic beta-plane case, the EP flux is defined as

$$ F = \left( 0, -\rho \mathbf{u} \mathbf{v}, \frac{f^2}{N^2} - \nabla \cdot \left( \frac{\partial \mathbf{u}}{\partial z} \right) \right). $$

The $y$-component is equal to the eddy momentum flux, and the $z$-component is the eddy heat flux. EP flux vectors are parallel to the direction of wave propagation in the $y-z$ plane, and the EP flux divergence represents the sole eddy forcing on the mean zonal flow (for a review, see Andrews et al. [1987]). Figure 3 shows the EP flux vectors and divergence every 15 days for one period of the oscillation. Regions of EP flux convergence, where the mean flow is

![Figure 2. Zonal mean zonal wind at 60°N for $H_k = 125$ m and 3.0 m s$^{-1}$ km$^{-1}$ first mode radiative equilibrium wind shear, $\frac{\partial u}{\partial z}$. (a) First mode, $U_A$; (b) Second mode, $U_C$. Contour interval is 15 m s$^{-1}$ in (a) and 5 m s$^{-1}$ in (b). The zero and negative contours are dashed. Shading indicates westerly winds.](image1)

![Figure 3. Latitude versus height plots of EP flux vectors and divergence for the total flow. Day indicated at the top of each plot. Contour interval is $4 \times 10^{-6}$ m s$^{-2}$ with convergence contours dashed. Zero contour omitted.](image2)
The effect of increased GHG concentrations is to warm the troposphere where GHG emit less radiation than they absorb from longwave radiation due to the positive lapse rate [Shindell et al., 2001]. In the stratosphere the lapse rate is negative so GHG emit more radiation than they receive, which causes cooling. Since the height of the tropopause decreases with latitude, GHG act to increase the meridional temperature gradient in a layer near the tropopause.

To simulate such effects in the model, integrations are performed for different values of the first mode radiative equilibrium wind shear, \( \frac{\partial w}{\partial z} \). \( U_{R_{A}} \) is in thermal wind balance with the radiative equilibrium meridional temperature gradient, so changing the shear of \( U_{R_{A}} \) changes the radiative equilibrium temperature difference across the channel. Since the increase in the equator to pole radiative equilibrium temperature gradient related to stronger GHG forcing is largest near the tropopause, the radiative equilibrium shear is adjusted only between 6 and 14 km. Results with 2.7 and 3.3 m s\(^{-1}\) km\(^{-1}\) radiative equilibrium wind shears will be shown. These shear values correspond to a five degree decrease and increase of the control case radiative equilibrium meridional temperature difference. The 2.7 and 3.3 m s\(^{-1}\) km\(^{-1}\) runs will be referred to as the weak and strong shear cases, respectively. Hartmann et al. [2000] note that decreasing ozone concentrations will also yield an increase in the radiative equilibrium meridional temperature gradient through cooling of the polar stratosphere. Since radiative forcing changes are applied solely in terms of the radiative equilibrium meridional temperature gradient, the current model is unable to distinguish between an increase of GHG concentrations or a depletion of ozone.

### 3.2. Radiative Equilibrium Shear

Over the past two centuries GHG concentrations have been increasing as a result of anthropogenic activity. The effect of increased GHG concentrations is to warm the troposphere where GHG emit less radiation than they absorb from longwave radiation due to the positive lapse rate [Shindell et al., 2001]. In the stratosphere the lapse rate is negative so GHG emit more radiation than they receive, which causes cooling. Since the height of the tropopause decreases with latitude, GHG act to increase the meridional temperature gradient in a layer near the tropopause.
response could possibly lead to interannual variability, since the model is run at winter mean conditions.

The only mechanism forcing the mean flow away from radiative balance is the EP flux of vertically propagating planetary waves. Since planetary waves act to weaken the polar vortex, the strength of the polar vortex provides a proxy for planetary wave activity in the stratosphere. The stratospheric winds are stronger for the case with increased radiative equilibrium shear and weaker for the decreased shear case. This implies that the strength of the westerly winds in the 6 km to 14 km layer greatly affects the ability of planetary waves to propagate into the stratosphere. This agrees with the modeling work of Chen and Robinson [1992] and the linear theory presented by Charney and Drazin [1961].

Figure 5. Time versus height plot of $U_A$ at 60°N for 125 m forcing and the following $U_{R_i}$ shear between 6 and 14 km: (a) 2.7 m s$^{-1}$ km$^{-1}$, (b) 3.3 m s$^{-1}$ km$^{-1}$. Contour interval is 15 m s$^{-1}$ with the zero and negative contours dashed. Shading indicates westerly winds.

