Pathways for Communicating the Effects of Stratospheric Ozone to the Polar Vortex: Role of Zonally Asymmetric Ozone

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ABSTRACT

A mechanistic model that couples quasigeostrophic dynamics, radiative transfer, ozone transport, and ozone photochemistry is used to study the effects of zonal asymmetries in ozone (ZAO) on the model’s polar vortex. The ZAO affect the vortex via two pathways. The first pathway (P1) hinges on modulation of the propagation and damping of a planetary wave by ZAO; the second pathway (P2) hinges on modulation of the wave–ozone flux convergences by ZAO. In the steady state, both P1 and P2 play important roles in modulating the zonal-mean circulation. The relative importance of wave propagation versus wave damping in P1 is diagnosed using an ozone-modified refractive index and an ozone-modified vertical energy flux. In the lower stratosphere, ZAO cause wave propagation and wave damping to oppose each other. The result is a small change in planetary wave drag but a large reduction in wave amplitude. Thus in the lower stratosphere, ZAO “precondition” the wave before it propagates into the upper stratosphere, where damping due to photochemically accelerated cooling dominates, causing a large reduction in planetary wave drag and thus a colder polar vortex. The ability of ZAO within the lower stratosphere to affect the upper stratosphere and lower mesosphere is discussed in light of secular and episodic changes in stratospheric ozone.

1. Introduction

Among the vast number of ozone dynamics studies of the stratosphere, most have focused on the role of zonal-mean ozone (ZMO) (WMO 2007). The relatively few studies that have focused on zonal asymmetries in ozone (ZAO) have shown that they can have a significant impact on the large-scale circulation in the stratosphere. Indeed, mechanistic modeling studies have shown that ZAO can affect zonal-mean ozone transport (Garcia and Hartmann 1980; Hartmann and Garcia 1979), photochemical damping of inertia–gravity waves (Zhu and Holton 1986; Xu 1999; Xu et al. 2001), photochemical destabilization of free Rossby waves (Nathan 1989; Nathan and Li 1991; Nathan et al. 1994), modulation of the quasi-biennial oscillation (Cordero et al. 1998; Cordero and Nathan 2000), solar cycle modulation of both the quasi-biennial oscillation (Cordero and Nathan 2005) and the Arctic polar vortex (Nathan et al. 2011), and planetary wave drag (Nathan and Cordero 2007, hereafter NC07).

A few studies have employed global circulation models (GCMs) to show that ZAO may impact climate variability (Sassi et al. 2005; Gabriel et al. 2007; Crook et al. 2008; Waugh et al. 2009; Gillett et al. 2009; McCormack et al. 2011). For example, McCormack et al. (2011) used a GCM with interactive ozone chemistry and found that including ZAO altered the strength of stratospheric planetary wave drag and increased the frequency of Northern Hemisphere sudden stratospheric warmings. Gabriel et al. (2007) used a GCM and found that ZAO resulted in warming of 4–6 K in the middle to lower stratosphere and cooling of 4–8 K in the upper stratosphere and lower mesosphere during Northern Hemisphere winter. Crook et al. (2008) determined that ZAO produce changes in the middle atmosphere circulation that are similar in magnitude to those caused by severe stratospheric ozone depletion. In addition, they found that regions of maximum (minimum) ozone due to ZAO correlated with regions of negative (positive) temperature anomalies. Crook et al. concluded that changes in the temperature structure induced by ZAO were due to changes in dynamical heating, not radiative heating. Crook et al. and Gillett
et al. both hypothesized that the differences in dynamical heating associated with ZAO were likely caused by a decrease in the upward flux of planetary wave activity. This hypothesis, however, remains untested.

Our goal in this study is to isolate the physics that forms pathways for communicating changes imparted by ZAO to the polar vortex. As we show through numerical experiments and ozone-modified wave diagnostics, the effects of ZAO are communicated vertically along two pathways that combine to alter the model’s polar vortex (see Fig. 1). Along each pathway ZAO play a crucial role. Along pathway 1 (P1), ZAO modulate wave propagation and wave damping (attenuation), which together modulate the vertical energy flux (VEF) and planetary wave drag (PWD). Along pathway 2 (P2), ZAO produce divergences of wave–ozone flux that modulate the zonal-mean ozone heating and thus zonal-mean temperature. In the steady state, we show that both P1 and P2 play important roles in communicating the effects of ozone to the polar vortex.

To see how ZAO modulate the zonal-mean circulation along P1, we consider first the expression for the ozone-modified planetary wave drag derived in NC07 [their Eq. (14)]. They showed that the PWD for a vertically propagating planetary wave in a slowly varying background flow can be written as

\[ \text{PWD} \approx -|A|^2 m_j m_i \exp \int_0^z (-m_j) dz', \quad (1) \]

where \( m_j \) and \( m_i \) measure, respectively, wave propagation and wave damping and \( A \) is the wave amplitude, which also depends on \( m_j \) and \( m_i \). The propagation and damping effects may either oppose or augment each other to produce either a small or large change in PWD. As we show later, in the case where the local change in ozone-modified PWD is small, the consequences for the greater circulation are nonetheless important. We explain this apparent paradox as follows. Imagine a planetary wave moving vertically through the lower stratosphere, where the ZAO cause an increase in wave propagation that is offset by an increase in wave damping. Locally, then, the PWD does not change and thus will not affect the zonal-mean flow. The wave amplitude, however, does change; it is reduced by the increase in wave damping. So, as the wave moves through the lower stratosphere, the zonal-mean flow is unaltered, but the wave becomes “preconditioned,” leaving the lower stratosphere with a smaller amplitude than it would have had in the absence of ZAO.

The scenario just described hinges largely on the phasing between the zonally asymmetric wind, temperature, and ozone fields (NC07). As our numerical simulations show, depending on the ratio of advective to photochemical time scales, the three wave fields may combine to produce either wave damping or wave amplification, or positive or negative wave–ozone flux convergences. In the dynamically controlled lower stratosphere, where the advective time scales are much faster than the photochemical time scales, ZAO produce either heating or cooling, depending on the relative magnitudes of the meridional and vertical gradients of zonal-mean ozone. We show that in the lower stratosphere there is little change in PWD due to ozone but a significant reduction in wave amplitude, consistent with our discussion of Eq. (1). As the preconditioned wave enters the middle stratosphere—termed the transition region—where ozone transport and photochemistry are of comparable importance, wave damping dominates over wave propagation, causing changes in PWD and thus the zonal-mean circulation. In the photochemically controlled upper stratosphere and lower mesosphere, where photochemical time scales are much shorter than advective time scales, a positive temperature perturbation produces a negative ozone perturbation (Craig and Ohring 1958). The negative correlation between the temperature and ozone perturbations causes thermal relaxation, that is, photochemically accelerated cooling, which damps the fields. We show that in the upper stratosphere wave damping strongly dominates over wave propagation. The result is a large reduction in the PWD and thus a colder polar vortex.

