The Influence of Zonally Asymmetric Stratospheric Ozone Changes on the Arctic Polar Vortex Shift

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ABSTRACT

Recent studies have found a shift of the Arctic stratospheric polar vortex toward Siberia during late winter since 1980, intensifying the zonally asymmetric ozone (ZAO) depletion in the northern middle and high latitudes with a stronger total column ozone decline over Siberia compared with that above other regions at the same latitudes. Using observations and a climate model, this study shows that zonally asymmetric stratospheric ozone depletion gives a significant feedback on the position of the polar vortex and further favors the stratospheric polar vortex shift toward Siberia in February for the period 1980–99. The polar vortex shift is not significant in the experiment forced by zonal mean ozone fields. The February ZAO trend with a stronger ozone decline over Siberia causes a lower temperature over this region than over the other regions at the same latitudes, due to shortwave radiative cooling and dynamical cooling. The combined cooling effects induce an anomalous cyclonic flow over Siberia, corresponding to the polar vortex shift toward Siberia. In addition, the ZAO depletion also increases the meridional gradient of potential vorticity over Siberia, which is favorable for the upward propagation of planetary wave fluxes from the troposphere over this region. Increased horizontal divergence of planetary waves fluxes over the region 60°–75°N, 60°–90°E associated with ZAO changes accelerates the high-latitude zonal westerlies in the middle stratosphere, further enhancing the shift of the stratospheric polar vortex toward Siberia. After 2000, the ZAO trend in February is weaker and induces a smaller polar vortex shift than that in the period 1980–99.

1. Introduction

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Stratospheric ozone depletion not only leads to a potential increase in ultraviolet radiation at the surface that is harmful for human health and to terrestrial and aquatic ecosystems, but also has important impacts on atmospheric temperature and circulation through ozone

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radiative heating (e.g., [Ramaswamy et al. 1996](#); [Forster and Shine 1997](#); [Shindell et al. 1999](#); [Tian and Chipperfield 2005](#); [Xie et al. 2008](#); [Wang et al. 2014](#); [Nowack et al. 2015](#); [Garfinkel 2017](#)). Although previous studies have shown that the ozone change in the Antarctic stratosphere can exert a significant influence on the tropospheric climate in the Southern Hemisphere (e.g., [Sexton 2001](#); [Shindell and Schmidt 2004](#); [Son et al. 2008](#); [Polvani et al. 2011](#); [Thompson et al. 2011](#); [Keeble et al. 2014](#)), there is not similarly robust evidence of the response of the Northern Hemispheric tropospheric climate to stratospheric ozone depletion due to large interannual ozone variability and comparatively weaker ozone depletion (e.g., [Thompson and Solomon 2005](#)). Using numerical model simulations, most previous studies only found significant regional surface responses over the northern middle and high latitudes when comparing cases of extremely high and low Arctic stratospheric ozone ([Cheung et al. 2014](#); [Karpechko et al. 2014](#); [Smith and Polvani 2014](#); [Calvo et al. 2015](#); [Xie et al. 2016, 2017](#); [Harari et al. 2019](#)). One possible reason is that the dynamical variability of the Arctic stratospheric polar vortex is much larger than its Antarctic counterpart ([Hu and Guan 2018](#); [Hu et al. 2019a](#); [Mai et al. 2020](#)) and as a result Arctic stratospheric ozone depletion is less persistent than Antarctic ozone depletion ([Manney et al. 2011](#)). Therefore, one would expect a weaker impact of ozone depletion on surface conditions in the Northern Hemisphere compared to the Southern Hemisphere. In fact, the stratospheric polar vortex condition may play a critical role in the potential impacts of stratospheric ozone depletion on the tropospheric climate in the Northern Hemisphere.

However, the impact of stratospheric ozone changes on the Arctic stratospheric polar vortex remains unclear. Stratospheric ozone depletion since 1980, especially in the 1990s, may have led to a strengthening of the Arctic stratospheric polar vortex in spring through absorbing less shortwave radiation and thereby reducing the Arctic stratospheric temperature ([Bohlinger et al. 2014](#); [Ivy et al. 2016](#); [Xie et al. 2016](#)). The influence of stratospheric ozone changes on the wintertime polar vortex is more complicated, due to the competing effects between radiative and dynamical processes in the polar region during winter ([Bohlinger et al. 2014](#); [Hu et al. 2015](#); [Ivy et al. 2016](#)). Both [Bohlinger et al. \(2014\)](#) and [Ivy et al. \(2016\)](#) found that a warming in the Arctic stratosphere due to increased planetary wave activity in winter is partially cancelled by a radiative cooling associated with stratospheric ozone depletion, leading to a nonsignificant warming during late winter since 1980 in the Arctic lower stratosphere. [Manzini et al. \(2003\)](#) pointed out that the negative dynamical feedback between ozone depletion and increases in wave forcing

could partly explain the slight warming of the Arctic stratosphere. This dynamical response to ozone depletion has also been discussed in other studies (e.g., [McLandress et al. 2010](#); [Lin et al. 2017](#); [Haase and Matthes 2019](#)). [Manzini et al. \(2003\)](#) proposed that this negative dynamical feedback is caused by the differential filtering effect of the stratospheric polar vortex on gravity waves, while [Albers and Nathan \(2013\)](#) attributed it to the amplification of planetary waves in the lower stratosphere. In addition, [Hu et al. \(2015\)](#) pointed out that the combined effects of reduced upward stationary wave flux and enhanced upward transient wave flux due to stratospheric ozone decreases may cause a rather weak cooling in the middle and upper Arctic stratosphere.

In addition to the zonal mean ozone depletion observed in the past several decades, stratospheric ozone also exhibits a zonally asymmetric variation, which can have a different impact on climate variability compared with that induced by zonally symmetric ozone variation (e.g., [Sassi et al. 2005](#); [Gabriel et al. 2007](#); [Crook et al. 2008](#); [Vaughan et al. 2009](#); [Gillett et al. 2009](#); [McCormack et al. 2011](#); [Albers and Nathan 2012](#); [Peters et al. 2015](#); [Rae et al. 2019](#)). Previous studies have showed that numerical simulations with zonally asymmetric ozone (ZAO) variations in the model can produce a warmer and more dynamically active polar stratosphere in the Northern Hemisphere than that without ZAO variations ([Gabriel et al. 2007](#); [Crook et al. 2008](#); [Gillett et al. 2009](#); [McCormack et al. 2011](#)). [Gillett et al. \(2009\)](#) found that ZAO variations could lead to up to a 3-K increase of the zonal mean temperature in the Arctic stratosphere. [McCormack et al. \(2011\)](#) pointed out that simulations including ZAO variations in the model produce more frequent major stratospheric sudden warmings compared to simulations forced by zonal mean ozone fields, indicating that ZAO variations can lead to a weaker stratospheric polar vortex. The weakening of the polar vortex is related to increases in the amplitude of zonal wave 1 in the stratosphere due to ZAO variations ([Gabriel et al. 2007](#); [McCormack et al. 2011](#); [Albers and Nathan 2012](#)).

Although the above-mentioned studies revealed that stratospheric ozone variations and its zonally asymmetric component can alter polar vortex strength, little is known regarding the effects of stratospheric ozone changes on stratospheric polar vortex position. [Zhang et al. \(2016\)](#) found that the Arctic stratospheric polar vortex has been shifted toward Siberia during late winter over the period 1980–2015. They attributed the shift of the stratospheric polar vortex to enhanced upward propagation of baroclinic waves associated with Arctic sea ice loss and increased Siberian snow cover. They hypothesized that other factors, such as

stratospheric ozone depletion, may also affect the vortex position. One purpose of this study is to verify this hypothesis.

