Stationary waves weaken and delay the near-surface response to stratospheric ozone depletion

Chaim I. Garfinkel, a Ian White, a Edwin P. Gerber, b Seok-Woo Son, c Martin Jucker, d

a The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra
Campus, Givat Ram, Jerusalem, Israel
b Courant Institute of Mathematical Sciences, New York University, New York, USA
c School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea
d Climate Change Research Centre and ARC Centre of Excellence for Climate Extremes,
University of New South Wales, Sydney, Australia

Corresponding author: Chaim I. Garfinkel, The Hebrew University of Jerusalem, Institute of Earth Sciences, Edmond J. Safra Campus, Givat Ram, Jerusalem, Israel, chaim.garfinkel@mail.huji.ac.il.
ABSTRACT: An intermediate complexity moist General Circulation Model is used to investigate the factor(s) controlling the magnitude of the surface impact from Southern Hemisphere springtime ozone depletion. In contrast to previous idealized studies, a model with full radiation is used, and further, the model can be run with a varied representation of the surface, from a zonally uniform aquaplanet to a highly realistic configuration. The model captures the positive Southern Annular Mode response to ozone depletion evident in observations and comprehensive models in December through February. It is shown that while synoptic waves dominate the long-term poleward jet shift, the initial response includes changes in planetary waves which simultaneously moderate the polar cap cooling (i.e., a negative feedback), but also constitute nearly half of the initial momentum flux response that shifts the jet polewards. This model also can disentangle the role of surface cooling for the jet response, and it is found that surface cooling helps strengthen the poleward shift. The net effect is that stationary waves weaken the circulation response to ozone depletion, and also delay the response until summer rather than spring when ozone depletion peaks. Finally, the surface response is shown to be linear with respect to the magnitude of the imposed stratospheric perturbation.
1. Introduction

Antarctic springtime ozone concentrations decreased in the last few decades of the twenty-first century due to anthropogenic emissions of chlorofluorocarbons (Solomon et al. 1986), and only recently have begun the slow process of recovery (Weber et al. 2018). Ozone depletion is known to have been the dominant contributor over the late 20th century to a poleward shift of the austral summer Southern Hemisphere (SH) tropospheric midlatitude jet, Southern Annular Mode (SAM), precipitation, and storm tracks, and to have led to an expansion of the summer Hadley Cell (Trenberth and Stepaniak 2002; Gillett and Thompson 2003; Son et al. 2010; Thompson et al. 2011; Kang et al. 2011; Polvani et al. 2011; McLandress et al. 2011; Eyring et al. 2013; Gerber and Son 2014; Gonzalez et al. 2014; Previdi and Polvani 2014; Waugh et al. 2015; Seviour et al. 2017; Son et al. 2018). Over the next ~50 years, ozone recovery is expected to nearly cancel out changes in the tropospheric jet and Hadley Cell that would otherwise be forced by greenhouse gases (Son et al. 2008; Polvani et al. 2011; Arblaster et al. 2011; Barnes and Polvani 2013; Gerber and Son 2014; Banerjee et al. 2020). Despite this clear evidence, the mechanism whereby ozone depletion leads to a downward impact, and the details of how this mechanism governs the magnitude of the impact, are still unclear, e.g. as noted in successive WMO Ozone assessments (World Meteorological Organization 2011, 2014; Karpechko et al. 2018).

This study focuses on the role of stationary vs. transient waves for the downward impact. While SH stationary waves are weaker than their counterparts in the Northern Hemisphere, they contribute roughly half of the heat flux in spring in the lower stratosphere (Källberg et al. 2005) and contribute to the inter-model spread in the timing of the ozone-hole breakup (Hurwitz et al. 2010). A commonly used model in studies focusing on the mechanism(s) for the surface response to ozone depletion is a dry dynamical core with a flat bottom (e.g. Kushner and Polvani 2004; Sun et al. 2014; Yang et al. 2015; Smith and Scott 2016) allowing for transient planetary waves only, or a highly idealized mountain (Gerber and Polvani 2009; Domeisen et al. 2013). The importance of stationary waves in the SH for a surface response cannot be readily evaluated in such setups by construction. Many of these studies using flat-bottomed models nevertheless conclude that planetary waves are crucial for the surface response. For example, Smith and Scott (2016) find that the response to a
stratospheric perturbation is weaker if interactions between planetary- and synoptic-scale waves are suppressed, while Domeisen et al. (2013) find that the jet shifts in the opposite direction if only planetary waves are present, ruling out the possibility that the jet shift occurs purely as a response to changes in the planetary- or synoptic-scale wave fields alone. However the lack of stationary planetary waves in these models resembling those in the SH may lead to a mis-representation of the total impact of planetary waves. The goal of this study is to answer this question: what is the relative role of synoptic vs. planetary waves for the downward impact?

A secondary goal of this study is to disentangle the role of the surface temperature cooling in response to ozone depletion for the jet response. The SAM response appears to account for around half of the observed surface warming over the Antarctic Peninsula, nearly all of the observed cooling over East Antarctica, and much of the warming over Patagonia (Trenberth and Stepaniak 2002; Previdi and Polvani 2014). Nevertheless, radiative effects may also be important for the tropospheric (Grise et al. 2009) and the surface temperature (Yang et al. 2014) response to ozone depletion, though Chiodo et al. (2017) found the net radiative effect at the surface to be weak. Regardless of how the tropospheric cooling arises, the role of this tropospheric cooling for the jet shift, as compared to other mechanisms for the downward impact, has not been isolated in previous work.