The effect of P2 on the zonal-mean circulation depends on the ability of wave–ozone flux convergences due to ZAO to modify the zonal-mean ozone distribution; the change in the zonal-mean ozone distribution then alters the zonal-mean heating rate and thus the thermodynamic structure of the middle atmosphere. Thus the first part of P2 is an ozone transport issue; this topic has been covered in detail in Hartmann and Garcia (1979) and Garcia and Hartmann (1980). However, the effect of wave–ozone flux convergences on the zonal-mean circulation of the middle atmosphere and its importance relative to P1 remains unknown.

By examining the pathways that communicate changes in ozone to the polar vortex, we focus on the role of ZAO. In so doing, we address several questions: what role do ZAO play in coupling the lower and upper stratosphere; can changes to planetary wave activity in the lower stratosphere impact the upper stratosphere and lower mesosphere; if ZAO are important to communicating local changes in ozone throughout the greater stratosphere, what is the relative importance of ozone transport versus ozone photochemistry in modulating the interaction between the planetary wave field and the polar vortex.

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\(^1\) Equation (1) differs slightly from NC07’s PWD equation in that we have stripped it of factors that do not bear on its physical interpretation.
and, if ZAO cause a decrease in the vertical energy flux into the stratosphere, is this a result of changes to the curvature and strength of the polar vortex (i.e., wave propagation) or do ZAO directly affect the vertical energy flux via wave damping?

We organize the paper as follows. Section 2 describes the numerical model, section 3 presents the theoretical framework and diagnostics that we use to interpret our numerical results in section 4, and section 5 summarizes our conclusions.

2. Model

We employ a mechanistic wave–mean flow interaction model of the extratropical atmosphere that couples radiative transfer, ozone transport, ozone photochemistry, and the dynamical circulation. The dynamical portion of the model is based on Holton and Mass (1976), while the radiative–photochemical portion of the model is based on Nathan and Li (1991). In particular, we consider a quasi-geostrophic atmosphere that is confined to a beta-plane channel of width \( L = 6000 \text{ km} \) centered at \( \theta_0 = 60^\circ \text{N} \) latitude. The model consists of coupled equations for wind, temperature, and ozone in which each field is partitioned into a zonal-mean part (denoted by an overbar) and a wave part (denoted by a prime). As in other studies, we use the zonal-mean zonal wind as a proxy for the polar vortex (e.g., Harnik 2009). The wave equations for potential vorticity, ozone, and temperature are

\[
\frac{\partial}{\partial t} \left( \frac{\partial}{\partial x} + \bar{u} \frac{\partial}{\partial x} \right) q' + \frac{\partial \psi' \partial Q}{\partial x} = \frac{\kappa f_0}{\rho H} \frac{\partial}{\partial z} \left( \frac{\rho}{N^2} \frac{\partial q'}{\partial z} \right),
\]

where the perturbation potential vorticity and basic-state potential vorticity gradient are

\[
q' = \nabla^2 \psi' + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{\rho f_0^2}{N^2} \frac{\partial \psi'}{\partial z} \right),
\]

\[
\frac{\partial Q}{\partial y} = \beta \frac{\partial^2 \bar{u}}{\partial y^2} + \frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{\rho f_0^2}{N^2} \frac{\partial \bar{u}}{\partial z} \right).
\]

The wave-heating rate and wave-ozone production/destruction rate are, respectively,

\[
J' = \left( -\frac{f_0 H}{\kappa} \frac{\partial \psi'}{\partial z} \right) + \left[ \Gamma_1 \gamma' - \Gamma_2 \int_z \frac{\rho(z')}{\rho_0} \gamma' \, dz' \right],
\]

\[
S' = -\xi_1 \gamma' + \xi_2 \int_z \frac{\rho(z')}{\rho_0} \gamma' (x, y, z, t) \, dz' - \frac{f_0 H}{R} \xi_T \frac{\partial \psi'}{\partial z},
\]

The zonal-mean equations for zonal wind, ozone, temperature, and meridional circulation are

\[
\frac{\partial}{\partial t} \left( \frac{\partial}{\partial y} \right) \bar{u} = -\frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{\rho f_0^2}{N^2} \frac{\partial \bar{u}}{\partial z} \right) + \frac{1}{\rho H} \frac{\partial}{\partial y} \left( \frac{\rho f_0^2}{N^2} \frac{\partial \bar{u}}{\partial y} \right),
\]

\[
\frac{\partial}{\partial t} \frac{\partial \gamma}{\partial y} = -\left( \frac{\partial \gamma}{\partial y} + \bar{w} \frac{\partial \gamma}{\partial z} \right) - \frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{\rho w' \gamma'}{\rho} \right) + S,
\]

\[
\frac{\partial^2 \bar{T}}{\partial y^2} + \frac{H N^2}{R} \frac{\partial \gamma}{\partial y} = -\frac{1}{c_p} \frac{\partial}{\partial y} \left( \frac{\rho}{\rho} \frac{\partial \bar{T}}{\partial y} \right) + \frac{1}{c_p} \frac{\partial}{\partial y} \left( \frac{\rho}{\rho} \frac{\partial \bar{T}}{\partial y} \right),
\]

where the zonal-mean meridional and vertical wind velocities are given by

\[
(\bar{v}, \bar{w}) = \left( \frac{\partial \gamma}{\partial y}, \frac{\partial \gamma}{\partial z} \right).
\]

The zonal-mean ozone heating rate and ozone production/destruction rate are, respectively,

\[
J = \Gamma_1 (\bar{\gamma} - \bar{\gamma}) - \Gamma_2 \int_z \frac{\rho(z')}{\rho_0} (\bar{\gamma} - \bar{\gamma}) \, dz',
\]

\[
S = -\xi_1 (\bar{\gamma} - \bar{\gamma}) + \xi_2 \int_z \frac{\rho(z')}{\rho_0} (\bar{\gamma} - \bar{\gamma}) \, dz' - \xi_T (\bar{T} - \bar{T}),
\]

\[2\] For ozone transport studies, it is often more illuminating to cast the meridional circulation equation in terms of its transformed Eulerian-mean form, wherein residual meridional and vertical velocities are used. Because ozone transport is outside the scope of this study, we opt to write the meridional circulation equation in its Eulerian form [Eq. (10)].
The subscript $R$ denotes fields that are in radiative and photochemical equilibrium (we discuss the equilibrium fields later). All symbols in Eqs. (2)–(13) are defined in Table 1.

We base the radiative–photochemical parameterizations in the thermodynamic and ozone continuity equations on Nathan and Li (1991) and have updated the parameterizations to include updated reaction rates based on Sander et al. (2006); an accounting of ozone catalytic loss cycles involving hydrogen, nitrogen, and chlorine based on the approximation of Haigh and Pyle (1982); temperature-dependent ozone absorption cross sections from Molina and Molina (1986); and enhancement of solar radiation due to multiple scattering, surface reflection, and clouds and aerosols using the approximations of Meier et al. (1982) and Nicolet et al. (1982). We give additional details about the parameterizations in the appendix.

The diabatic heating rates $J'$ and $\overline{J}$ each contain three terms: Newtonian cooling, local ozone heating, and non-local ozone heating due to perturbations in ozone above a given level, termed the shielding effect (Hartmann 1978; Zhu and Holton 1986). The Newtonian cooling approximation is based on Dickinson (1973), which models cooling to space due to the 15-μm band of carbon dioxide and 9-μm band of ozone. The ozone production and destruction terms $S'$ and $\overline{S}$, which appear in the wave–ozone equation [Eq. (6)] and zonal-mean ozone equation [Eq. (13)], comprise local effects, shielding effects, and coupling with temperature.