Zhang et al. (2018) pointed out that the stratospheric polar vortex shift could transport air, which is both ozone-poor and high in active chlorine, from the Arctic regions to the Eurasian continent, leading to a ZAO structure with a stronger total column ozone (TCO) decline over Eurasia than over other regions at the same latitudes. This conclusion suggests that the polar vortex shift could further enhance the ZAO structure in the middle and high latitudes. Their studies used only observations and a chemical transport model forced by fixed meteorological fields to analyze the impact of the polar vortex shift on ozone; however, they did not discuss whether the ZAO variations associated with the polar vortex shift could give a further feedback on the polar vortex position. Until now there has been no evidence to answer this question. Better understanding of the potential impacts of stratospheric ozone changes on the polar vortex position would be helpful for accurately predicting the Arctic vortex shift and its associated extremely cold events in the middle and high latitudes. Here we attempt to address these scientific issues. Section 2 describes the data, methods, and climate model experiment designs used in this study. Section 3 presents the ZAO impact on the polar vortex position, section 4 attempts to analyze its associated mechanisms, and the conclusions and discussion are shown in section 5.

2. Data, methods, and model

a. Data

In this study, we focus on a detailed analysis on the period 1980–99 when significant stratospheric ozone depletion occurs, but we also consider the changes during 2000–09. We use monthly mean data from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) to calculate the polar vortex edge and center. The data have a horizontal resolution of 1° latitude $\times$ 1° longitude and are subsequently interpolated vertically from pressure coordinates (1000–1 hPa) onto isentropic coordinates (400–1400 K). Monthly mean geopotential height, potential vorticity, and temperature fields are also derived from the ERA-Interim dataset.

To investigate the impact of variations in stratospheric ozone on the polar vortex, we use monthly mean ozone fields derived from the National Aeronautics and Space Administration Modern Era-Retrospective Analysis for Research and Applications, (NASA-MERRA) (Rienecker

et al. 2011). The TCO data derived from MERRA reanalysis data are also employed. MERRA ozone data have been extensively evaluated against various satellite datasets and were found to well represent northern middle- and high-latitude ozone concentrations (Xie et al. 2016).

b. Methods

The polar vortex edge is defined as the location of the largest potential vorticity (PV) gradient on isentropic surfaces, with an additional constraint of close proximity to a strong westerly jet, according to the method of Nash et al. (1996). The center of the polar vortex is calculated by the PV-weighted average method described by Eqs. (1)–(3) (Zhang et al. 2019a):

$$PV = -g(\zeta + f) \frac{\partial \theta}{\partial p}, \quad (1)$$

$$\text{latc} = \frac{\sum_{i=1}^n \text{lat}(i) \times PV(i) \times \cos[\text{lat}(i)]}{\sum_{i=1}^n PV(i) \times \cos[\text{lat}(i)]}, \quad \text{and} \quad (2)$$

$$\text{lonc} = \frac{\sum_{i=1}^n \text{lon}(i) \times PV(i) \times \cos[\text{lat}(i)]}{\sum_{i=1}^n PV(i) \times \cos[\text{lat}(i)]}, \quad (3)$$

where θ is the potential temperature, ζ is the relative vorticity, p is the pressure coordinate, g is the gravitational acceleration that is equal to 9.8 m s^{-2} , and f is the Coriolis parameter. Also, latc and lonc are the latitude and longitude of the polar vortex center, respectively, and n is the number of grid points in the polar vortex circled by the polar vortex edge; $PV(i)$, $\text{lat}(i)$, and $\text{lon}(i)$ are the PV, latitude, and longitude of the i th grid, respectively. Before calculating the longitude of the polar vortex center, the longitude coordinate is set up within the range from 180°W to 180°E .

PV increases exponentially with altitude in the stratosphere, indicating that averaging PV over a vertical layer without scaling the PV will automatically and significantly weight the average toward the higher levels. Here, scaled PV is used according to the definition by Dunkerton and Delisi (1986) as follows:

$$PV_s = \frac{PV}{g \left| \frac{\partial \theta}{\partial p} \right|}, \quad (4)$$

where $\partial \theta / \partial p$ is the standard atmosphere value of $\partial \theta / \partial p$ at the standard height of the isentropic surface. Figure S1 in the online supplemental material shows

that there are similar patterns in scaled PV anomalies and polar vortex shift over different isentropic surfaces (e.g., 600, 800, 900, and 1000 K), suggesting that it is reasonable to average the vertical layers between 600 and 1000 K for analysis.

The Plumb wave activity fluxes (Plumb 1985) are used to diagnose the propagation of planetary waves in the three-dimensional (3D) space. The Plumb flux components are calculated as follows:

$$F_{p\lambda} = \frac{p}{2a^2 \cos\phi} \left[\left(\frac{\partial\psi'}{\partial\lambda} \right)^2 - \psi' \frac{\partial^2\psi'}{\partial\lambda^2} \right], \quad (5)$$

$$F_{p\phi} = \frac{p}{2a^2} \left(\frac{\partial\psi'}{\partial\lambda} \frac{\partial\psi'}{\partial\phi} - \psi' \frac{\partial^2\psi'}{\partial\lambda\partial\phi} \right), \quad \text{and} \quad (6)$$

$$F_{pz} = \frac{2\Omega^2 p \sin^2\phi}{N^2 a} \left(\frac{\partial\psi'}{\partial\lambda} \frac{\partial\psi'}{\partial z} - \psi' \frac{\partial^2\psi'}{\partial\lambda\partial z} \right). \quad (7)$$

Its horizontal divergence is written as follows:

$$\nabla \mathbf{F}_{xy} = \frac{1}{a \cos\phi} \frac{\partial F_{p\lambda}}{\partial\lambda} + \frac{1}{a \cos\phi} \frac{\partial F_{p\phi}}{\partial\phi} \cos\phi, \quad (8)$$

where p is the pressure, λ is the longitude, and ψ is the streamfunction. The prime represents a perturbation quantity. The streamfunction ψ is calculated using u and v by integrating the equations

$$u = -\frac{\partial\psi}{\partial y} \quad \text{and} \quad (9)$$

$$v = \frac{\partial\psi}{\partial x}. \quad (10)$$

The stationary wave refractive index n_k^2 (Andrews et al. 1987) is able to characterize the atmospheric environment for planetary wave propagation and is defined as follows:

$$n_k^2 = \frac{\bar{q}_\phi}{\bar{u}} - \left(\frac{k}{a \cos\phi} \right)^2 - \left(\frac{f}{2NH} \right)^2, \quad (11)$$

where

$$\bar{q}_\phi = \frac{2\Omega}{a} \cos\phi - \frac{1}{a^2} \left[\frac{(\bar{u} \cos\phi)\phi}{\cos\phi} \right]_\phi - \frac{f^2}{\rho_0} \left(\frac{\bar{u}_z}{\rho_0 N^2} \right)_z. \quad (12)$$

Expansion of the third term on the right-hand side of Eq. (12) yields

$$-\frac{f^2}{\rho_0} \left(\frac{\bar{u}_z}{\rho_0 N^2} \right)_z = \left(\frac{f^2}{HN^2} + \frac{f^2}{N^4} \frac{dN^2}{dz} \right) \bar{u}_z - \frac{f^2}{N^2} \bar{u}_{zz}. \quad (13)$$

Here, $\bar{q}_\phi$, H , N^2 , Ω , and k denote the meridional gradient of the zonal-mean PV gradient, scale height, buoyancy frequency, Earth rotation frequency, and zonal wave-number, respectively. Planetary waves can propagate in regions with positive n_k^2 , whereas they avoid areas with negative n_k^2 . However, due to the overlapping of positive and negative refractive indices, the composite refractive index squared is difficult to interpret. Thus, in this study, we use the meridional PV gradient $\bar{q}_\phi$ that is proportional to the refractive index to analyze whether ZAO variations can affect the wave propagation conditions.