We take advantage of a recently developed intermediate complexity model that can delineate the role of different waves types and of surface cooling. Namely, it can be run alternately with realistic stationary waves or without any zonal asymmetry in the bottom boundary (e.g., topography), and thus clarify the role of stationary waves for the surface response. This model also allows us to carefully isolate the importance of surface temperature changes in response to ozone depletion by studying the jet response for different surface albedos over Antarctica.

After introducing this model in Section 2 and our diagnostics in Section 3, we demonstrate in Section 4 that the model in its most realistic configuration simulates a quantitatively realistic response to ozone depletion, but that the response is significantly stronger in an aquaplanet configuration. We consider reasons for this effect in Section 5, isolate the role of
surface cooling in Section 6, and then summarize our results and place them in the context of previous work in Section 7.

2. A model of an idealized moist atmosphere (MiMA)

We use the Model of an idealized Moist Atmosphere (MiMA) introduced by Jucker and Gerber (2017), Garfinkel et al. (2020b), and Garfinkel et al. (2020a). This model builds on the aquaplanet models of Frierson et al. (2006), Frierson et al. (2007), and Merlis et al. (2013). Very briefly, the model solves the moist primitive equations on the sphere, employing a simplified Betts-Miller convection scheme (Betts 1986; Betts and Miller 1986), idealized boundary layer scheme based on Monin-Obukhov similarity theory, and a purely thermodynamic (or slab) ocean. An important feature for this paper is that we use a realistic radiation scheme Rapid Radiative Transfer Model (RRTMG) (Mlawer et al. 1997; Iacono et al. 2000), which allows us to explicitly simulate the radiative response to ozone depletion, unlike previous studies using more idealized models with Newtonian cooling. Please see Jucker and Gerber (2017) for more details.

This model can be run alternately as an aquaplanet, or with stationary waves quantitatively similar to those in comprehensive models (Garfinkel et al. 2020b,a). The most realistic configuration of MiMA used in this study has boundary forcings that are identical to those of Garfinkel et al. (2020a), and this configuration is referred to as STAT in the rest of this paper. MiMA has no true land, rather the properties of the surface at gridpoints that are land on Earth are modified to mimic land (Figure 3 of Jucker and Gerber 2017). The net effect is that the STAT configuration includes three sources of zonal asymmetry in the lower boundary: orography, prescribed east-west ocean heat transport, and land-sea contrast (i.e., difference in heat capacity, surface friction, and moisture availability between “ocean” gridpoints and “land” gridpoints). The specifications of these forcings can be found in Garfinkel et al. (2020a). Note that the same albedo value is applied to all wavelengths of incoming solar radiation.

We analyze the response to an identical ozone hole for four different tropospheric configurations: (i) the Southern Hemisphere (SH) of STAT, (ii) the Northern Hemisphere (NH) of STAT (STATNH), (iii) an aquaplanet with albedo of 0.27 globally (including over “Antar-
tica"), and (iv) and an aquaplanet but in which the albedo over “Antarctica" is increased to
0.8 and elsewhere lowered to 0.23 (as in STAT, see equation A3 of Garfinkel et al. 2020a) to
help maintain a similar global mean and “Antarctic" temperature to STAT. We refer to these
last two experiments as AQUA27 and AQUA80 in the rest of this paper. The AQUA runs
have no stationary waves, but both aquaplanet integrations still include north-south ocean
heat transport (Eq. A4 of Garfinkel et al. 2020a). The aquaplanet runs use a mixed-layer
depth of 75m everywhere, in contrast to STAT which has a mixed layer depth of 2.5m over
“land" and a varying depth for ocean gridpoints (see Eq. A2 of Garfinkel et al. 2020a).
The NH STAT configuration is not meant to simulate a boreal winter ozone “hole", either as
observed in 1997, 2011 or 2020 (Hurwitz et al. 2011; Manney et al. 2011; Rao and Garfinkel
2020; Lawrence et al. 2020; Rao and Garfinkel 2021) or as in a world avoided scenario (New-
man et al. 2009; García et al. 2012). Rather, it explores how the exact same change of ozone
impacts the circulation with a very different climatology of stationary (and synoptic) waves.

For all tropospheric configurations, we compare a pair of simulations: (1) a preindustrial
simulation forced with the monthly varying climatology of ozone in the CMIP6 ozone speci-
fication averaged from 1860 to 1899 (PI simulation; Checa-García et al. 2018; Checa-García
2018); and (2) a simulation forced with the monthly varying climatology of ozone in the
CMIP6 ozone specification averaged from 1990 to 1999, which we then further reduce by a
factor of 4 between 150hPa and 30hPa and poleward of 65S following:

\[ \Phi(\varphi) = 1 - \frac{3}{8} \left(1 - \tanh \left(\frac{\varphi + 65^\circ}{3^\circ}\right)\right), \]  

(1)

where \( \varphi \) denotes latitude (ozone hole simulation). This additional reduction in the polar
lower stratosphere is intended to capture springs with stronger than average ozone deple-
tion (Previdi and Polvani 2014), and is included to enhance the signal to noise ratio. An
experiment without this additional reduction leads to a weaker surface response, which is
consistent with previous work that has argued that interannual variability of ozone concen-
trations can be used to improve the skill of seasonal and subseasonal forecasting (Son et al.
2013; Bandoro et al. 2014; Hendon et al. 2020; Jucker and Goyal 2021). The linearity of the
response is discussed in more detail in Section 5b.
The ozone hole runs branch from October 1st of the last 65 years of the respective preindustrial control runs for a total of 65 ensemble members, and are then integrated for at least 150 days. The results are shown in terms of the difference between the ozone hole simulation and the PI simulation (ozone hole - PI), though all conclusions are just as applicable to ozone recovery. The net effect on ozone is shown in Figure 1abc, which show days 1 to 30 (October), 31 to 70 (November and early December), and 71 to 120 (rest of December and January). The ozone perturbation is evident throughout the spring and decays in early summer. In the polar lower stratosphere, more than 90% of the preindustrial ozone is locally depleted, and this reduction is within the range of realistic values (Solomon et al. 2005; Previdi and Polvani 2014). Ozone actually increases slightly in the upper stratosphere in summer due to dynamical feedbacks (Stolarski et al. 2006).