[30] $U_C$ at 6 km and 60°N is shown in Figure 7 for both the weak and strong shear cases. The amplitude of the oscillation is slightly larger, and the time mean $U_C$ is more westerly, which corresponds to higher NAM index values, for the strong shear case. In addition, the time evolution of $U_C$ for the two shear cases is quite similar with both spending nearly equal time in high and low index states. This result agrees with that of Fyfe et al. [1999] who used the Canadian Centre for Climate Modeling and Analysis (CCCma) coupled climate model to study the effects of GHG forcing on the NAM. Shindell et al. [1999], using the NASA Goddard Institute for Space Sciences (GISS) atmosphere GCM, also report increased NAM index values due to GHG forcing.

[31] The reason for the stronger winds at 6 km is found by examining the y-component of the EP flux. Recall that the y
component of the EP flux, $F^y$, is equal to the eddy momentum flux and indicates the direction of planetary wave propagation in the meridional plane. Figure 8 shows that the strong shear case has larger negative values of $F^y$, which correspond to a northward flux of zonal momentum. Thus, the stronger westerly winds at 6 km and 60°N result from an increase in the eddy momentum flux that is associated with the equatorward propagation of planetary waves.

To illustrate that the second mode zonal mean zonal wind is stronger throughout the atmosphere for the strong shear case, the time average second mode zonal mean zonal wind deviation from the control case at each level for both the weak and strong shear cases is computed at 60°N (Figure 9). At nearly all levels, the second mode winds are on average weaker for the weak shear case; while for an increased temperature gradient, the winds are on average more westerly. Thus by increasing the radiative forcing, we find that the model prefers higher NAM index values.

It is also of interest to investigate the time mean response of the total fields to the increased radiative forcing. Figure 10a shows the time-average difference of the EP flux between the strong shear case and the control run. The EP flux vectors clearly show increased equatorward propagation of planetary waves during the strong shear case throughout most of the model domain. This is consistent with the Figure 9. In addition, Figure 10a generally shows increased vertical wave propagation below 35 km and decreased vertical wave propagation above. This suggests the possibility of a region of EP flux convergence centered near 35 km.

Figure 6. Time versus height plot of $U_C$ at 60°N for 125 m forcing and the following $U_{Ri}$ shear between 6 and 14 km: (a) 2.7 m s$^{-1}$ km$^{-1}$, (b) 3.3 m s$^{-1}$ km$^{-1}$. Contour interval is 5 m s$^{-1}$ with the zero and negative contours dashed. Shading indicates westerly winds.
The time-average deviation of the strong shear case zonal mean zonal wind from the control case is displayed in Figure 10b. The average second mode zonal mean zonal wind, $U_c$, is stronger for the strong shear case at all levels (Figure 9); thus, the majority of the vertical structure in Figure 10b is associated with the first mode zonal mean zonal wind. Therefore, the negative average deviation at 35 km indicates that the polar night jet is on average slightly weaker around 35 km in the strong shear case. This decrease in wind speed is most likely a result of the EP flux convergence region seen in Figure 10a. It should be noted that since the zonal mean zonal wind oscillates between positive and negative values, the negative deviation shown in Figure 10b could be caused by either weaker westerly winds or stronger easterly winds. Earlier figures suggest that both factors contribute to the negative deviation. Despite the small decrease in the average strength of the polar night jet, the average response of the zonal mean zonal wind in the troposphere appears to be consistent with higher NAM values.

3.3. Upper Boundary Height

The GCM used by Shindell et al. [1999] was able to produce a positive trend of the NAM in response to increased GHG forcing for simulations where the stratosphere was fully resolved. These model runs had an upper boundary of 85 km. When Shindell et al. [1999] lowered the upper boundary of the GCM to 30 km, so that only two dynamically active layers were above the tropopause, the model still produced a NAM, but the NAM was unaffected by increases in GHG forcing. In contrast, Fyfe et al. [1999] was able to produce a GHG induced trend in the NAM without a fully resolved stratosphere. In order to examine the influence of the stratosphere on our model results, a series of experiments is performed for different heights of the upper boundary. All parameters other than the height of the upper boundary are the same as before;

Figure 7. Time versus $U_c$ at 6 km and 60°N for both the 2.7 m s$^{-1}$ km$^{-1}$ and 3.3 m s$^{-1}$ km$^{-1}$ shear cases.

Figure 8. Time versus $F_y$ at 6 km and 60°N for both the 2.7 m s$^{-1}$ km$^{-1}$ and 3.3 m s$^{-1}$ km$^{-1}$ shear cases.
the radiative equilibrium shear is 3.0 m s\(^{-1}\) km\(^{-1}\), unless stated otherwise.