c. Numerical model and experiments

In addition, to probe the possible mechanisms by which stratospheric ozone depletion may affect the polar vortex, five time-slice simulations are performed using the Whole Atmosphere Community Climate Model, version 3 (WACCM3). WACCM3 has 66 vertical levels extending from the ground to ~ 145 -km geometric altitude. WACCM3 has been extensively evaluated against various satellite datasets and has a good performance in simulating stratospheric chemistry and dynamics (Eyring et al. 2006). Details of the model can be found in Garcia et al. (2007). All time-slice simulations presented in this study were performed at a horizontal resolution of 1.9° latitude $\times$ 2.5° longitude, with interactive chemistry and QBO forcing disabled. Three runs (O3R1–3D, O3R2–3D, and O3R3–3D) were conducted using the same greenhouse gas emissions and mixing ratio lower boundary conditions for chlorofluorocarbons as well as the same sea surface temperature forcing, except that the ozone forcings in O3R1–3D, O3R2–3D, and O3R3–3D are the decadal mean of the monthly 3D ozone field derived from MERRA reanalysis data over the periods 1980–89, 1990–99, and 2000–09, respectively. All the experiments use the fixed surface mixing ratios for GHGs in the year 1980 derived from the scenario A1B of the IPCC (Stocker et al. 2001). The mixing ratios for CFC11 and CFC12 in the year 1980 are taken from Table 4B-2 of WMO (2003). Sea surface temperature and sea ice concentrations used in all the experiments are the observed climatology annual cycle for the time period from 1980 to 2009 derived from the Met Office Hadley Center (Rayner et al. 2003). In addition, to analyze the impacts of zonally symmetric ozone changes on the polar vortex, we performed another two runs (O3R1–2D and O3R2–2D) that are forced by zonal mean ozone fields derived from MERRA reanalysis data over the periods 1980–89 and 1990–99, respectively. All the simulations were run for 45 years with the first 5 years excluded for the model spinup and the remaining 40 years of data used for analysis.

FIG. 1. Linear trends in TCO in the Northern Hemisphere in (a) January, (b) February, and (c) March for the period 1980–99 derived from MERRA reanalysis data. Dotted regions indicate statistical significance at the 90% confidence level according to a two-tailed Student's t test.

3. Impacts of stratospheric ozone changes on the polar vortex

Figure 1 presents the linear trends in MERRA TCO for January, February, and March, respectively, over the period 1980–99. Note that there are positive TCO trends without statistical significance in January, while the Arctic TCO depletion is significant in March. Compared to the TCO trends in January and March, the trends in TCO in February show a more asymmetric structure in the high latitudes with a strong TCO decline over Siberia. This feature is consistent with the conclusion of Zhang et al. (2016, 2018, 2019b) that the shift of the Arctic stratospheric polar vortex toward Eurasia is more significant in February. Considering that the most significant ZAO trend occurs in February, we will now focus on the impacts of ZAO changes on the polar vortex in that month. A composite analysis is performed with respect to this anomalous ZAO structure in February (Fig. 1b). Strong ZAO events are selected based on the criteria that the normalized time series of the difference between the zonal TCO anomaly (defined as the departure of TCO at a given location from its corresponding zonal mean) in February within the sector 0° – 90° E, 60° – 75° N and those within 90° W– 180° , 60° – 75° N should be less than -0.5 , while the normalized difference for weak ZAO events is larger than 0.5 .

Figures 2a and 2b show the climatological mean zonal ozone anomaly (contour lines) and composites for zonal ozone anomaly (color shaded) during strong and weak ZAO events according to our definition. For either strong or weak ZAO events, the zonal ozone anomaly is in phase with that for the climatological mean. However,

positive and negative anomaly centers are stronger during strong ZAO events than during weak ZAO events, underlining that the selected strong and weak ZAO events correspond to stronger and weaker zonally asymmetric ozone variations, respectively. Figures 2c and 2d show cross sections along 60° – 75° N of differences in zonal ozone anomaly between strong and weak ZAO events, and between 1990s and 1980s ozone conditions, respectively. Note that during strong ZAO events (in the 1990s), the negative center of ozone difference within the sector 0° – 90° E is stronger than that during weak ZAO events (in the 1980s), whereas the positive center within the sector 90° W– 180° shows higher value during strong ZAO events (in the 1990s). This result suggests that the ZAO structure in the 1990s is stronger than that in the 1980s. We found 5 out of 7 strong ZAO events are from the period 1990–99, while 5 out of 7 weak ZAO events are from the period 1980–89.

Figure 3a shows the composite of scaled PV differences in February within the layer 600–1000 K between strong and weak ZAO events derived from the ERA-Interim data for the period 1980–99. Note that there are significant positive PV anomalies over Siberia and Europe and negative PV anomalies over North America. Correspondingly, the composited mean location of the polar vortex edge during the strong ZAO events is closer to Siberia than that during the weak ZAO events. This result suggests that the polar vortex shift toward Siberia is closely related to strong ZAO. Figure 3b shows that the polar vortex in the 1990s is shifted toward Siberia compared with that in the 1980s though the PV anomalies and polar vortex shift are weaker than those in the composite analysis based on strong/weak ZAO events, which may be due to cancelling effects of

FIG. 2. Isentropic surface–longitude cross section between 60° and 75°N of composites for zonal ozone anomaly (defined as the departure of ozone at a given location from its corresponding zonal mean; color shaded) in the (a) strong and (b) weak ZAO events, and (c) their differences (strong ZAO minus weak ZAO). (d) Isentropic surface–longitude cross section between 60° and 75°N of ozone differences between the 1990s and the 1980s. The climatological mean zonal ozone anomaly for the period 1980–99 is also shown (contour lines; interval: 0.1 ppmv; solid and dashed lines represent positive and negative values, respectively). All data are derived from MERRA reanalysis data. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's t test.

internal variabilities of the climate system. However, this composite analysis only suggests a coincidence between ZAO events and a polar vortex shift, and the feedback of ZAO on the polar vortex position is still not verified. Here, the model simulations O3R2–3D and O3R1–3D, which are forced by the MERRA 3D ozone field in the 1990s and 1980s (see section 2 for more details), respectively, could provide further evidence. Figure 3c shows the scaled PV differences in February within the layer 600–1000 K between O3R2–3D and O3R1–3D. There are also positive PV differences over Siberia and negative PV differences over North America, although the magnitude of modeled PV responses is relatively smaller than that composited from the reanalysis data. The polar vortex edge and center in the O3R2–3D run are also closer to Siberia than those in the O3R1–3D run, which is consistent with the composite results (Figs. 3a,b).

Table 1 shows changes in the latitudes and longitudes of the polar vortex center associated with stratospheric ozone changes using reanalysis data composite and WACCM3 simulations. Note that the latitude of the polar vortex center during strong ZAO events is more southward than that during the weak ZAO events, corresponding to the shift of the polar vortex to lower latitudes. In addition, the longitude of the polar vortex center during the strong ZAO events is more eastward than that during the weak ZAO events, corresponding to a displacement of polar vortex center toward Siberia. In the 1990s composite, the latitude and longitude of the polar vortex center do not show a significant shift compared with those in the 1980s composite, which may be masked by various climate forcings. Thus we used WACCM3 simulations with fixed SST and greenhouse gases (i.e., O3R1–3D and O3R2–3D) to

FIG. 3. (a) The average of scaled PV differences within the layer 600–1000 K (color shaded) in February between the strong and weak ZAO events. The color contour lines and color crosses represent the mean polar vortex edges and centers, respectively, in strong (blue) and weak (red) ZAO events. (b) As in (a), but for the scaled PV differences between the 1990s and 1980s and vortex edges and centers in the 1990s (blue) and 1980s (red). (c) As in (a), but for the scaled PV differences between the runs O3R2–3D and O3R1–3D (see section 2 for more details) and vortex edges and centers in O3R2–3D (blue) and O3R1–3D (red). White and green dotted regions indicate statistical significance at the 90% and 95% confidence level according to a two-tailed Student's t test, respectively.

isolate the impacts of ZAO on the polar vortex center. The polar vortex center in the O3R2–3D run shows a more eastward and southward displacement compared with that in the O3R1–3D run, in agreement with the comparison between strong and weak ZAO events. The above-mentioned analysis results indicate that there may exist a feedback effect of ZAO anomalies induced by the polar vortex shift on the position of polar vortex in February (Fig. 3).