In order to isolate any effect of ozone on surface shortwave absorption (Chiodo et al. 2017), and also more cleanly connect our results to studies using dry models with an imposed diabatic heating (Kushner and Polvani 2004; Sheshadri and Plumb 2016; White et al. 2020), we also performed simulations in which a diabatic heating perturbation is imposed in the lower stratosphere. Our goal is to match the stratospheric diabatic heating perturbation due to ozone, and thus we show in Figure 1d-f the diabatic heating perturbation due to the reduced ozone as computed by the model. The diabatic heating rate is \(-0.5\)K/day in the polar lower stratosphere. The upper stratospheric diabatic tendency is due to the dynamically induced warming resulting in enhanced longwave emission (Manzini et al. 2003; McLandress et al. 2010; Orr et al. 2012a). Motivated by this, we impose a diabatic perturbation between 150hPa and 30hPa of the form of equation 1, and hold it constant in time with no seasonality. The effect of this diabatic heating perturbation is explored both for a diabatic heating perturbation similar in magnitude and location to the one due to ozone depletion (peaking at -0.5K/day; DIAB simulation) and also a factor of five larger (peaking at -2.5K/day; DIAB5x simulation). Note that the net effect on the stratospheric vortex of the \(-0.5\)K/day perturbation is slightly weaker than the corresponding ozone hole depletion run, as the \(-0.5\)K/day perturbation in turn leads to a negative feedback: cooler lower stratospheric temperatures lead to less longwave emission.
Table 1 summarizes all experiments included in this paper. For all integrations, the model is forced with $CO_2$ concentrations fixed at 390ppmv and seasonally varying solar insolation. All simulations in this paper were run with a triangular truncation at wavenumber 42 (T42) with 40 vertical levels.

Table 1. MiMA Experiments, with “Y” indicating a forcing is on and “N” indicating a forcing is off. For ozone, we compare a “preindustrial” simulation using ozone concentrations from the CMIP6 read-in file over the years 1860-1899 to a simulation using ozone concentrations from the CMIP6 read-in file over the years 1990-1999, which were then modified in the Antarctic lower stratosphere (see section 2) to capture a deeper ozone hole evident in some years. The jet latitude is included for November in the SH.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Surface zonal structure</th>
<th>“Antarctica” albedo</th>
<th>“Antarctica” mixed layer</th>
<th>Nov jet latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>STAT, hole-PI</td>
<td>Y</td>
<td>0.8</td>
<td>2.5m</td>
<td>50.1S</td>
</tr>
<tr>
<td>AQUA80, hole-PI</td>
<td>N</td>
<td>0.8</td>
<td>75m</td>
<td>46.5S</td>
</tr>
<tr>
<td>AQUA27, hole-PI</td>
<td>N</td>
<td>0.27</td>
<td>75m</td>
<td>43.1S</td>
</tr>
<tr>
<td>STATNH, hole-PI</td>
<td>Y</td>
<td>0.8</td>
<td>2.5m</td>
<td></td>
</tr>
<tr>
<td>STAT, diab-PI</td>
<td>Y</td>
<td>0.8</td>
<td>2.5m</td>
<td>50.1S</td>
</tr>
<tr>
<td>AQUA80, diab-PI</td>
<td>N</td>
<td>0.8</td>
<td>75m</td>
<td>46.5S</td>
</tr>
<tr>
<td>STAT, diabox-PI</td>
<td>Y</td>
<td>0.8</td>
<td>2.5m</td>
<td>50.1S</td>
</tr>
<tr>
<td>AQUA80, diabox-PI</td>
<td>N</td>
<td>0.8</td>
<td>75m</td>
<td>46.5S</td>
</tr>
</tbody>
</table>

3. Diagnostics

The role of synoptic and planetary waves in driving the poleward jet shift is diagnosed using the Eulerian mean zonal momentum budget:
\[
\frac{\partial \bar{u}}{\partial t} = -\left(\frac{1}{a \cos^2 \varphi} \frac{\partial}{\partial \varphi} (\cos^2 \varphi \bar{u}'_{k \leq 3}) + \frac{1}{\rho_0} \frac{\partial}{\partial z} (\rho_0 \bar{u}'_{k < 3})\right)_{edd \, 1-3} - \left(\frac{1}{a \cos^2 \varphi} \frac{\partial}{\partial \varphi} (\cos^2 \varphi \bar{u}'_{k > 3}) + \frac{1}{\rho_0} \frac{\partial}{\partial z} (\rho_0 \bar{u}'_{k > 3})\right)_{edd \, 4+} + \int_{f_0} - \frac{\partial \bar{u}}{\partial z} + \frac{\partial}{\partial \varphi} (\bar{u} \cos \varphi) \right)_{advect} + X + \text{res} \tag{2}
\]

(e.g., Andrews et al. 1987; Hitchcock and Simpson 2016) where the acceleration of the zonal-mean zonal wind on the left hand side is contributed to by processes associated with (from left to right on the right hand side): eddy momentum flux convergence due to planetary waves (\textit{edd} 1-3), eddy momentum flux convergence due to synoptic waves (\textit{edd} 4+), Coriolis torques acting on the meridional motion (\textit{f }), mean flow momentum advection (\textit{advect}), and parameterised processes including the zonal wind tendency due to vertical and horizontal diffusion and gravity-wave drag in the model (\textit{X}). All variables follow standard notation (e.g., see Andrews et al. 1987). The final term (\textit{res}) is the budget residual and is contributed to by issues associated with sampling and truncation errors.