[37] The model response within the troposphere and lower stratosphere is very similar for model runs with an upper boundary of at least 30 km. The first mode zonal mean zonal wind, \(U_A\) is nearly constant at lower levels, and the second mode zonal mean zonal wind oscillates between easterly and westerly values which correspond to low and high-index events, respectively, at 60\(^\circ\)N (see Figure 2). Since the oscillation of the polar vortex in Figure 2 occurs primarily above 30 km, placing the upper boundary at 30 km effectively eliminates such variability from the model. Despite this, the model still produces a NAM signature throughout the troposphere and stratosphere. If the upper boundary is set at 16 km or lower, though, then a NAM signature is absent from the model simulations.

[38] A regression map of the total zonal mean zonal wind from a model run with the upper boundary at 30 km is shown in Figure 11a. The regression values were computed by regressing the total zonal mean zonal wind onto the first standardized principle component of the zonal mean zonal wind. The principle component was computed using zonal mean zonal wind values over the entire domain. (If instead the principle component is calculated using only tropospheric wind data, the leading regression map is very similar.) The first principle component explains 41% of the variability and is well separated from the other principle components. Figure 11a shows a dipole in the zonal mean zonal wind field with a node between 45\(^\circ\) and 50\(^\circ\)N, and
centers of action near 30° and 60°N. This regression map compares well with the calculations of Baldwin and Dunkerton [1999] and Thompson and Wallace [2000] where zonal mean zonal wind observations are regressed onto the NAM index. The main discrepancy between our model data and observations lies in the strength of the southern branch of the dipole. Both Baldwin and Dunkerton [1999] and Thompson and Wallace [2000] show the southern branch as having less amplitude than the northern branch. The model shows nearly equal strength in both branches of the dipole, which is probably a result of the β-plane geometry and the severe truncation in the model’s meridional direction. In addition, the regression amplitudes are slightly less than observations in the lower stratosphere and troposphere. This is conceivably due to the model not resolving synoptic-scale waves, which also contribute to the wave forcing of the NAM [Limpasuvan and Hartmann, 1999]. Regression maps of the zonal mean zonal wind were also calculated for model integrations with the upper boundary at altitudes above 30 km. These maps yield a similar dipole structure in the troposphere and lower stratosphere. However, the variability at upper levels is dominated by a single maximum centered slightly north of 45°N and is strongly tied to the strength of the polar vortex.

[39] The regression map of the zonal mean zonal wind from a model integration with the upper boundary at 16 km is shown in Figure 11b. The first principle component explains 60% of the variance. (The process used to calculate this regression map is analogous to that described above with the upper boundary at 30 km.) This regression map differs markedly from the previous case, though, as the leading mode of variability shows no characteristics identifiable with the NAM. Additional runs of the model were made with the upper boundary at 10 km which yielded very similar results. Increasing the vertical resolution to 1 km did not alter this. Thus, when the upper boundary is at 16 km or below, a NAM signature is essentially absent from the model.

[40] Results were also obtained from model integrations analogous to the stronger shear case with the upper boundary at both 30 km and 16 km. The results from the 30 km top strong shear case are very similar to those shown with the upper boundary at 90 km. In particular, eddy momentum fluxes associated with increased equatorward propagation of planetary waves were again responsible for forcing the model toward higher NAM values. The 16 km top strong shear case, though, yielded only small departures from the control run. Examining the EP flux vectors for the two different 16 km top cases reveals that the vertical component is dominant. In both the control and strong shear cases, the meridional component of the EP flux is very small, implying that there is very little meridional propagation of the waves.

[41] The results presented here imply that a model with the current formulation needs to have an upper boundary height of approximately 30 km or higher to be able to produce NAM-like variability. When the upper boundary is lower than 30 km, the meridional propagation of stationary waves is inhibited reducing the eddy momentum fluxes that force the zonal wind anomalies associated with the NAM. The lack of NAM variability in the model when the upper boundary is placed at 16 km or below reveals the dependence of NAM variability in the model on the eddy fluxes of momentum but also a limitation of the current model. Synoptic-scale baroclinic eddies, which are also known to force zonal wind anomalies associated with the NAM [Limpasuvan and Hartmann, 1999], are not incorporated into the current model. If such effects were included into the model, it is possible that NAM variability would be present with the model boundary at 16 km. Because of this limitation, it is difficult to make general statements concerning the necessity of the stratosphere for resolving NAM variability. Nevertheless, our results do show that if baroclinic eddies are not incorporated into a model, then the model needs to be able to properly account for the meridional propagation of planetary waves to produce a NAM signature.

3.4. Conclusions

[42] The results shown here reveal that a mechanistic model describing planetary wave-mean flow interaction on a β-plane integrated with perpetual winter conditions is able to qualitatively describe characteristics of the observed NAM: the structure and amplitude of the leading regression map. The results also suggest that our model lacks the necessary stratospheric scales to reproduce the observed NAM variability. Nevertheless, our results do show that if baroclinic eddies are not incorporated into a model, then the model needs to be able to properly account for the meridional propagation of planetary waves to produce a NAM signature.

References


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