To cross-check whether the polar vortex is shifted on daily time scales, we calculate the probability density function (PDF) of the daily-mean fractional vortex area over Siberia and North America. The fractional area of the polar vortex is defined as the percentage area of the polar vortex that covers one region (e.g., Siberia or North America) divided by the total area of the polar vortex. Figures 4a and 4c show the PDFs of the daily mean fractional area of the polar vortex over Siberia during strong and weak ZAO events as well as in the 1990s and 1980s. During strong ZAO events (1990s), the probability of the polar vortex covering a larger fractional area of Siberia is greater than during weak ZAO events (1980s). This feature is consistent with the polar vortex shift toward Siberia shown in the composite results of ERA-Interim data (Figs. 3a,b). However, there is no significant difference in the PDF of fractional area over North America between strong ZAO events (1990s) and weak ZAO events (1980s), which is contradictory to the polar vortex shift away from North America (Figs. 3a,b). This may be related to limited sampling for the composite analysis and large

interannual variability among different samples. The Kolmogorov-Smirnov test results show that the difference in PDF of polar vortex area over Siberia between strong and weak ZAO events is statistically significant at 0.01 level, while that over North America is not. We further analyzed the changes in PDFs of fractional polar vortex area over Siberia in the WACCM3 simulations (Fig. 4e) and found that the large fractional area over Siberia is more likely to occur in O3R2–3D than in O3R1–3D, consistent with the results from the

TABLE 1. The latitudes and longitudes of the polar vortex center during February derived from ERA-Interim reanalysis data and different WACCM3 experiments. The positive and negative values in the longitude column denote east and west longitude, respectively.

	Latitude (°)	Longitude (°)
Anomalous ZAO cases from ERA-Interim data		
Weak ZAO cases	82.4	−25.5
Strong ZAO cases	78.7	14.5
Decadal mean from ERA-Interim data		
1980s	80.0	−0.2
1990s	79.8	15.0
ZAO forcing WACCM experiments		
O3R1–3D (1980s)	85.4	15.6
O3R2–3D (1990s)	83.6	44.3
O3R3–3D (2000s)	83.9	33.3
Zonally symmetric ozone forcing WACCM experiments		
O3R1–2D (1980s)	84.7	32.0
O3R2–2D (1990s)	84.9	35.0

FIG. 4. Probability density function (PDF) of February fractional area of (a) Siberian (45° – 85° N, 45° – 135° E) and (b) North American (45° – 85° N, 45° – 135° W) regions covered by the stratospheric polar vortex during strong (blue) and weak (red) ZAO events. (c),(d) As in (a) and (b), respectively, but for the 1980s (red) and 1990s (blue) composite. (e),(f) As in (a) and (b), respectively, but for runs O3R1–3D (red) and O3R2–3D (blue). The fractional area of polar vortex is defined as the percentage area of the polar vortex that covers one region divided by the total area of the polar vortex.

composite analysis. In contrast, there is a significant lower occurrence in the simulations of the polar vortex being located over North America in O3R2–3D than in O3R1–3D (Fig. 4f). The Kolmogorov-Smirnov test results show that all the differences in histograms between O3R2–3D than in O3R1–3D are statistically significant at the 0.01 level.

A question arises to whether the zonally symmetric ozone changes can also modulate the polar vortex position. We performed two additional experiments forced by the decadal mean of monthly and zonal mean ozone fields derived from MERRA reanalysis data during the 1980s and 1990s (O3R1–2D and O3R2–2D; see section 2 for more details). Figure 5a shows the scaled

FIG. 5. (a) The average of scaled PV differences within the layer 600–1000 K (color shaded) in February between the runs O3R2–2D and O3R1–2D. The color contour lines and color crosses represent the mean polar vortex edges and centers, respectively, in O3R2–2D (blue) and O3R1–2D (red). (b) As in (a), but for the scaled PV differences between the runs O3R2–3D and O3R2–2D (see section 2 for more details) and vortex edges and centers in O3R2–3D (blue) and O3R2–2D (red). Dotted regions indicate statistical significance at the 90% confidence level according to a two-tailed Student's t test.

PV differences between the runs O3R2–2D and O3R1–2D along with the decadal mean of polar vortex edges in each period. Note that the changes in polar vortex position are totally different from those forced by MERRA 3D ozone data (Fig. 3c). There are no significant changes in the polar vortex edge and center between the 2D runs in the 1990s and 1980s, although the Siberian PV anomalies are marginally positive, which is different from the significant polar vortex shift toward Siberia during February in the 1990s forced by 3D ozone data. Table 1 further shows that the latitude and longitude of the polar vortex center in the zonally symmetric ozone run O3R2–2D are not significantly different from those in O3R1–2D. In addition, when comparing the O3R2–3D and O3R2–2D runs, we also found that the polar vortex shows more of a shift toward Siberia in the ZAO run than that in the zonally symmetric ozone run in the 1990s (Fig. 5b). The discernible differences between the 2D runs and the 3D runs suggest that the impacts of ZAO variations on the polar vortex position are not negligible.

4. Mechanisms responsible for the ZAO impacts on the polar vortex

First, we attempt to clarify the relative importance of radiative processes and dynamical processes in the change of polar vortex position associated with ZAO

variation. Figure 6 shows the differences in longwave and shortwave heating rates and temperature tendency due to dynamical processes between the WACCM3 simulations. Stratospheric ozone depletion leads to a smaller shortwave heating rate in the O3R2–3D run forced by the 1990s ozone than that in the O3R1–3D forced by the 1980s ozone (Fig. 6a), with a maximum reduction of shortwave heating rate over Siberia due to the strong ozone decline here. By contrast, the differences in the shortwave heating rate between the zonally symmetric ozone runs O3R2–2D and O3R1–2D show a nearly uniform pattern in the middle and high latitudes (Fig. 6c), in agreement with the zonal mean ozone forcing used in the two runs. In addition, there is a larger longwave heating rate over Siberia in the O3R2–3D run (Fig. 6b) compared with that in the O3R1–3D run. This is because lower stratospheric ozone concentrations in the middle and lower stratosphere allow more longwave radiation from the Earth surface to reach higher altitudes (Chipperfield and Pyle 1988; Gabriel et al. 2007). However, the positive differences of the longwave heating rate over Siberia between the zonally symmetric ozone runs O3R2–2D and O3R1–2D are statistically insignificant (Fig. 6d). Note that the longwave heating rate differences are larger than shortwave heating rate differences, which is related to relatively weak solar radiation at high latitudes during late winter. The differences in dynamical heating rate between

FIG. 6. Differences of (a) shortwave heating rate, (b) longwave heating rate, (e) temperature tendency due to dynamical processes, and (f) temperature averaged between 30 and 10 hPa during February between the runs O3R2-3D and O3R1-3D. (c),(d) As in (a) and (b), respectively, but for the differences between the runs O3R2-2D and O3R1-2D. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's t test.

the ZAO runs O3R2–3D and O3R1–3D are opposite to the net radiative heating rate differences (Fig. 6e), suggesting that adiabatic cooling associated with dynamical processes roughly cancels with the net radiative heating due to ozone (Birner and Charlesworth 2017). In summary, the dynamical cooling and the shortwave cooling overwhelms the longwave heating, which results in lower midstratospheric temperature over Siberia in the O3R2–3D run than in the O3R1–3D run (Fig. 6f).