Solutions to Eqs. (2)–(13) require initial and boundary conditions. We initialize the model with climatological distributions of zonal-mean wind, temperature, and ozone, which are constructed for late winter (February) using data tabulated by Randel et al. (2002). At the channel boundaries ($y = 0$ and $y = L$) the kinematic boundary condition is applied, which requires that the northward velocity vanish there. At the lower boundary the initial zonal-mean wind is set to its climatological value, while at the upper boundary we impose $\partial v / \partial z = 0$ at $z = z_u$. As in Holton and Mass (1976), the planetary wave is periodic in $x$ and vanishes at the upper boundary ($z_u = 90$ km), which is placed at sufficiently great height to prevent spurious reflections. At the lower boundary ($z_R = 10$ km) we impose a wave that grows monotonically from zero to an asymptotic steady state. We write this boundary condition as $\psi'(x, y, z_R, t) = gh_B [1 - \exp(-\tau')] \exp(ikx) \sin y$, where $k = v/a_e$, $\cos \theta_0$ is the zonal wavenumber, $n$ is the quantized zonal wavenumber, $a_e$ is the earth’s radius, $\theta_0 = 60^\circ$, $l = \pi/a_e$, and $h_B$ is the asymmetrical steady-state amplitude of the geopotential height perturbation. Depending on the size of $h_B$, the zonal-mean wind may either reverse from westerly to easterly, corresponding to a stratospheric warming, or remain westerly as it vacillates. In either case, the zonal-mean wind eventually asymptotes to a steady state. Here we choose $h_B = 35$ gpm, a value that causes the zonal-mean flow to remain westerly as it evolves to a steady state after $\sim 150$ days. We have carried out model simulations for planetary waves $1 (n = 1$ and $2 (n = 2$ and the results were robust across a wide range of parameter settings; results are shown for $n = 1$ and $\tau = 2.5 \times 10^5$ s. All comparisons between the ozone and no-ozone simulations, presented later, are based on steady-state conditions. We defer examination of the effects of ozone on the transient properties of the model circulation to a future study.

We seek solutions to Eqs. (2)–(13) in the form of highly truncated Fourier series for which we write the planetary wave, zonal-mean fields, and radiative–photochemical coefficients each as a single Fourier mode:

Table 1. List of symbols.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
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<tbody>
<tr>
<td>$t, x, y, z = -H \ln(p/p_0)$</td>
<td>Time and distances in the eastward, northward, and vertical directions</td>
</tr>
<tr>
<td>$\rho(z) = \rho_0 \exp(-z/H), \rho_0$</td>
<td>Basic-state pressure, reference pressure at the surface of the earth</td>
</tr>
<tr>
<td>$\rho(z) = \rho_0 \exp(-z/H), \rho_0$</td>
<td>Basic-state density, reference density at the surface of the earth</td>
</tr>
<tr>
<td>$f_0, \beta, H$</td>
<td>Planetary vorticity, planetary vorticity gradient evaluated at $\theta_0 = 60^\circ$ latitude, and mean scale height (7 km)</td>
</tr>
<tr>
<td>$N^2$</td>
<td>Brunt–Väisälä frequency squared ($4 \times 10^{-4}$ s$^{-2}$)</td>
</tr>
<tr>
<td>$k = R/C_p$</td>
<td>$R$ is the gas constant and $C_p$ is the specific heat at constant pressure</td>
</tr>
<tr>
<td>$\psi', T, \psi', v', w'$</td>
<td>Perturbation geostrophic streamfunction, temperature, ozone volume mixing ratio, and meridional and vertical winds</td>
</tr>
<tr>
<td>$\pi, T, \gamma$</td>
<td>Zonal-mean zonal wind, temperature, and ozone volume mixing ratio</td>
</tr>
<tr>
<td>$\pi, \overline{w}, \overline{x}$</td>
<td>Zonal-mean meridional and vertical wind, and zonal-mean meridional streamfunction</td>
</tr>
<tr>
<td>$\Gamma(z; \gamma, T, \mu)$ ($j = 1, 2$)</td>
<td>Radiative–photochemical heating coefficients</td>
</tr>
<tr>
<td>$\alpha(z)$</td>
<td>Newtonian cooling coefficient</td>
</tr>
<tr>
<td>$\xi(z; \gamma, T, \mu)$ ($j = 1, 2, T$)</td>
<td>Radiative–photochemical ozone production and destruction coefficients</td>
</tr>
<tr>
<td>$\mu$</td>
<td>Solar zenith angle</td>
</tr>
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</table>
(ψ', γ', w') = \left[ \Psi'(z,t), \gamma'(z,t), w'_R(z,t) \right] \sin(ly) + c.c.
(\pi, \nu, \chi) = \left[ U_0(z,t), \nu_0(z,t), \chi_0(z,t) \right] \sin(ly)
\bar{w} = \frac{U_0(z,t) \cos(ly)}{\bar{w}_0(z,t) \cos(ly)}.
(14a–e)
\left( \gamma, T \right) = \left[ \gamma_0(z,t), T_0(z,t) \right] + \left[ \gamma_1(z,t), T_1(z,t) \right] \cos(ly)
(i = 1, 2)
\left( \Gamma_i, \xi_i, \xi_T \right) = \left[ \Gamma_{i,0}(z), \xi_i0(z), \xi_{T,0}(z) \right] + \left[ \Gamma_{i,1}(z), \xi_i1(z), \xi_{T,1}(z) \right] \cos(ly),

Insertion of Eq. (14) into Eqs. (2)–(13), subsequently projecting onto \sin ly and \cos ly, yields a coupled set of equations governing the time–height evolution of the wind, temperature, and ozone fields. To determine the evolution of the system, however, first requires specifying the radiative–photochemical equilibrium state of the model.

In the absence of wave driving, the model relaxes back to a state in which wind, temperature, and ozone are in equilibrium. Given the equilibrium temperature, the equilibrium zonal wind is then determined from thermal wind balance. We use two methods to calculate the radiative–photochemical equilibrium distributions of temperature and ozone. In the first method, we integrate the nonlinear ozone continuity and thermodynamic energy equations forward in time until radiative and photochemical equilibrium is reached [i.e., we use Eqs. (A1) and (A2) in the appendix]. In the second method, we specify the equilibrium temperature profile and integrate the ozone continuity equation [Eq. (A1)] forward in time until equilibrium is also reached. For the second method, we use the Kushner and Polvani (2006) analytic function for temperature, which produces equilibrium temperature profiles that are in excellent agreement with those obtained by Fels (1985), who combined a radiative–photochemical model of the stratosphere and mesosphere with a radiative-convective model of the troposphere. Because the second method, rather than the first, produces equilibrium wind, temperature, and ozone profiles that are in slightly better agreement with Fels, we use the second method for our equilibrium calculations.