The Southern Annular mode (SAM) and the e-folding timescale of the corresponding principle component timeseries is computed following the methodology of Baldwin et al. (2003) and Gerber et al. (2008). Jet latitude is computed by fitting the 850hPa zonal mean zonal wind near the jet maxima (as computed at the model’s T42 resolution) to a second order polynomial, and then evaluating the polynomial at a meridional resolution of 0.12°. The latitude of the maximum of this polynomial is the jet latitude (Garfinkel et al. 2013a).

4. The response to an identical ozone perturbation in STAT and in AQUA80

We begin by showing that in the STAT configuration of MiMA, ozone loss leads to impacts similar to those shown in previous works using reanalysis or comprehensive models. Figure 1ghi shows the temperature response to reduced ozone. Temperatures in the polar lower stratosphere gradually decrease over the first two months and reach -15K by November, and
Fig. 1. Zonal-mean responses to ozone loss [i.e., ozone hole minus preindustrial (PI)] in the most realistic configuration, STAT, in (left) days 1-30 after branching, i.e. October; (middle) days 31 to 70, i.e. November and December 1-10; (right) days 71 to 120, i.e. December 11 through January 30. (a-c) ozone perturbation; (d-f) diabatic heating rate computed as the sum of the temperature tendency due to longwave, shortwave, and latent heat release; (g-i) temperature; (j-l) zonal wind. The bottom row is as in (j) through (l) but for an aquaplanet configuration with "Antarctic" albedo=0.8. Stippling indicates anomalies statistically significant at the 95% level. The zero contour is shown in magenta.
Fig. 2. Map view of ozone loss response (ozone hole - PI) in the most realistic configuration in (left) days 1-30 after branching, i.e. October; (middle) days 31 to 70; (right) days 71 to 120. (a-c) geopotential height at 500hPa; (d-f) precipitation; (g-i) zonal wind at 850hPa. Stippling indicates anomalies statistically significant at the 95% level.

The anomaly propagates downward to near the tropopause in late December (Figure 1i). This cooling is similar to that observed during years with a particularly strong ozone hole relative to 1960s conditions (Randel et al. 2009; Previdi and Polvani 2014). The zonal wind response is shown in Figure 1jkl, and captures the response evident in reanalysis, CMIP, and CCMi data (Previdi and Polvani 2014; Son et al. 2018). Changes in 500hPa geopotential height also resemble the canonical SAM pattern (Figure 2bc, Kidson 1988; Thompson and Wallace 2000; Thompson et al. 2011) with lower heights in subpolar latitudes and higher heights between 40S and 50S. The model also simulates the precipitation response to
ozone depletion (unlike dry models used in many mechanistic studies). Figure 2def shows an increase in precipitation over Southeastern Australia and Southeastern South America and drying over New Zealand (in agreement with observed trends; Hendon et al. 2007; Ummenhofer et al. 2009; Gonzalez et al. 2014).

The increase in subpolar zonal wind peaks near day 90 at 70hPa (January 1st; Figure 3a), though higher in the stratosphere the response peaks earlier, and is followed by a zonal wind and SAM response in the troposphere (Figure 3b for 850hPa wind and 3c for geopotential height). While a tropospheric response begins to develop in November, the response peaks in January and persists into February in agreement with previous work.

Encouraged by the quantitative accuracy of the response in the most realistic configuration, we now take advantage of the flexibility of the idealized model in order to understand the role of stationary waves for the surface response. As discussed in Section 2, the same ozone perturbation has also been imposed in two aquaplanet configurations of the model (differing only in the polar albedo) and in the Northern Hemisphere. We begin with the aquaplanet configuration with a polar albedo of 0.8 (AQUA80), as this turns out to be the tropospheric configuration with the largest surface response to ozone depletion. Even though the ozone perturbations are identical, the wind response (Figure 1, bottom row) is larger in AQUA80. The difference in zonal wind response between the two configurations is statistically significant at the 5% level after day 45 in both the stratosphere and troposphere (Figure 4). The geopotential height response in the troposphere is more than twice as large in AQUA80 than in STAT (Figure 2abc vs 5abc and Figure 3c vs. 6c), and the precipitation response is also stronger although less regionally focused due to the lack of Antarctic orography (Figure 5def). The difference in response is evident both in November and in December/January (Figure 6b and Figure 7abc).

5. Why the stronger response for AQUA80 than STAT?

a. A stationary wave negative feedback for the jet response

Adding stationary waves leads to less of a cooling of the polar lowermost stratosphere in STAT relative to AQUA80 (Figure 7a) and less of a vortex strengthening (Figure 4c) even though the ozone perturbation is identical. This difference in response to an identical
Fig. 3. Development and downward propagation of the response to the ozone perturbation in the most realistic configuration. (a) 70hPa zonal wind; (b) 850hPa zonal wind; (c) 850hPa polar cap geopotential height; upper tropospheric meridional Eliassen-Palm flux due to (d) planetary and (e) synoptic waves. The tropospheric jet latitude is shown in (a) and (b) with gray diamonds. Stippling indicates anomalies statistically significant at the 95% level.