The stratospheric ozone decline over Siberia not only leads to local temperature changes over this region, but also induces a zonally asymmetric heating in the stratosphere. The spatial pattern of the temperature anomalies in the midstratosphere shows a dipolar structure (Fig. 6f), resembling a pattern of zonal wave 1, suggesting that planetary wave 1 has an important contribution to the zonally asymmetric heating. Actually, previous studies have shown that a strengthening of zonal wave 1 has an important contribution to the upward planetary wave flux in the stratosphere and favors the shift of polar vortex toward Siberia (Zhang et al. 2016; Hu et al. 2018, 2019b; Huang et al. 2018). Therefore, it is necessary to analyze the changes in zonal wave 1 associated with ZAO. Figure 7a shows zonal wave-1 components of temperature differences between the 1990s and 1980s (color shaded) along with the climatological mean of the zonal wave 1 component (contour lines). There are negative temperature anomalies over Siberia and positive anomalies over the North American continent, in phase with the ozone differences between the 1990s and 1980s (Fig. 2d). The zonal wave-1 temperature anomalies enhance the climatological mean zonal wave 1 to some extent. Lower temperatures in the middle stratosphere over Siberia result in compressed air thickness and thereby cyclonic flows, showing that the polar vortex is shifted toward Siberia. Conversely, higher temperatures in the middle stratosphere over North America result in expansion of air thickness and anticyclonic flows.

An interesting feature is that the negative and positive centers of temperature anomalies show an eastward shift by about 50° compared with the climatological mean, consistent with the findings of Peters et al. (2015) that ZAO heating could induce an eastward shift of planetary wave 1. This phenomenon is also seen in the pressure–longitude cross section of temperature anomalies in the stratosphere (Fig. 7c) with a strengthening and an eastward shift of baroclinic wave 1. Furthermore, the wave-1-like temperature differences between O3R2–3D and O3R1–3D (Figs. 7b,d) show similar results to those in the composite analysis. The anomalous cyclonic flows over Siberia in O3R2–3D forced by the 1990s ozone field

(Fig. 7b) further support that ZAO trends could drive a polar vortex shift toward Siberia. It is worth noting that the warming in the upper stratosphere over Siberia (Figs. 7c,d), which may be caused by enhanced upward propagation of planetary wave, supports the finding that the upper stratospheric temperature changes due to ozone changes is partially caused by dynamics (Keeble et al. 2014). The Plumb flux (see section 2 for more details), which represents well the propagation of planetary waves, is used to analyze wave–mean flow interactions involved in the shift of the polar vortex during late winter. Figures 7e and 7f show that an enhancement of zonal wave 1 exists in the geopotential height anomalies associated with ZAO changes and the zonal wave-1 pattern of geopotential height also shows an eastward shift during strong ZAO events (O3R2–3D) compared to that during weak ZAO events (O3R1–3D), in agreement with the eastward shift of the center of the temperature anomalies (Figs. 7c,d). Correspondingly, the planetary waves within the sector 60° – 180° E show upward propagation (Figs. 7e,f). By contrast, there is less upward propagation of planetary waves within the sector 180° W– 0° .

The above-mentioned results indicate that the polar vortex shift toward Siberia in the 1990s, when the ZAO structure is stronger, is closely related to more upward propagation of planetary wave, particularly for wave 1. To further confirm the role of ZAO in modulating the planetary wave propagation, Fig. 8a shows the differences in the vertical component of the Plumb flux F_z in the upper troposphere and lower stratosphere between the two runs forced with zonal mean ozone fields (i.e., O3R2–2D and O3R1–2D). Note that there are relatively weak negative F_z anomalies over Siberia, opposite to the significant positive F_z anomalies over Siberia between the ZAO runs O3R2–3D and O3R1–3D (Fig. 8b). This result suggests that the upward propagation of planetary waves into the stratosphere over Siberia is stronger in the ZAO runs. The changes in upward propagation of planetary waves are closely related to the wave propagation conditions, which can be described by the refractive index (Andrews et al. 1987). Simpson et al. (2009) noted that the changes in the refractive index at the middle and high latitudes are mainly due to the change in the meridional gradient of PV (PV_y). Figure 8d shows the differences in PV_y between O3R2–3D and O3R1–3D. Note that there is a larger PV_y over Siberia in the O3R2–3D run than in the O3R1–3D run, favorable for the upward propagation of planetary waves over this region (Fig. 8b). By contrast, there are negative PV_y differences over North America between O3R2–3D and O3R1–3D, and hence, weak upward propagation of planetary waves (Fig. 8b). However, the

FIG. 7. Differences of zonal wave-1 temperature (color shaded) and horizontal winds (vectors; m s^{-1}) in the layer 600–1000 K during February between the (a) 1990s and 1980s and (b) runs O3R2–3D and O3R1–3D. Height–longitude cross section of zonal wave-1 temperature differences (color shaded) averaged over 60° – 75°N between the (c) 1990s and 1980s and (d) runs O3R2–3D and O3R1–3D. The climatological mean of zonal wave-1 temperature structure (white contour lines) is also shown. Height–longitude cross section of zonal wave-1 geopotential height differences (color shaded) averaged over 60° – 75°N between the (e) 1990s and 1980s and (f) runs O3R2–3D and O3R1–3D. The vectors in (e) and (f) represent zonal and vertical components of Plumb fluxes ($\text{m}^2 \text{s}^{-2}$), and the vertical component is magnified by a factor of 500. The climatological mean of zonal wave-1 geopotential height structure (white contour lines) is also shown in (e) and (f). The contour intervals in (a)–(f) are 1 m s^{-1} , 0.5 m s^{-1} , 1 K , 0.5 K , 100 m , and 50 m , respectively. The solid and dashed lines denote positive and negative values, respectively. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's t test. The reference vector is shown in the lower right corner.

FIG. 8. Differences of February vertical component of Plumb fluxes in the layer 100–300 hPa between the runs (a) O3R2–2D and O3R1–2D and the runs (b) O3R2–3D and O3R1–3D. Differences of the February meridional PV gradient in the layer 600–1000 K between the runs (c) O3R2–2D and O3R1–2D and the runs (d) O3R2–3D and O3R1–3D. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's *t* test.

PV_y to the north of 60°N over Siberia in the O3R2–2D run without ZAO structure is smaller than that in the O3R1–2D run, consistent with weak upward wave propagation and relatively weak polar vortex shift toward Siberia in O3R2–2D, further confirming that the ZAO can affect the planetary wave propagation and hence the polar vortex position.

Figure 9b shows the differences in horizontal components of the Plumb flux and its horizontal divergence

between O3R2–3D and O3R1–3D. Note that there exist poleward horizontal wave fluxes within the sector 0°–90°E, while the wave fluxes to the east of 90°E show anomalous southeastward propagation, which may result from wave reflection by the circumpolar jet, in the O3R2–3D run compared with those in the O3R1–3D run, leading to a larger Plumb flux divergence around 90°E between 60° and 75°N in the O3R2–3D run. By definition, the Plumb flux divergence is proportional to

FIG. 9. Differences of February horizontal components of Plumb fluxes (vector; $\text{m}^2 \text{s}^{-2}$; the meridional component is magnified by a factor of 5) and its divergence (color shaded) in the layer 600–1000 K between the runs (a) O3R2-2D and O3R1-2D and (b) O3R2-3D and O3R1-3D. Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's t test. The reference vector is shown in the lower right corner.

the meridional PV flux and zonal wind acceleration (Plumb 1985; Holton and Hakim 2013), in agreement with the positive PV anomalies over Siberia and the shift of the polar vortex edge toward Siberia in the O3R2–3D run (Fig. 3c). By contrast, there is a larger Plumb flux convergence over North America in the O3R2–3D run than that in the O3R1–3D run, corresponding to a zonal wind deceleration and a retreat of the polar vortex edge over North America. Figure 9a shows the differences in horizontal components of the Plumb flux and its horizontal divergence between two runs forced with zonal mean ozone fields (i.e., O3R2–2D and O3R1–2D). The magnitude of the Plumb flux divergence differences between the O3R2–2D and O3R1–2D runs is smaller than that between O3R2–3D and O3R1–3D, consistent with a weaker shift of the polar vortex toward Siberia in the zonally symmetric ozone runs (Fig. 5a).