We calculate the equilibrium profiles for zonal-mean temperature and ozone at 40°, 60°, and 80°N. We then use these profiles to obtain equilibrium profiles for the meridional gradients of ozone and temperature at mid-channel:

\[ \gamma_{1,R} = -\frac{1}{l} \frac{\partial \gamma_R}{\partial y} = -\frac{1}{l} \left[ \gamma_{0,R}(80°) - \gamma_{0,R}(40°) \right] \]
\[ T_{1,R} = -\frac{1}{l} \frac{\partial T_R}{\partial y} = -\frac{1}{l} \left[ T_{0,R}(80°) - T_{0,R}(40°) \right], \]

where \( l = \pi/a_e \). Given Eqs. (15) and (16), we then determine the equilibrium zonal-mean wind by the thermal wind relation

\[ U_{0,R} = \frac{-R}{\bar{f}_0 H} \frac{\partial T}{\partial y} dz + U_{0,z_R} = \frac{RI}{\bar{f}_0 H} \frac{\partial T}{\partial y} dz + U_{0,z_R}, \]

where we have assumed \( U_0(z_R) = 15 \text{ m s}^{-1} \).

3. Theoretical basis

To ease interpretation of the numerical results, we first provide a theoretical framework that allows us to assess how radiation, ozone, and dynamics combine to modulate the wave and zonal-mean fields along the two pathways shown in Fig. 1. To understand how ZAO operate along P1, we adopt the NC07 ozone-modified refractive index (OMRI), which describes how ozone heating (OH) and Newtonian cooling (NC) combine to affect wave propagation and wave damping (attenuation). We use the OMRI to form diagnostics for planetary wave drag (PWD) and vertical energy flux (VEF). To understand how ZAO operate along P2 we use a steady-state form of the zonal-mean ozone equation [Eq. (8)]. We then examine the relative importance of P1 and P2.

The net driving of the zonal-mean wind is due to the combined effects of planetary wave drag \( \bar{u}_{PWD} \), Newtonian cooling \( \bar{u}_{NC} \), and ozone heating \( \bar{u}_{OH} \). In the steady state, the zonal-mean wind equation [Eq. (7)] can be written as

\[ 0 = \frac{\partial^2}{\partial y^2} \left[ \frac{\partial}{\rho \frac{\partial f_0^2}{N^2}} \frac{\partial \gamma}{\partial z} \right] + \frac{1}{\kappa} \frac{\partial}{\partial z} \left( \frac{\rho f_0^2}{N^2} \frac{\partial T}{\partial y} \right) \]
\[ -\frac{1}{\rho} \frac{\partial f_0^2}{N^2} \alpha \frac{\partial u}{\partial z} - \frac{\partial \bar{u}_R}{\partial z} \]

As the zonal-mean wind evolves to the steady-state balance described by Eq. (18), the PWD term \( \bar{u}_{PWD} \), which is modulated by ozone, decelerates the zonal-mean wind, driving it away from radiative equilibrium to create a weaker and warmer polar vortex. In contrast, the \( \bar{u}_{NC} \)-term relaxes the zonal-mean wind toward radiative equilibrium, corresponding to a colder and stronger polar vortex. Depending on the ratio of photochemical to dynamical time...
scales, the $\bar{u}_{\text{OH}}$ term may either damp or enhance the zonal-mean wind, driving the zonal-mean wind either toward or away from radiative equilibrium.

Figure 2 shows schematically the nonlinearly coupled pathways and physical processes that contribute to $\bar{u}_{\text{PWD}}$, $\bar{u}_{\text{OH}}$, and $\bar{u}_{\text{NC}}$ in the steady-state zonal-mean wind equation [Eq. (18)]. The contribution by the planetary wave drag $\bar{u}_{\text{PWD}}$ depends on NC and OH, which correspond to the terms on the rhs of the wave-heating equation [Eq. (5)]. The wave OH, which operates along P1 and arises from ZAO, is governed by the wave–ozone equation [Eq. (3)] and comprises two diabatic heating effects: advection of zonal-mean ozone by the wave and production and destruction of wave–ozone. As shown in NC07, $\bar{u}_{\text{PWD}}$ vanishes in the absence of NC and OH (Charney and Drazin 1961).

The contribution to the mean wind by the mean ozone heating term $\bar{u}_{\text{OH}}$ depends on the mean ozone distribution [see Eq. (12)]. The mean ozone distribution depends on three processes [see Eq. (8)]: advection of zonal-mean ozone by the zonal-mean meridional circulation, divergence of wave–ozone flux (shown as $\nabla \cdot \vec{F}$ in Figs. 1 and 2), and zonal-mean ozone production and destruction. In the steady state, and in the absence of momentum fluxes (as in our model), $\vec{v}$ and $\vec{w}$ vanish so that the advection of zonal-mean ozone by the wave–ozone production and destruction and $\nabla \cdot \vec{M}$. The zonal-mean ozone production and destruction consists of photochemical relaxation [$\xi_1$ and $\xi_2$ in Eq. (13)] and temperature-dependent effects [$\xi_T$ in Eq. (13)]; only $\nabla \cdot \vec{M}$, which operates along P2, depends directly on ZAO.

To determine how the physical processes operating along P1 vary with altitude and are modulated by wave propagation and wave damping, we consider next the OMRI derived by NC07. From the OMRI we obtain expressions for the PWD and VEF, which we use to interpret the numerical results in section 4.

NC07 give a detailed derivation and analysis of the OMRI; we summarize their analysis here to the extent necessary to interpret our model results. Briefly, if the background fields of wind, temperature, and ozone are
slowly varying in the vertical, we seek solutions for the streamfunction and ozone fields of the form

\[
\psi(x, y, z; \Delta), \gamma(x, y, z; \xi; \Delta) = [\psi(\xi; \Delta), \gamma(\xi; \Delta)] \exp(\zeta/2H) \exp(ikx) \sin^2 \theta + \text{c.c.}
\]

(19)
in which \( \zeta = \Delta z \) is the slowly varying vertical coordinate and \( \Delta \ll 1 \) is nondimensional. The slowly varying vertical structures for the streamfunction and ozone fields are chosen using the WKB approximation in the form (Bender and Orszag 1978):

\[
[\psi(\zeta), \gamma(\zeta)] = [A(\zeta), B(\zeta)] \exp \left\{ i \left[ \frac{1}{\Delta} \int_0^\zeta m(\zeta') \, d\zeta' \right] \right\}.
\]

(20)

In Eq. (20), \( m = m + im_t \) is the OMRI, or equivalently, the local (complex) vertical wavenumber, where \( m_r \) measures wave propagation and \( m_t \) measures wave damping; \( A(\zeta) \) and \( B(\zeta) \) are the streamfunction and ozone amplitudes, respectively, which are both functions of \( m_r \) and \( m_t \). The OMRI and amplitudes [see NC07, their Eqs. (11) and (12)] are nonlinear functions of NC and OH. This means that NC and OH are coupled rather than additive.

The PWD is a measure of the wave driving of the zonal-mean circulation. As shown in NC07, insertion of Eq. (20) into Eqs. (2)–(6) yields the ozone-modified PWD equation [Eq. (1)]. In addition to PWD, we also consider the vertical energy flux. Consideration of the VEF is motivated in part by Crook et al. (2008) and Gillett et al. (2009), who speculated that small changes in the strength and shape of the polar vortex induced by the ZAO might modulate the VEF in the stratosphere.