The ozone perturbation occurs because the strengthened vortex in late fall and early summer due to ozone depletion favors more upward wave propagation, and the subsequent enhanced wave convergence within the stratosphere leads to dynamical warming of the polar cap via downwelling of the transformed Eulerian mean vertical wind. This cancels a part of the radiatively driven cooling near the tropopause (not shown, but as in Manzini et al. 2003; Li et al. 2010; McLandress et al. 2010; Orr et al. 2012a). However this increase in upward propagating waves is more dramatic in the presence of stronger wave forcing from below, and in STAT these upward propagating waves are indeed stronger due to the presence
Fig. 4. Evolution of zonal wind from 54S to 80S for the ozone hole-PI runs with (a) realistic stationary waves, (b) an aquaplanet, with “Antarctic“ albedo equal to 0.8. (c) difference between (a) and (b). The contour interval is 2m/s in (a) and (b) and 0.5m/s in (c). The 1m/s contour is indicated in red in (a) and (b). Stippling indicates anomalies statistically significant at the 95% level.

of stationary waves. We demonstrate this effect in Figure 7d, which shows the vertical component of the Eliassen-Palm flux at 40hPa, though other levels in the mid- and lower-stratosphere show a similar response. In STAT, an ozone hole leads to increased upward wave flux by late October, and the anomaly stays positive throughout the duration of the run. The increase in AQUA80 is weaker however, and the difference between STAT and AQUA80 is statistically significant between days 75 and 90 though if we time average the significance begins earlier. The net effect is a warmer polar stratosphere and less accelerated vortex in STAT (Figure 7a and Figure 4c). Hence, stationary waves act as a negative feedback on the surface and stratospheric response to ozone, acting to partially offset the ozone-induced cooling and poleward jet shift.

This negative feedback caused by the presence of stationary waves can be further demonstrated by imposing the same ozone hole in the Northern Hemisphere. We compare the response of subpolar zonal wind in the (y-axis) lower stratosphere and (x-axis) lower troposphere in Figure 8a. Consider STAT (blue line); the average wind anomaly for days 61 to
Fig. 5. As in Figure 2 but for an aquaplanet configuration with “Antarctic" albedo=0.8. Note that the color scale for the top row differs from Figure 2. Continental outlines are included for reference only.

75 is 8.8m/s at 70hPa and 1.0m/s at 850hPa, whereas in the two AQUA runs the wind responses are stronger (for AQUA80 in black, 10.8m/s at 70hPa and 2.0m/s at 850hPa). The corresponding changes for the NH (in gold) are much weaker both in the lower stratosphere and troposphere despite cooling aloft (4.0m/s and 0.3m/s respectively). The net effect is that stationary waves (of which there is more activity in the NH) help damp the surface response to ozone depletion.

It is important to note that while stationary waves damp the surface response, transient planetary waves help contribute to the surface response, in agreement with Smith and Scott (2016). We demonstrate this by considering the Eulerian mean momentum budget for
Fig. 6. As in Figure 3 but for aquaplanet with “Antarctic” albedo=0.8.

AQUA80 which only contains transient planetary waves. The zonal wind tendency calculated explicitly is shown in Figure 9abc, and the various terms in the budget (equation 2) are shown in the rest of Figure 9. Figure 9def shows the sum of all terms on the right-hand size of equation 2, which should be equal to the zonal wind tendency in Figure 9abc. This is indeed the case, as the budget closes in nearly all regions, though some of the fine-scale details of the wind tendencies differ. The dominant terms are the eddy forcing term (Figure 9ghi) and the coriolis torque (Figure 9jkl), with the acceleration in most regions and lags provided by the eddy forcing term. The sum of the eddy forcing and coriolis terms (Figure 9mno) already resembles the total tendency in most regions/lags (Figure 9def), but crucially in the mid- and upper-stratosphere changes in gravity wave absorption act as a negative feedback in days 31 to 70 (late spring), and dominate the response in days 71 to 120 (summer). In other words, the zonal wind anomaly peaks in December before weakening in January and
Fig. 7. Summary of responses to ozone depletion [ozone hole-P1]. (a) polar cap temperature at 250hPa [K]; area-weighted average of zonal wind from 80S to 55S [m/s] at (b) 70hPa and (c) 850hPa; (d) vertical component of the EP flux at 40hPa area-weighted average from 80S to 45S [k /s²]. Blue and magenta lines are for most realistic configuration with an ozone perturbation and diabatic perturbation respectively. Red line is for an aquaplanet with “Antarctic” albedo equal to 0.27. Black and grey lines are for an aquaplanet with “Antarctic” albedo equal to 0.8 with an ozone perturbation and diabatic perturbation respectively. A thick line indicates regions in which a null hypothesis of no effect can be rejected at the 95% confidence level. The legend also includes the SAM e-folding timescale of each configuration in January. (e) only shows the runs discussed in Section 6 for clarity.
Fig. 8. Evolution of subpolar U for the (a) [ozone hole-PI] runs with (blue) realistic stationary waves, (black) an aquaplanet with “Antarctic” albedo equal to 0.8, and (gold) Northern Hemisphere with realistic stationary waves. (b) runs analogous to [ozone hole-PI] but in which a diabatic heating perturbation is imposed directly (see methods). The mean of each fifteen day segment after branching is indicated with a dot, and is labeled by the last day included in the fifteen day segment (e.g. 30 is for days 16 to 30). For (b), for the runs with a factor of five increase in diabatic heating rate, we divide the response by a factor of five. A dashed gray line indicates a constant reference slope of 3.5.