Finally, it should be noted that the magnitude of the ZAO may influence the extent of the polar vortex shift toward Siberia. Figure 10a shows the scaled PV differences between the 2000s and 1980s, and the polar vortex positions in the two decades. Note that the negative PV differences over North America between the 2000s and 1980s are slightly weaker and not significant compared with those between the 1990s and 1980s, although the positive PV differences over Siberia are similar to those in the former period (cf. Figs. 3c, 10a). In addition, the

polar vortex in the O3R3–3D run (green line and cross) shows a less significant shift toward Siberia than that in the O3R2–3D run (blue line and cross) relative to the polar vortex position in the O3R1–3D run (red line and cross). This may be because the magnitude of ZAO differences between the 2000s and 1980s (Fig. 10b) is smaller than those between the 1990s and 1980s (Fig. 2d), which might be related to the stratospheric ozone recovery after 2000 (Chipperfield et al. 2017).

5. Conclusions and discussions

Using WACCM3 forced by 3D ozone data in the 1980s, 1990s, and 2000s, the impacts of ZAO changes on the polar vortex positions are analyzed in this study. We found that, in the experiment with MERRA 3D ozone forcing in the 1990s, there is a significant shift of the polar vortex toward Siberia in February compared with that in the experiment with 1980s ozone forcing (Fig. 3c). It is found the stratospheric ozone in the 1990s shows a more zonally asymmetric structure than that in the 1980s (Figs. 1 and 2). Using reanalysis data to make composite analysis, we confirmed that the polar vortex is shifted toward Siberia during strong ZAO events or in the 1990s compared with that during weak ZAO events or in the 1980s (Figs. 3a,b), supporting the modeled results (Fig. 3c).

FIG. 10. (a) The average of scaled PV differences within the layer 600–1000 K (color shaded) in February between the runs O3R3–3D and O3R1–3D (see section 2 for more details). The color contour lines and color crosses represent the mean polar vortex edges and centers, respectively, in O3R3–3D (green), O3R2–3D (blue), and O3R1–3D (red). (b) Isentropic surface–longitude cross section between 60° and 75°N of ozone differences between the 2000s and the 1980s derived from MERRA reanalysis data. The climatological mean zonal ozone anomaly for the period 1980–99 is also shown (contour lines; interval: 0.1 ppmv; solid and dashed lines represent positive and negative values, respectively). Dotted regions indicate statistical significance at the 95% confidence level according to a two-tailed Student's *t* test.

The latitude and longitude of the polar vortex center in the experiment with 1990s ozone forcing (strong ZAO periods) shows more southward and eastward, respectively, than those in the experiment with 1980s ozone forcing (weak ZAO periods) (Table 1). Furthermore, on

daily time scales, there is a higher occurrence of stratospheric polar vortex fractional area over Siberia in the 1990s than that in the 1980s (Fig. 4). By contrast, the stratospheric polar vortex does not show a shift toward Siberia in the 1990s experiment with zonally mean ozone forcing (Fig. 5a), suggesting that the ZAO impact on the polar vortex position is not negligible. In addition, the shift of the polar vortex toward Siberia in the O3R3–3D run with 2000s ozone forcing is weaker than that in the O3R2–3D run with 1990s ozone forcing, corresponding to a weaker ZAO trend after 2000 (Fig. 10).

The feedback of ZAO on the polar vortex position is mainly achieved through the chemical–radiative–dynamical feedback processes (Fig. 6). In winter, the shortwave cooling (Fig. 6a) and dynamical cooling over Siberia (Fig. 6e) associated with the ZAO changes leads to a compression of isentropic surface layers and anomalous cyclonic flows over Siberia (Figs. 7a,b), corresponding to a polar vortex shift toward Siberia. In addition, a larger meridional PV gradient over Siberia associated with ZAO structure is found in the 1990s run than that in the 1980s run (Fig. 8d), favorable for more upward propagation of planetary waves over Siberia (Fig. 8b), which is not seen in the zonally symmetric ozone runs (Fig. 8a). Enhanced divergence (convergence) of horizontal wave fluxes over Siberia (North America) accelerates (decelerates) the high-latitude zonal westerlies over Siberia (North America) (Fig. 9), further enhancing the polar vortex shift toward Siberia.

Previous studies (e.g., Gabriel et al. 2007; Crook et al. 2008; McCormack et al. 2011) have reported that ZAO variations could reduce to some extent the strength of the stratospheric polar vortex and attributed the weakening of the polar vortex to increases in the amplitude of stratospheric wave 1. The results in the present study support the findings in the abovementioned studies. Furthermore, the present study first reveals that ZAO variations not only can affect the polar vortex strength, but also can modulate the polar vortex position; that is, the ZAO trend characterized by stronger Siberian ozone decline favors a shift of the polar vortex toward Siberia (Fig. 3). We found that this phenomenon is most pronounced in February due to an evident ZAO trend at that month (Figs. 1 and 2).

Zhang et al. (2018, 2019b) demonstrated that the linear trend in TCO for the period 1980–2009 shows a zonally asymmetric structure with a stronger TCO decline over Siberia, which is closely related to the shift of the Arctic stratospheric polar vortex toward the Eurasian continent, particularly in February. It should be pointed out that the results presented here do not contradict the findings of Zhang et al. (2018, 2019b). Instead, we highlight that the ZAO trend induced by the position change of the Arctic polar vortex in turn

provides a positive feedback on the polar vortex position, and this feedback has not been reported in previous literature. The positive feedback between ZAO variations and the polar vortex shift would be an important process accounting for the interactions between stratospheric ozone changes and polar stratospheric dynamics. The results in the present study further suggest that including realistic ZAO variations in climate models is essential for a better simulation of the climate system.

Finally, it should be pointed out that the version of WACCM used in this study does not include the turbulent mountain stress parameterization that is used in more recent versions of the model (Richter et al. 2010). The inclusion of this parameterization improved the planetary wave propagation and led to a larger number of sudden stratospheric warmings, which suggests that the newer WACCM versions would have a generally more disturbed polar stratospheric vortex. The sensitivity of our results related to different climate models deserves more research in the future.