We determine whether changes in the VEF are due to changes in strength and curvature of the polar vortex (wave propagation) or whether ZAO directly affects the VEF via an increase in wave damping. Therefore, in addition to the PWD, another important diagnostic for measuring the effects of lower-stratospheric OH on the upper-stratospheric polar vortex, which NC07 did not consider, is the zonally averaged VEF,
where the $\text{VEF}_{HF}$, $\text{VEF}_{NC}$, and $\text{VEF}_{OH}$ denote wave contributions from the northward heat flux, Newtonian cooling, and ozone heating, respectively. As we show in section 4, the VEF is dominated by the $\text{VEF}_{HF}$, which itself is a function of NC and OH. In terms of the OMRI, the $\text{VEF}_{HF}$ can be written to lowest order as

$$\text{VEF}_{HF} \propto \frac{\pi}{m_i} \text{PWD}, \quad (22)$$

where, recall, PWD depends on wave damping and wave propagation and is given by Eq. (1).

We measure changes in wave damping by wave amplitude [Eq. (14a)], which, as we noted earlier, is a function of $m_i$ (as $m_i$ increases, wave amplitude decreases). We approximate the wave propagation by using $m_r \approx m_o$, where

$$m_0^2 = N^2 \left[ \frac{\pi}{\tau} \frac{\partial Q}{\partial y} - (k^2 + \ell^2) \right] - \frac{1}{4H^2} \quad (23)$$

is the lowest-order approximation to the propagation, which corresponds to the square of the classical one-dimensional refractive index originally derived by Charney and Drazin (1961). To a first approximation, the height of the reflecting surface is located where $m_0 = 0$. The ozone physics is indirectly contained in the propagation term $m_0$ since $\overline{u}$ and $\overline{Q}$ each depend on ozone [see Eq. (7)].

Planetary waves propagate in regions where $m_0^2 > 0$ and evanesce in regions where $m_0^2 < 0$; larger values of $m_0^2$ foster greater wave propagation. Specifically, larger (positive) values of the zonal-mean wind $\overline{u}$ decrease $m_0^2$, while larger (positive) values of the curvature of the zonal-mean wind, measured by the background potential vorticity gradient $\overline{Q}_y$, increase $m_0^2$.

4. Numerical experiments

We present three experiments that show how zonal asymmetries in ozone modulate the zonal-mean circulation along the two pathways shown in Figs. 1 and 2. In the first experiment (No-O3), we exclude ozone; this experiment reproduces the wave–mean flow interaction results of Holton and Mass (1976). In the second experiment (Full-O3), we include both zonal-mean ozone (ZMO) and zonal asymmetries in ozone (ZAO). In the third experiment (Mean-O3), we artificially suppress ZAO but retain ZMO. For each experiment, we use the boundary and initial conditions stated in the model section (section 2).

Results

1) ZONAL-MEAN WIND

Figure 3 shows the steady-state vertical distribution of zonal-mean wind for the no-ozone (No-O3), full-ozone (Full-O3), and zonal-mean ozone (Mean-O3) experiments.
We also show the radiative equilibrium distribution of the wind. To facilitate comparison of the experiments, we remind the reader of the strong altitude dependence of the ozone physics: dynamical processes control ozone below \( \sim 35 \) km and photochemical processes control ozone above \( \sim 45 \) km (see NC07, their Fig. 3). Dynamical and photochemical processes together control ozone in the transition region (35–45 km). The response of the zonal-mean wind is different for each region. Above \( \sim 40 \) km, the zonal-mean wind of Mean-O3 and Full-O3 are both stronger than for No-O3; below \( \sim 40 \) km, the zonal-mean wind of Full-O3 is slightly weaker than for Mean-O3 and No-O3; the zonal-mean wind shows little change between the experiments in the center of the transition region, that is, near 40 km.

Figure 3 also shows that the largest difference in the zonal-mean wind is between No-O3 and Full-O3 where the difference is about \(-3, 10, \) and 17 m s\(^{-1}\) at heights of 25, 50, and 60 km, respectively. In contrast, the difference between the mean wind of No-O3 and Mean-O3 is about 0, 5, and 8 m s\(^{-1}\) at the same heights. Thus within the photochemical control region, ZAO are largest (and positive) in the photochemical control region and smallest (and negative) in the dynamical control region. We discuss these sharp differences in mean wind response later in this section. For now, we simply note that ZMO and ZAO together cause the polar vortex to be stronger and colder above 40 km and slightly weaker and warmer below 40 km.

To understand the relative importance of planetary wave drag, ozone heating, and Newtonian cooling to the zonal-mean wind response shown in Fig. 3, we show in Fig. 4 the contributions from \( \bar{u}_{PWD} \), \( \bar{u}_{OH} \), and \( \bar{u}_{NC} \) [see Eq. (18) and Fig. 2]. Consider first the No-O3 simulation. As expected, \( \bar{u}_{PWD} \) and \( \bar{u}_{NC} \) balance at all heights (cf. Figs. 4a,b). This balance reflects two opposing mechanisms: the tendency of cooling due to longwave thermal emission \( \bar{u}_{NC} \) to relax the wind toward radiative equilibrium and the tendency of dynamical heating due to wave activity \( \bar{u}_{PWD} \) to drive the zonal-mean wind away from radiative equilibrium.

ZMO (Mean-O3) changes the steady-state zonal-mean wind [Eq. (18)] by modifying the zonal-mean ozone heating rate term \( \bar{u}_{OH} \) [Fig. 4c] within the photochemical control and upper transition regions (above \( \sim 40 \) km). The changes in \( \bar{u}_{OH} \) arise from photochemical production and destruction of zonal-mean ozone, where the ozone
production/destruction is a function of photochemical relaxation and temperature-dependent effects.

Adding ZAO (Full-O₃) further changes the steady-state zonal-mean wind by modifying \( \bar{u}_{\text{PWD}} \) and \( \bar{u}_{\text{OH}} \) (P1 and P2, respectively). Figure 4a shows that changes in the zonal-mean wind by \( \bar{u}_{\text{PWD}} \) are strongest above \( \sim 40 \) km. In contrast, ZAO produce wave–ozone flux divergences that affect \( \bar{u}_{\text{OH}} \) throughout the depth of the model (Fig. 4c).

Because of the strong altitude dependence of \( \bar{u}_{\text{PWD}}, \bar{u}_{\text{NC}}, \) and \( \bar{u}_{\text{OH}} \), we next consider the effect of ZMO and ZAO in the photochemically controlled upper stratosphere and lower mesosphere, and follow with a similar discussion for the dynamically controlled lower stratosphere.

(i) Photochemical control region

Within the photochemical control and upper transition regions (above \( \sim 40 \) km), the differences between the zonal-mean wind profiles of Mean-O₃ and Full-O₃ are mostly driven by changes in PWD, which operate along P1 (Fig. 1). Figure 4a shows that in the photochemical control region, \( \bar{u}_{\text{PWD}} \) is \( \sim 30\% \) smaller for Full-O₃ than for Mean-O₃. To understand how ZAO cause changes in \( \bar{u}_{\text{PWD}} \), we first recall from Eq. (1) that PWD is a function of wave propagation and wave damping. Here we measure wave propagation using the refractive index [Eq. (23)]; we measure wave damping using the planetary wave amplitude [Eq. (14a)].