February because the already accelerated vortex allows for more gravity wave absorption in the mid-stratosphere. The advection term also contributes in regions with strong wind gradients (bottom, Figure 9). The net effect is that the dominant term for the subpolar zonal acceleration is the resolved eddy term in Figure 9ghi, and importantly this acceleration extends from the stratosphere to the surface.

Figure 10 decomposes the eddy forcing into its various wavenumber components. At early lags, the tropospheric response arises mostly through wave-2 and wave-3 (Figure 10def), while for days 71 to 120 synoptic wavenumbers are most important (Figure 10ghi). The wave-2 and wave-3 present in AQUA80 are transient planetary waves, and it is clear that they help set up the initial jet shift and then contribute a continued acceleration at subpolar latitudes. Wave-1 does not contribute to forcing the jet shift (Figure 10abc). These conclusions are true of the STAT runs as well (Figure 11) even though observed and STAT SH stationary waves are dominated by wave-1 even in the troposphere (Garfinkel et al. 2020a).
Fig. 9. Eulerian mean momentum budget for the [ozone hole-PI] aquaplanet runs, with “Antarctic” albedo equal to 0.8 in (left) days 1-30 after branching, i.e. October; (middle) days 31 to 70; (right) days 71 to 120. (a-c) total wind tendency; (d-f) sum of all terms; (g-i) eddy forcing terms \((u'v'\text{ and } u'w')\); (j-l) coriolis torque; (m-o) sum of eddy forcing and coriolis torque; (p-r) gravity wave drag; (s-u) advection of mean zonal wind. Note that the color-bar for (g-i) and (j-l) differ from that in (m-o) due to the strong cancellation between eddy forcing and coriolis torque (as expected).
Fig. 10. Decomposition of the eddy forcing term in 9ghi into the various wavenumber components. (a-c) wavenumber 1; (d-f) wavenumber 2 through 3; (g-i) wavenumbers 4 and larger.

Fig. 11. As in 10 but for STAT.
The importance of both planetary and synoptic waves is also evident using the Transformed Eulerian mean budget (as in Orr et al. 2012b). The time evolution of the upper tropospheric (200-400hPa) meridional component of the Eliassen Palm flux ($EP$) is shown in Figure 3de and 6de for STAT and AQUA80; both synoptic and planetary waves are important. The timing of the increase in $EP$ is similar for both synoptic and planetary waves, however, and thus it is unclear if one can be argued to help induce the other. That being said, these figures (and also Figure 10) show that at later lags, synoptic wavenumbers dominate the response. A similar relative role for planetary waves vs. synoptic waves for the tropospheric jet shift is evident for both AQUA80 and STAT (in both Figure 3de and 6de), and hence the presence of stationary waves does not appear to affect the ability of planetary waves to contribute to the jet shift. However the jet shift is weaker for STAT (due to a weaker stratospheric response as discussed above) and consistent with this the overall eddy forcing is weaker too (Figure 3de vs. 6de).

b. Response of STAT and AQUA80 to stratospheric diabatic heating

In addition to the ozone hole runs presented thus far, we have also performed integrations in which a diabatic heating perturbation replaces the ozone perturbation. As discussed in Section 2, the spatial structure of the diabatic heating perturbation follows the ozone perturbation, and its magnitude (-0.5K/day) mimics that due to ozone (Figure 1d-f). The benefit from these diabatic heating runs are two-fold: first, we can increase the amplitude of this diabatic heating perturbation at will and hence explore linearity of the response. (In contrast, the impact of ozone saturates as concentrations cannot be negative.) Second, there is no shortwave heating perturbation by construction (the effects of UV on the surface energy budget discussed in Chiodo et al. 2017, are turned off), and hence the stationary waves present in STAT but absent in AQUA80 are the only factor that can lead to a difference in the surface response.

We begin with the linearity of the response. Figure 8b is similar to Figure 8a, but showing the response to a diabatic heating perturbation imposed on STAT and AQUA80. By construction, the lower stratospheric and tropospheric wind response for a -0.5K/day perturbation (the dark purple and dark gray lines) in Figure 8b resemble qualitatively their
counterpart in Figure 8a, though the magnitude of the effect in both the stratosphere and
troposphere is slightly weaker (Figure 7); this slight weakening is due to the fact that the
imposed -0.5K/day perturbation is compensated by reduced longwave emission. The exper-
iments with a factor of five times stronger perturbation (-2.5K/day) are shown in Figure 8b
but with the subsequent response divided by a factor of five. It is clear that the response
is generally linear. (The response in AQUA80 is slightly weaker than might be expected by
linearity, though the response for STAT is stronger). This result highlights the fact that
interannual variability in ozone concentrations should be useful for seasonal predictability of
surface climate (Son et al. 2013; Bandoro et al. 2014; Hendon et al. 2020; Jucker and Goyal
2021).

Next, we use these diabatic forcing experiments to isolate the role of stationary waves
for the downward response, as these experiments do not allow for any effect of shortwave
radiation on the surface. The subpolar zonal wind response for STAT and AQUA80 to an
identical perturbation is shown in Figure 12a and 12b, and the difference between the two is
in Figure 12c. Initially, the diabatic perturbation causes a larger response in STAT, but the
zonal wind response in AQUA80 in both the stratosphere and troposphere becomes larger
after day 45. Hence, stationary waves lead to a negative feedback on the response even if
surface shortwave effects are suppressed, as shown in Figure 7d. A comparison of Figure 12c
and Figure 4c suggests that the response in austral summer is nearly identical, though the
apparently different response in October before the stationary wave feedback kicks in should
be considered for future work. Overall, these results support the conclusion of Chiodo et al.
(2017) that shortwave surface effects are not important for the tropospheric response.