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REFERENCES

- Albers, J. R., and T. R. Nathan, 2012: Pathways for communicating the effects of stratospheric ozone to the polar vortex: Role of zonally asymmetric ozone. *J. Atmos. Sci.*, **69**, 785–801, <https://doi.org/10.1175/JAS-D-11-0126.1>.
- , and —, 2013: Ozone loss and recovery and preconditioning of upward propagating planetary wave activity. *J. Atmos. Sci.*, **70**, 3977–3994, <https://doi.org/10.1175/JAS-D-12-0259.1>.
- Andrews, D. G., C. B. Leovy, and J. R. Holton, 1987: *Middle Atmosphere Dynamics*. Academic Press, 489 pp.
- Birner, T., and E. J. Charlesworth, 2017: On the relative importance of radiative and dynamical heating for tropical tropopause temperatures. *J. Geophys. Res. Atmos.*, **122**, 6782–6797, <https://doi.org/10.1002/2016JD026445>.
- Bohlinger, P., B. M. Sinnhuber, R. Ruhnke, and O. Kirner, 2014: Radiative and dynamical contributions to past and future Arctic stratospheric temperature trends. *Atmos. Chem. Phys.*, **14**, 1679–1688, <https://doi.org/10.5194/acp-14-1679-2014>.
- Calvo, N., L. M. Polvani, and S. Solomon, 2015: On the surface impact of Arctic stratospheric ozone extremes. *Environ. Res. Lett.*, **10**, 094003, <https://doi.org/10.1088/1748-9326/10/9/094003>.
- Cheung, J. C. H., J. D. Haigh, and D. R. Jackson, 2014: Impact of EOS MLS ozone data on medium-extended range ensemble weather forecasts. *J. Geophys. Res. Atmos.*, **119**, 9253–9266, <https://doi.org/10.1002/2014JD021823>.
- Chipperfield, M. P., and J. A. Pyle, 1988: Two-dimensional modeling of the Antarctic lower stratosphere. *Geophys. Res. Lett.*, **15**, 875–878, <https://doi.org/10.1029/GL015i008p00875>.
- , and Coauthors, 2017: Detecting recovery of the stratospheric ozone layer. *Nature*, **549**, 211–218, <https://doi.org/10.1038/nature23681>.
- Crook, J. A., N. P. Gillett, and S. P. E. Keeley, 2008: Sensitivity of Southern Hemisphere climate to zonal asymmetry in ozone. *Geophys. Res. Lett.*, **35**, L07806, <https://doi.org/10.1029/2007GL032698>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Dunkerton, T. J., and D. P. Delisi, 1986: Evolution of potential vorticity in the winter stratosphere of January–February 1979. *J. Geophys. Res.*, **91**, 1199–1208, <https://doi.org/10.1029/JD091iD01p01199>.
- Eyring, V., and Coauthors, 2006: Assessment of temperature, trace species, and ozone in chemistry–climate model simulations of the recent past. *J. Geophys. Res.*, **111**, D22308, <https://doi.org/10.1029/2006JD007327>.
- Forster, P. M. D., and K. P. Shine, 1997: Radiative forcing and temperature trends from stratospheric ozone changes. *J. Geophys. Res.*, **102**, 10 841–10 855, <https://doi.org/10.1029/96JD03510>.
- Gabriel, A., D. Peters, I. Kirchner, and H. F. Graf, 2007: Effect of zonally asymmetric ozone on stratospheric temperature and planetary wave propagation. *Geophys. Res. Lett.*, **34**, L06807, <https://doi.org/10.1029/2006GL028998>.
- Garcia, R. R., D. R. Marsh, D. E. Kinnison, B. A. Boville, and F. Sassi, 2007: Simulation of secular trends in the middle atmosphere, 1950–2003. *J. Geophys. Res.*, **112**, D09301, <https://doi.org/10.1029/2006JD007485>.
- Garfinkel, C. I., 2017: Might stratospheric variability lead to improved predictability of ENSO events? *Environ. Res. Lett.*, **12**, 031001, <https://doi.org/10.1088/1748-9326/aa60a4>.
- Gillett, N. P., J. F. Scinocca, D. A. Plummer, and M. C. Reader, 2009: Sensitivity of climate to dynamically-consistent zonal asymmetries in ozone. *Geophys. Res. Lett.*, **36**, L10809, <https://doi.org/10.1029/2009GL037246>.
- Haase, S., and K. Matthes, 2019: The importance of interactive chemistry for stratosphere–troposphere coupling. *Atmos. Chem. Phys.*, **19**, 3417–3432, <https://doi.org/10.5194/acp-19-3417-2019>.
- Harari, O., and Coauthors, 2019: Influence of Arctic stratospheric ozone on surface climate in CCMI models. *Atmos. Chem. Phys.*, **19**, 9253–9268, <https://doi.org/10.5194/acp-19-9253-2019>.
- Holton, J. R., and G. J. Hakim, 2013: *An Introduction to Dynamic Meteorology*. Academic Press, 532 pp.
- Hu, D. Z., and Z. Guan, 2018: Decadal relationship between the stratospheric Arctic vortex and Pacific decadal oscillation.