Figure 5a shows that in the photochemical control and upper transition regions (above \( \sim 40 \) km), the refractive indices [Eq. (23)] of Mean-O₃ and Full-O₃ are nearly equal, meaning there is very little change in wave propagation between the two cases. The wave amplitudes of Mean-O₃ and Full-O₃, however, are noticeably different. Figure 5b shows that, for Full-O₃, ZAO produce a sharp decrease in the wave amplitude. This means that in the photochemical control and upper transition regions, wave damping is the controlling factor in the PWD and thus the primary mechanism for communicating ZAO to the zonal-mean circulation along P1. We provide further support for the primacy of P1 in driving the change in the zonal-mean circulation in the photochemical control region by returning to Fig. 4 and considering \( \bar{u}_{\text{NC}} \) and \( \bar{u}_{\text{OH}} \).

Figures 4b,c show that, in contrast to ZMO alone, including ZAO produces smaller values of \( \bar{u}_{\text{NC}} \) and \( \bar{u}_{\text{OH}} \) in the photochemical control and transition regions. The decrease in \( \bar{u}_{\text{OH}} \) is due to two processes (see Fig. 2); decreased production and destruction of ozone (Figs. 6a,b) and addition of wave–ozone flux divergence, \( \mathbf{V} \cdot \mathbf{M} \) (Fig. 6c). Nevertheless, if the contribution from \( \bar{u}_{\text{PWD}} \) is ignored, the smaller values of \( \bar{u}_{\text{NC}} \) and \( \bar{u}_{\text{OH}} \) resulting...
from ZAO would produce a weaker zonal-mean wind. Yet the zonal-mean wind of Full-O₃ is stronger and closer to radiative equilibrium than for Mean-O₃ (see Fig. 3), implying that the stronger zonal-mean wind of Full-O₃ must result from the decrease in $u_{PWD}$ due to ZAO, as described above. These results present two questions: does the relationship between wave amplitude and PWD seen in the photochemical control region hold for the dynamically controlled lower stratosphere and does knowledge of PWD alone provide a complete assessment of the impact of ZAO on the zonal-mean circulation? We consider these questions next.

(ii) Dynamical control region

To evaluate the response of the zonal-mean wind to ZAO in the dynamically controlled lower stratosphere (below ~35 km), we again consider the contributions to the steady-state zonal-mean wind [Eq. (18)]. The response of $u_{PWD}$ to ZAO in the dynamical control region is sharply different from that in the photochemical control region and transition regions (see Fig. 4a); yet, Fig. 5b shows that in the dynamical control region, ZAO decreases the wave amplitude by an amount similar to that seen in the photochemical control region. Why do $u_{PWD}$ and the zonal-mean wind show a relatively small response to the effects of full-ozone feedbacks (Full-O₃) in the dynamical control region, whereas the wave amplitude shows a relatively large response? As explained in the introduction, if wave propagation and wave damping are offsetting, it is possible to have a large change in wave amplitude but a relatively small change in PWD. A small change in PWD means a small change in the zonal-mean wind. To examine the possibility that the decrease in wave amplitude due to wave damping is offset by an increase in wave propagation, we again examine the refractive index [Eq. (23)].

Figure 5a shows that between ~15 and 45 km, which spans the dynamical control and transition regions, the refractive index of Full-O₃ is larger than for Mean-O₃. To interpret this change in refractive index, recall how ZAO changed the speed and curvature of the zonal-mean wind of the upper versus lower stratosphere (Fig. 3). The change in the zonal-mean wind between Full-O₃ and Mean-O₃ is larger in the photochemical control region than in the dynamical control region (~17 versus ~3 m s⁻¹). In contrast, the change in curvature of the wind profile...
between Full-O$_3$ and Mean-O$_3$ is more significant in the dynamical control region than in the photochemical control region; the result is the increase in wave propagation shown in Fig. 5a. Thus in the dynamical control region, ZAO cause an increase in propagation that is offset by an increase in wave damping; consequently, there is a negligible change in $\bar{u}_{P WD}$ and a much smaller change in the zonal-mean circulation than in the photochemical control region. Changes in $a_{P WD}$ do not, however, fully explain how the zonal-mean circulation of the lower stratosphere responds to ZAO. In fact, the slight decrease in $\bar{u}_{P WD}$ due to ZAO is inconsistent with the slight decrease in the zonal-mean wind of Full-O$_3$. Therefore, although ZAO modulate the propagation and damping characteristics of the planetary wave, $\bar{u}_{P WD}$ cannot be fully responsible for the change in the zonal-mean wind in the lower transition and dynamical control regions (below $\sim 40$ km).

To explain the change in the zonal-mean wind within the dynamical control region, we note that the inclusion of ZAO causes a slight increase in $\bar{u}_{OH}$ below $\sim 40$ km (Fig. 4c). Because there is very little change in $\bar{u}_{P WD}$ and $\bar{u}_{NC}$ (Figs. 4a and 4b, respectively), the increase in $\bar{u}_{OH}$ becomes the dominant process, and the zonal-mean wind of Full-O$_3$ is slightly weaker than that of Mean-O$_3$ (Fig. 3). In contrast to the photochemical control region where changes in $\bar{u}_{OH}$ were shown to be dominated by photochemical production and destruction of ozone (Figs. 6a,b), changes in $\bar{u}_{OH}$ in the lower transition and dynamical control regions are primarily governed by the divergence of wave--ozone fluxes ($\mathbf{V} \cdot \mathbf{M}$, Fig. 6c). Indeed, Figs. 6a and 6b show that the production and destruction of ozone is negligible below $\sim 40$ km; this reflects the increase in photochemical time scales with decreasing height. Thus the reduction of $\bar{u}_{OH}$ and the resultant decrease in the zonal-mean wind of Full-O$_3$ are primarily due to $\mathbf{V} \cdot \mathbf{M}$ along P2.

To summarize, we have presented evidence showing that $\bar{u}_{P WD}$ and $\bar{u}_{OH}$ (P1 and P2, respectively) both play a role in determining the final state of the zonal-mean wind. However, within the dynamically controlled lower stratosphere, ZAO caused offsetting changes in wave propagation and wave damping that produce a negligible change in PWD; therefore, using PWD as the sole diagnostic for evaluating the impact of ZAO on the zonal-mean circulation may yield the mistaken conclusion that ZAO are unimportant in the lower stratosphere. To gain further insight into the interaction between ZAO and the planetary wave field, we next examine the vertical energy flux diagnostic given by [Eq. (21)].

2) **Vertical Energy Flux**

Figure 7a shows that the total VEF for Full-O$_3$ is smaller than for Mean-O$_3$ from $\sim 10$ to 60 km; the largest decrease ($\sim 50\%$) occurs in the dynamical control and transition regions (below $\sim 45$ km). The decrease in VEF is predominantly driven by the large decrease in the northward heat flux (VEF$_{HF}$, Fig. 7b) between $\sim 20$ and 45 km; however, ozone heating (VEF$_{OH}$, Fig. 7c) plays a secondary role in the photochemical control region (above $\sim 45$ km). Newtonian cooling (Fig. 7d, VEF$_{NC}$) does not contribute to the decrease in the VEF because, when ZAO are included, VEF$_{NC}$ becomes less negative. Recall, $\bar{u}_{P WD}$ (Fig. 4a) shows the largest change due to ZAO in the photochemical control region. Clearly, VEF and PWD respond differently to ZAO. To reconcile this difference, consider the following.