6. Role of surface warming

Surface temperature over Antarctica cools in response to ozone depletion (Grise et al. 2009;
Yang et al. 2014; Previdi and Polvani 2014), and while much of this change is likely due to
the SAM, this cooling can still feedback onto the jet shift. We now use the idealized model
to isolate the impact of the surface temperature change on the jet shift independent of any
change in the SAM.
Fig. 12. Evolution of zonal wind from 54S to 80S for the Diabatic-PI runs with (a) realistic stationary waves, (b) an aquaplanet, with “Antarctic” albedo equal to 0.8. (c) difference between (a) and (b). The 1m/s contour is indicated in red in (a) and (b).

Recall that the albedo in both AQUA80 and STAT is 0.8 over Antarctica and 0.23 elsewhere, and in order to disentangle the role of the surface temperature changes over Antarctica on the jet shift, we have performed an additional aquaplanet integration with an albedo of 0.27 everywhere (AQUA27). AQUA80 and AQUA27 differ only in the specification of albedo, and by summer, surface temperatures rise over Antarctica by 1K due to enhanced ultraviolet absorption in AQUA27, rather than cooling by 4K as in AQUA80 (Figure 7c). The warmer near-surface tropospheric polar cap in AQUA27 leads to higher heights throughout the column, as can be quantified using the the hypsometric equation (not shown). The net effect is that the meridional gradient in geopotential is more extreme in AQUA80 than in AQUA27, and thus the stratospheric zonal wind response (Figure 7b, red vs. black; Figure 13) and tropospheric jet shift (Figure 7c; Figure 13) are therefore stronger in AQUA80. In other words, the polar surface cooling in AQUA80 reinforces the ozone-induced poleward shift, and hence is a positive feedback.

This weakening of the response in AQUA27 relative to that in AQUA80 occurs even as the jet latitude is further equatorward and the annular mode timescale of the SAM is slightly larger in AQUA27 (Figure 7, Table 1), two factors which would be expected
Fig. 13. Evolution of zonal wind from 54S to 80S for the [ozone hole-PI] runs for an aquaplanet (a) with “Antarctic” albedo equal to 0.27 (AQUA27), (b) with “Antarctic” albedo equal to 0.8 (AQUA80). (c) difference between (a) and (b). The 1m/s contour is indicated in red in (a) and (b).

to lead to a stronger response (Garfinkel et al. 2013b). This indicates that the surface temperature effect in AQUA27 overwhelms the jet latitude/eddy feedback strength effect. In order to cleanly assess the eddy feedback strength effect highlighted by Garfinkel et al. (2013b), we have performed an experiment using the AQUA80 configuration but in which the jet is pushed 7° further poleward. This is achieved by imposing a stronger and more poleward meridional ocean heat transport gradient following equation A8 of Garfinkel et al. (2020a) with an amplitude of 50 W m⁻², which leads to a poleward shift of the sea surface temperature gradient. The response to ozone depletion is shown in supplemental Figure 1, and it is clear that the tropospheric response is weaker as expected. The stationary waves and surface shortwave effects are identical in this simulation to those in AQUA80, and hence the weakened tropospheric response is due to jet latitude and weakened eddy feedback. Note that this run includes a stronger sea surface temperature front than AQUA80 yet has a weaker response, suggesting that the results of Ogawa et al. (2015) may have more to do with the eddy feedback strength in their simulations than the well-defined sea surface temperature front per se.
7. Discussion and Conclusions

Ozone depletion is known to have been the dominant contributor to a poleward shift of the Southern Hemisphere (SH) tropospheric midlatitude jet, precipitation, and storm tracks over the late 20th century. Over the next 50 years, ozone recovery is expected to nearly cancel out changes in the jet and Hadley Cell that would otherwise be forced by greenhouse gases (Polvani et al. 2011; Arblaster et al. 2011; Barnes and Polvani 2013; Gerber and Son 2014; Waugh et al. 2015; Sivir et al. 2017; Son et al. 2018; Banerjee et al. 2020). The degree of cancellation is uncertain and model dependent, however, leading to uncertainty in future projections (Gerber and Son 2014). The mechanism whereby ozone depletion leads to a downward impact, and the details of how this mechanism governs the magnitude of the impact, are still unclear (as noted in WMO Ozone assessments in 2010, 2014, and 2018). While previous work has shown that jet latitude (Garfinkel et al. 2013b) and the details of the ozone forcing (Neely et al. 2014; Young et al. 2014) are important, we have demonstrated two additional processes that are important for the downward impact: surface cooling and stationary waves.

This study takes advantage of a recently developed intermediate complexity model that can delineate the role of these two effects. We integrate it with realistic stationary waves, comparing it to runs without any zonal asymmetry in the bottom boundary. For both configurations of the bottom boundary, we compared integrations with an ozone hole in which surface shortwave feedbacks are present, to integrations with a diabatic temperature tendency that mimics the shortwave effects of ozone depletion in the stratosphere only. By comparing these runs, we isolated the role of stationary waves for the surface response, and demonstrated that the response is twice as strong for many of the diagnostics examined when no stationary waves are present (Figure 1mno, 5, 6, and 12ab). The presence of stationary waves leads to a weaker response starting in late November and extending into February to an identical diabatic heating perturbation. This effect arises because a stationary wave negative feedback leads to a weaker stratospheric circulation response when there are stationary waves in the troposphere. Further, we demonstrated that radiative effects are not important due to the high albedo over Antarctica (in agreement with Chiado et al. 2017). Finally, by integrating the model in an aquaplanet configuration but with different surface albedos over
“Antarctica”, we isolated the role of surface temperature and showed that surface cooling enhances the jet response.