The VEF$_{HF}$ [Eq. (22)] is proportional to the product of PWD and zonal-mean wind and inversely proportional to wave damping $m_i$. Within the dynamical control region, ZAO increase $m_i$ (reflected in the decreased wave amplitude; Fig. 5b) but cause little change in PWD and the zonal-mean wind; the result is reflected in VEF$_{HF}$, which is much smaller in Full-O$_3$ than in Mean-O$_3$. In the upper transition and photochemical control regions (above $\sim 40$ km), ZAO increase $m_i$ and decrease PWD (Fig. 4a). However, the changes in $m_i$ and PWD are offset by the larger zonal-mean wind of Full-O$_3$. As a result, the VEF$_{HF}$ of Full-O$_3$ and Mean-O$_3$ eventually converge near $\sim 55$ km. The significant decrease in the VEF caused by ZAO in the dynamical control region highlights the need to understand both wave propagation and wave damping when evaluating the ability of the lower stratosphere to modulate the circulation of the upper stratosphere and lower mesosphere.

The height ($\sim 30$ km) of the local maximum of the refractive index shown in Fig. 5a qualitatively agrees with Chen and Robinson (1992), who found that small changes in the strength ($\sim 1–3$ m s$^{-1}$) and curvature of the zonal-mean wind within the lower stratosphere may greatly alter the flux of wave activity into the middle and upper stratosphere. Chen and Robinson describe the shear of the zonal-mean wind in the upper troposphere and lower stratosphere as a valve, either enhancing or diminishing planetary wave propagation into the upper stratosphere. The VEF in Fig. 6 shows that, in addition to wave propagation, wave damping may also act as a filter or valve, modulating the amount of wave energy that can escape the lower stratosphere and affect the upper stratosphere and lower mesosphere. Yet, because ZAO alter both wave propagation and wave damping, the use of a single diagnostic to evaluate the effects of ZAO becomes problematic. PWD and VEF, therefore, provide complimentary views of the impact of ZAO on the middle atmosphere; namely, PWD shows that the net effect of ZAO on the zonal-mean circulation is largest in the upper stratosphere and lower
mesosphere, while VEF exposes the effect of ZAO on the planetary wave amplitude within the lower stratosphere.

5. Conclusions

Relatively little is known about the pathways that communicate the effects of ZAO to the wave-driven circulation in the middle atmosphere. In this study, we have provided a theoretical framework that identifies the two key pathways (P1 and P2) along which ZAO modulate the stratospheric circulation (see Figs. 1 and 2). Along P1, ZAO alter the propagation and damping (attenuation) of vertically propagating planetary waves; along P2, wave–ozone flux convergences alter the zonal-mean heating rate. Using a quasigeostrophic model that couples radiation, ozone, and dynamics, we have evaluated the relative importance of P1 and P2 using a combination of zonal-mean and wave diagnostics, which include expressions for ozone-modified planetary wave drag (NC07) and ozone-modified vertical energy flux.

Our results show that P1 and P2 both play a role in communicating the effects of ZAO to the stratosphere and lower mesosphere. The relative importance of each pathway depends on altitude or, more specifically, on the ratio of photochemical to dynamical time scales. Comparison of simulations with and without ZAO shows that P1 imparts the largest change in the zonal-mean circulation in the photochemically controlled upper stratosphere and lower mesosphere; this variability is predominantly driven by a reduction in planetary wave drag (PWD) above 40 km and is associated with P1. Diagnosing the importance of P1 solely by changes in PWD within the upper stratosphere and lower mesosphere, however, is potentially misleading.

In particular, our analysis of the ozone-modified PWD reveals that, within the lower stratosphere, ZAO cause wave propagation and wave damping to oppose each other. The result is a small change in PWD [see Eq. (1)], but a large reduction in wave amplitude. Thus in the dynamically controlled lower stratosphere, ZAO “precondition” the wave before it propagates into the upper stratosphere where damping due to photochemically accelerated cooling dominates, causing a large reduction in PWD and thus a colder polar vortex. The preconditioning of the wave by ZAO in the lower stratosphere may explain the GCM
results obtained by Gabriel et al. (2007). They found that temperature changes in the lower mesosphere are induced indirectly by changes in planetary wave dynamics, despite the ZAO being prescribed only in the lower stratosphere of their GCM simulations.

The preconditioning of the wave in the lower stratosphere is also evidenced in our analysis of an ozone-modified vertical energy flux (VEF). Our numerical simulations show that ZAO cause a substantial decrease in zonal-mean wave energy arriving in the upper stratosphere and lower mesosphere; this decrease in the VEF is dominated by contributions from the northward heat flux (VEF_{HF}). This result confirms the previously untested hypothesis asserted by Crook et al. (2008) and Gillett et al. (2009) that ZAO cause a decrease in VEF that is at least partly responsible for changes in the polar vortex in their GCM experiments. Moreover, the VEF_{HF} clearly shows the decrease in wave amplitude within the lower stratosphere that cannot be discerned by evaluating PWD alone. Thus, while VEF is not a direct measure of the net effect of the planetary wave on the zonal-mean circulation, it may provide a better measure of the effect of ZAO on the planetary wave itself within the dynamically controlled lower stratosphere. This point has important implications for researchers attempting to diagnose the impact of ZAO in climate models: namely, PWD and VEF should be considered in tandem so as to fully expose the effects of ZAO along P1. While the PWD and VEF diagnostics indicate that P1 is the dominant pathway for zonally asymmetric ozone-driven variability, P2 also plays an important role.

Our numerical results show that, within the photochemically controlled upper stratosphere, wave–ozone fluxes (V · M) along P2 increase the strength of the zonal-mean wind. In contrast, V · M produces a heating effect in the dynamical control and transition regions; this heating is responsible for the weaker zonal-mean wind seen in the lower stratosphere when ZAO are included in the model. The nature and significance of ZAO in modulating transport processes was studied in detail by Hartmann and Garcia (1979), Garcia and Hartmann (1980), and Kirchner and Peters (2003). However, Garcia and Solomon (1983) found that one-dimensional models, including the 1D Garcia and Hartmann (1980) model, underestimate the magnitude of ozone transports due to planetary waves; thus, it is plausible that the model employed in this study actually underestimates the impact of P2 on the zonal-mean circulation. Nevertheless, when ZAO are included in the model, the combined effects of P1 and P2 lead to a stronger and colder polar vortex in the upper stratosphere and lower mesosphere and a slightly weaker and warmer polar vortex in the lower stratosphere.

Several issues of importance to stratospheric climate variability have not been addressed within this study. For example, the Northern Hemisphere lower stratosphere has undergone both secular ozone losses (WMO 2007) and episodic ozone losses (Jin et al. 2006). Yet, it remains unclear what role ZAO may play in these losses. This issue may be of particular importance in light of the ability of ZAO to alter the propagation and damping of vertically propagating planetary waves as they pass upward through the lower stratosphere and into the upper stratosphere and lower mesosphere (P1). In this way, the preconditioning of planetary waves along P1 may play an important role in communicating lower-stratospheric ozone losses to the entire middle atmosphere.

Our study relied on a one-dimensional (in height) dynamical–radiative–photochemical model. The next step is to extend our mechanistic model to allow for wave propagation in the latitude–height plane, which has been shown to play an important role in ozone-induced changes in planetary wave activity (Hu and Tung 2003). In particular, meridional wave propagation and meridional momentum fluxes may play an important role in offsetting or reinforcing the propagation and damping of planetary waves by ZAO along P1. Despite the limitations of our model, the pathways that we have identified in Fig. 1 provide a clear and general description of the physical mechanisms involved in zonally asymmetric ozone-driven climate variability.

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APPENDIX

Radiative–Photochemical Model

In the absence of dynamics, equations for the ozone volume mixing ratio and temperature are, respectively (Lindzen and Goody 1965; Nathan and Li 1991),

\[
\frac{\partial \gamma}{\partial t} = 2\frac{J_i n_s}{n_m} - 2\frac{J_3 k_3}{k_2 n_2} \gamma^2, \quad (A1)
\]

\[
\frac{\partial T}{\partial t} = \frac{q_3 N_0}{m_a c_p} \gamma + b - \alpha T, \quad (A2)
\]

where \(J_i\), \(n_s\), and \(k_i\) are the photodissociation rate, number density, and temperature-dependent reaction rate for molecular oxygen \((i = 2)\), ozone \((i = 3)\), and air \((i = m)\); \(q_3\) is the heating rate due to the absorption of solar radiation by ozone; the Newtonian cooling...
coefficients, $\alpha$ and $b$, are taken from Dickinson (1973) and Blake and Lindzen (1973), respectively; $m_a$ is the molecular weight of air; $c_p$ is the gas constant; and $N_0$ is Avogadro’s number. The temperature-dependent reaction rate for ozone $k_3$ has been modified to account for catalytic destruction of odd oxygen by hydrogen, nitrogen, and chlorine chemistry (Haigh and Pyle 1982).

Ozone heating and photodissociation rates are given by

$$q_3 = \int_0^\infty F(\nu)\sigma_3(\nu) \exp[-\sigma_3(\nu)x_3(z) - \sigma_2(\nu)x_2(z)] \, dv,$$

(A3)

$$J_i = \int_0^\infty \{F(\nu)\sigma_i(\nu) \exp[-\sigma_3(\nu)x_3(z) - \sigma_2(\nu)x_2(z)]/h\nu\} \, dv,$$

(A4)

respectively, where $F(\nu)$ and $\sigma_i(\nu)$ are the solar flux of photons at the top of the atmosphere and the absorption cross section at frequency $\nu$, respectively, for oxygen ($i = 2$) and ozone ($i = 3$), and

$$\overline{\sigma_i} = \frac{1}{\cos \mu} \int_0^\infty \tilde{n}_i \, dz' = \frac{1}{\cos \mu} \frac{N_0}{m_a} \int_0^\infty \rho(z') \overline{\tilde{g}_i}(z') \, dz'$$

(A5)

is the slant path column density of molecular oxygen ($i = 2$) and ozone ($i = 3$) at solar zenith angle $\mu$. The net ozone heating and photodissociation rates for the Chappuis band are then calculated as

$$q_C = \sum_i \varepsilon_i F_C \sigma_C \exp(-\sigma_C \overline{\sigma_i})$$

(A6)

and $J_C = q_C/(h\nu_C)$, where $\varepsilon$ is the efficiency factor. Similar calculations are performed for the Hartley, Huggins, and Herzberg bands yielding for the total ozone heating and photodissociation rates $q_3 = q_C + q_{Ha} + q_{Hd}$ and $J_3 = J_C + J_{Ha} + J_{Hd}$. Photodissociation of molecular oxygen due to the Schumann–Runge band is calculated using the zenith-angle-dependent effective cross sections of Allen and Frederick (1982) together with the solar flux tabulations from WMO (1985). Temperature-dependent absorption cross sections (Molina and Molina 1986) are used from 185 to 350 nm, and WMO (1985) cross sections are used otherwise. Enhancement of solar radiation due to multiple scattering, surface reflection, clouds, and aerosols is taken into account using the approximation of Meier et al. (1982) and Nicolet et al. (1982). The heating and photodissociation rates are then diurnally averaged (Cunnold et al. 1975; Cogley and Borucki 1976). We obtained data for the initial temperature profile from Randel et al. (2002); ozone profiles were obtained from Keating et al. (1996). We use the analytical radiative equilibrium temperature expression from Kushner and Polvani (2004, 2006). For our experiments, we have set $\gamma = 0.75 \, K \, km^{-1}$, $\phi_{0N} = 60^\circ$, and $p_T = 100 \, mb$ in their Eqs. (1)–(4) and (1)–(3), respectively.

Expressions for the ozone heating coefficients, Eqs. (5) and (12), and ozone production and destruction coefficients, Eqs. (6) and (13), are determined by linearizing Eqs. (A1) and (A2), yielding the heating rate coefficients

$$\Gamma_1 = \frac{q_3 N_0}{m_a}$$

(A7)

and

$$\Gamma_2 = \frac{L_i N_0^2 \rho_0}{m_a \cos \mu} \tilde{\gamma},$$

(A8)

and ozone production and destruction coefficients

$$\xi_1 = -\frac{4C_1 J_3}{n_2} \exp(-K/T),$$

(A9)

$$\xi_2 = \frac{N_0 \rho_0}{m_a \cos \mu} \left(-2\gamma_2 L_2 + \frac{\tilde{\xi}_1 \gamma_3}{2 J_3}\right),$$

(A10)

$$\xi_T = \frac{\tilde{\xi}_1 \gamma}{2T} \left(1 + \frac{K}{T}\right),$$

(A11)

where $q_3$ and $J_3$ are defined in Eqs. (A3) and (A4); $C_1 = 3.18 \times 10^{22} \, mol \, cm^{-3}$; $\gamma_2$ is the mixing ratio of molecular oxygen; and $L_i$ is defined as

$$L_i(z, t) = \int_0^\infty \{F(\nu)\sigma_i(\nu) \exp[-\sigma_3(\nu)x_3(z) - \sigma_2(\nu)x_2(z)]/h\nu\} \, dv,$$

(A12)

where $h$ is Planck’s constant. The temperature-dependent reaction rate $K$ is defined following the approximation of Haigh and Pyle (1982):

$$K = \frac{a^* + b^* f_1 + c^* f_2 + d^* f_3}{2 + f_1 + f_2 + f_3},$$

(A13)

where $f_1, f_2$, and $f_3$ are the ratios of the rate of destruction of odd oxygen due to catalytic cycles involving HO, NO, and ClO, respectively, to the rate of destruction due to oxygen-only reactions, and $a^* = 2570 \, K$, $b^* = 630 \, K$, $c^* = 1400 \, K$, and $d^* = 260 \, K$ (Sander et al. 2006).

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