Despite the negative stationary wave feedback on the magnitude of the stratospheric circulation response to ozone depletion, tropospheric planetary waves are important for the tropospheric jet response. Both planetary and synoptic waves are important for the tropospheric jet shift both in AQUA80 and STAT integrations (Figures 10 and 11). Waves 1-3 contribute roughly half of the torque in November, though by December and January their contribution is less (Figure 3de and 6de). This is true for both of the ozone depletion runs, and also the diabatic heating runs where we can increase the amplitude of the forcing to better capture the response (Figure 14).

Gravity waves also act as a negative feedback on the magnitude of the stratospheric circulation response to ozone depletion. Namely, the strengthened polar vortex allows more gravity waves to propagate into the stratosphere, and these gravity waves then break in the subpolar mid-stratosphere (Figure 9). This partial compensation between gravity waves and an externally imposed forcing is consistent with Cohen et al. (2013); Sigmond and Shepherd (2014); Scheffler and Pulido (2015); Watson and Gray (2015), and Garfinkel and Oman (2018).

The specific mechanism as to how the downward influence arises is not the main focus of this paper, however our results are of relevance to previously proposed theories. Wave-2 and wave-3 are crucial in the lower stratospheric zonal momentum response (Figures 10 and 11, consistent with Orr et al. 2012b). Both planetary and synoptic waves are important for the tropospheric impact, and it is impossible to distinguish whether one begins before the other. This difficulty occurs even if we enhance the signal-to-noise ratio by imposing a diabatic heating perturbation five times stronger than that associated with ozone depletion (Figure 14de), likely because of eddy-eddy interactions (Domeisen et al. 2013; Smith and Scott 2016). Synoptic waves are somewhat more important in summer, but in late fall the momentum forcing is more evenly split between synoptic and planetary waves. This balance is evident both in AQUA80 and in STAT even as stationary wave-1 is present only in STAT. The tropospheric response begins first at subpolar latitudes and only later, after synoptic eddies dominate, includes the midlatitudes.
In all runs, a tropospheric response does not begin until at least day 15, and in the diabatic heating runs with the forcing increased by a factor of five, there is even a weak equatorward shift in the first ten days (though not evident in Figure 14b using the chosen contour interval). This arises because a thermally driven cooling of the vortex will be balanced in part by downwelling over the pole and equatorward motion in the troposphere, which leads to an easterly Coriolis torque (Eliassen 1951). This opposite response is consistent with Yang et al. (2015) who also find that the residual circulation is of the wrong sign to explain the poleward shift, and also with White et al. (2020) who also impose a diabatic heating perturbation and find that the poleward shift does not occur for at least 15 days. This effect does not explain why the observed poleward shift is not robust until December, however, as this delay is far longer than 15 days.

On the other hand, our simulations help clarify the important factors for the onset of the response, and thereby help explain why the jet response in observations and in STAT only becomes robust in summer after the ozone hole is already filling up. Namely, the tropospheric response can begin in late October if the forcing is strong (Figure 14b) or in AQUA80 (Figure 6b). The net effect is that the delay until December in STAT is a consequence of the stationary waves and the relative weakness of the diabatic heating perturbation associated with ozone depletion. Our simulations also demonstrated that the near surface cooling over Antarctica evident in observations helps encourage the poleward jet shift. Future work should evaluate whether the stationary wave feedback or surface cooling response is crucial for the timing of the onset of the response in observations and comprehensive models as well.

The response to an identical ozone hole imposed in the Northern Hemisphere in STAT (STATNH) is significantly weaker than when imposed in the Southern Hemisphere (Supplemental Figure 2). In other words, the tropospheric circulation in the Northern Hemisphere is less sensitive to a stratospheric ozone perturbation. The negative stationary wave feedback likely plays a role. Northern Hemisphere stationary waves are stronger, and hence the stratospheric circulation response to an identical ozone depletion is weaker due to an offset by enhanced wave propagation into the stratosphere. In addition, the annular mode
timescale is shorter in the Northern Hemisphere (22 days; Figure 8), and hence synoptic
eddy feedbacks are weaker too.

In the most realistic configuration (STAT), the model simulates a response resembling
that observed and simulated by comprehensive models (Figure 1, 2, and 3). Nevertheless,
the model used in this work suffers from some limitations - there is no coupling of the
ozone with the dynamics, the imposed ozone hole has no zonal structure, and the albedo is
constant for all shortwave wavelengths. Despite these limitations, the results of our work
have implications for seasonal forecasting and for the interpretation of results from both
comprehensive and idealized models. First, interannual variability in ozone concentrations
can be used to enhance seasonal forecasting (Figure 8), consistent with Hendon et al. (2020)
and Jucker and Goyal (2021). Second, dry and flat idealized models miss the stationary
wave effect, which may lead to an exaggerated doubling of the response to an identical
ozone hole. Third, the Antarctic surface temperature response to ozone depletion helps
regulate the magnitude of the jet response, and it is not clear how well models can capture
the stable boundary layers common over Antarctica, the mixed-phase and ice clouds common
at these latitudes, or the properties of a glaciated land surface. Future work should explore
whether diversity in how models represent these processes can explain some of the diversity
in future projections of climate change in the Southern Hemisphere (Gerber and Son 2014),
and thereby help narrow projections as ozone recovers.

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APPENDIX

Data Statement: The updated version of MiMA used in this
study including the modified source code can be downloaded from
Fig. 14. As in Figure 3 but for a diabatic heating rate of 2.5K/day in the lower stratosphere and no ozone depletion. Note factor of 5 difference in colorbar for (a) and (b), and factor of 2 difference for (c)-(e).


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