- J. Climate*, **31**, 3371–3386, <https://doi.org/10.1175/JCLI-D-17-0266.1>.
- , W. S. Tian, F. Xie, C. X. Wang, and J. K. Zhang, 2015: Impacts of stratospheric ozone depletion and recovery on wave propagation in the boreal winter stratosphere. *J. Geophys. Res. Atmos.*, **120**, 8299–8317, <https://doi.org/10.1002/2014JD022855>.
- , Z. Y. Guan, W. S. Tian, and R. C. Ren, 2018: Recent strengthening of the stratospheric Arctic vortex response to warming in the central North Pacific. *Nat. Commun.*, **9**, 1697, <https://doi.org/10.1038/s41467-018-04138-3>.
- , Y. P. Guo, and Z. Y. Guan, 2019a: Recent weakening in the stratospheric planetary wave intensity in early winter. *Geophys. Res. Lett.*, **46**, 3953–3962, <https://doi.org/10.1029/2019GL082113>.
- , Z. Guan, Y. Guo, L. Chu, and D. Jin, 2019b: Dynamical connection between the stratospheric Arctic vortex and sea surface temperatures in the North Atlantic. *Climate Dyn.*, **53**, 6979–6993, <https://doi.org/10.1007/s00382-019-04971-2>.
- Huang, J. L., W. S. Tian, L. J. Gray, J. K. Zhang, Y. Li, J. L. Luo, and H. Y. Tian, 2018: Preconditioning of Arctic stratospheric polar vortex shift events. *J. Climate*, **31**, 5417–5436, <https://doi.org/10.1175/JCLI-D-17-0695.1>.
- Ivy, D. J., S. Solomon, and H. E. Rieder, 2016: Radiative and dynamical influences on polar stratospheric temperature trends. *J. Climate*, **29**, 4927–4938, <https://doi.org/10.1175/JCLI-D-15-0503.1>.
- Karpechko, A. Y., J. Perlwitz, and E. Manzini, 2014: A model study of tropospheric impacts of the Arctic ozone depletion 2011. *J. Geophys. Res. Atmos.*, **119**, 7999–8014, <https://doi.org/10.1002/2013JD021350>.
- Keeble, J., P. Braesicke, N. L. Abraham, H. K. Roscoe, and J. A. Pyle, 2014: The impact of polar stratospheric ozone loss on Southern Hemisphere stratospheric circulation and climate. *Atmos. Chem. Phys.*, **14**, 13 705–13 717, <https://doi.org/10.5194/acp-14-13705-2014>.
- Lin, P., D. Paynter, L. Polvani, G. J. P. Correa, Y. Ming, and V. Ramaswamy, 2017: Dependence of model-simulated response to ozone depletion on stratospheric polar vortex climatology. *Geophys. Res. Lett.*, **44**, 6391–6398, <https://doi.org/10.1002/2017GL073862>.
- Mai, Y., Z. Sheng, H. Shi, Q. Liao, and W. Zhang, 2020: Spatiotemporal distribution of atmospheric ducts in Alaska and its relationship with the Arctic vortex. *Int. J. Antenn. Propag.*, **2020**, 9673289, <https://doi.org/10.1155/2020/9673289>.
- Manney, G. L., and Coauthors, 2011: Unprecedented Arctic ozone loss in 2011. *Nature*, **478**, 469–475, <https://doi.org/10.1038/nature10556>.
- Manzini, E., B. Steil, C. Brühl, M. A. Giorgetta, and K. Krüger, 2003: A new interactive chemistry-climate model: 2. Sensitivity of the middle atmosphere to ozone depletion and increase in greenhouse gases and implications for recent stratospheric cooling. *J. Geophys. Res.*, **108**, 4429, <https://doi.org/10.1029/2002JD002977>.
- McCormack, J. P., T. R. Nathan, and E. C. Cordero, 2011: The effect of zonally asymmetric ozone heating on the Northern Hemisphere winter polar stratosphere. *Geophys. Res. Lett.*, **38**, L03802, <https://doi.org/10.1029/2010GL045937>.
- McLandress, C., A. I. Jonsson, D. A. Plummer, M. C. Reader, J. F. Scinocca, and T. G. Shepherd, 2010: Separating the dynamical effects of climate change and ozone depletion. Part I: Southern Hemisphere stratosphere. *J. Climate*, **23**, 5002–5020, <https://doi.org/10.1175/2010JCLI3586.1>.
- Nash, E. R., P. A. Newman, J. E. Rosenfield, and M. R. Schoeberl, 1996: An objective determination of the polar vortex using Ertel's potential vorticity. *J. Geophys. Res.*, **101**, 9471–9478, <https://doi.org/10.1029/96JD00066>.
- Nowack, P. J., N. L. Abraham, A. C. Maycock, P. Braesicke, J. M. Gregory, M. M. Joshi, A. Osprey, and J. A. Pyle, 2015: A large ozone-circulation feedback and its implications for global warming assessments. *Nat. Climate Change*, **5**, 41–45, <https://doi.org/10.1038/nclimate2451>.
- Peters, D. H. W., A. Schneider, M. Bugelmayer, C. Zulicke, and I. Kirchner, 2015: Atmospheric circulation changes in response to an observed stratospheric zonal ozone anomaly. *Atmos.–Ocean*, **53**, 74–88, <https://doi.org/10.1080/07055900.2013.878833>.
- Plumb, R. A., 1985: On the three-dimensional propagation of stationary waves. *J. Atmos. Sci.*, **42**, 217–229, [https://doi.org/10.1175/1520-0469\(1985\)042<0217:OTDPO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<0217:OTDPO>2.0.CO;2).
- Polvani, L. M., D. W. Waugh, G. J. P. Correa, and S. W. Son, 2011: Stratospheric ozone depletion: The main driver of twentieth-century atmospheric circulation changes in the Southern Hemisphere. *J. Climate*, **24**, 795–812, <https://doi.org/10.1175/2010JCLI3772.1>.
- Rae, C. D., J. Keeble, P. Hitchcock, and J. A. Pyle, 2019: Prescribing zonally asymmetric ozone climatologies in climate models: Performance compared to a chemistry-climate model. *J. Adv. Model. Earth Syst.*, **11**, 918–933, <https://doi.org/10.1029/2018MS001478>.
- Ramaswamy, V., M. D. Schwarzkopf, and W. J. Randel, 1996: Fingerprint of ozone depletion in the spatial and temporal pattern of recent lower-stratospheric cooling. *Nature*, **382**, 616–618, <https://doi.org/10.1038/382616a0>.
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, <https://doi.org/10.1029/2002JD002670>.
- Richter, J. H., F. Sassi, and R. R. Garcia, 2010: Toward a physically based gravity wave source parameterization in a general circulation model. *J. Atmos. Sci.*, **67**, 136–156, <https://doi.org/10.1175/2009JAS3112.1>.
- Rienecker, M. M., and Coauthors, 2011: MERRA: NASA's Modern-Era Retrospective Analysis for Research and Applications. *J. Climate*, **24**, 3624–3648, <https://doi.org/10.1175/JCLI-D-11-00015.1>.
- Sassi, F., B. A. Boville, D. Kinnison, and R. R. Garcia, 2005: The effects of interactive ozone chemistry on simulations of the middle atmosphere. *Geophys. Res. Lett.*, **32**, L07811, <https://doi.org/10.1029/2004GL022131>.
- Sexton, D. M. H., 2001: The effect of stratospheric ozone depletion on the phase of the Antarctic Oscillation. *Geophys. Res. Lett.*, **28**, 3697–3700, <https://doi.org/10.1029/2001GL013376>.
- Shindell, D. T., and G. A. Schmidt, 2004: Southern Hemisphere climate response to ozone changes and greenhouse gas increases. *Geophys. Res. Lett.*, **31**, L18209, <https://doi.org/10.1029/2004GL020724>.
- , D. Rind, N. Balachandran, J. Lean, and P. Lonergan, 1999: Solar cycle variability, ozone, and climate. *Science*, **284**, 305–308, <https://doi.org/10.1126/science.284.5412.305>.
- Simpson, I. R., M. Blackburn, and J. D. Haigh, 2009: The role of eddies in driving the tropospheric response to stratospheric heating perturbations. *J. Atmos. Sci.*, **66**, 1347–1365, <https://doi.org/10.1175/2008JAS2758.1>.

- Smith, K. L., and L. M. Polvani, 2014: The surface impacts of Arctic stratospheric ozone anomalies. *Environ. Res. Lett.*, **9**, 074015, <https://doi.org/10.1088/1748-9326/9/7/074015>.
- Son, S. W., and Coauthors, 2008: The impact of stratospheric ozone recovery on the Southern Hemisphere westerly jet. *Science*, **320**, 1486–1489, <https://doi.org/10.1126/science.1155939>.
- Stocker, T. F., and Coauthors, 2001: Physical climate processes and feedbacks. *Climate Change 2001: The Scientific Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 419–457.
- Thompson, D. W. J., and S. Solomon, 2005: Recent stratospheric climate trends as evidenced in radiosonde data: Global structure and tropospheric linkages. *J. Climate*, **18**, 4785–4795, <https://doi.org/10.1175/JCLI3585.1>.
- , —, P. J. Kushner, M. H. England, K. M. Grise, and D. J. Karoly, 2011: Signatures of the Antarctic ozone hole in Southern Hemisphere surface climate change. *Nat. Geosci.*, **4**, 741–749, <https://doi.org/10.1038/ngeo1296>.
- Tian, W. S., and M. P. Chipperfield, 2005: A new coupled chemistry–climate model for the stratosphere: The importance of coupling for future O₃–climate predictions. *Quart. J. Roy. Meteor. Soc.*, **131**, 281–303, <https://doi.org/10.1256/qj.04.05>.
- Wang, W., W. Tian, S. Dhomse, E. Xie, J. Shu, and J. Austin, 2014: Stratospheric ozone depletion from future nitrous oxide increases. *Atmos. Chem. Phys.*, **14**, 12 967–12 982, <https://doi.org/10.5194/acp-14-12967-2014>.
- Waugh, D. W., L. Oman, P. A. Newman, R. S. Stolarski, S. Pawson, J. E. Nielsen, and J. Perlwitz, 2009: Effect of zonal asymmetries in stratospheric ozone on simulated Southern Hemisphere climate trends. *Geophys. Res. Lett.*, **36**, L18701, <https://doi.org/10.1029/2009GL040419>.
- WMO, 2003: Scientific assessment of ozone depletion: 2002. Global Ozone Research and Monitoring Project Rep. 47, 490 pp.
- Xie, F., W. S. Tian, and M. P. Chipperfield, 2008: Radiative effect of ozone change on stratosphere–troposphere exchange. *J. Geophys. Res.*, **113**, D00B09, <https://doi.org/10.1029/2008JD009829>.
- , and Coauthors, 2016: A connection from Arctic stratospheric ozone to El Niño–Southern Oscillation. *Environ. Res. Lett.*, **11**, 124026, <https://doi.org/10.1088/1748-9326/11/12/124026>.
- , J. Zhang, W. Sang, Y. Li, Y. Qi, C. Sun, Y. Li, and J. Shu, 2017: Delayed effect of Arctic stratospheric ozone on tropical rainfall. *Atmos. Sci. Lett.*, **18**, 409–416, <https://doi.org/10.1002/asl.783>.
- Zhang, J. K., W. S. Tian, M. P. Chipperfield, F. Xie, and J. L. Huang, 2016: Persistent shift of the Arctic polar vortex towards the Eurasian continent in recent decades. *Nat Climate Change*, **6**, 1094–1099, <https://doi.org/10.1038/nclimate3136>.
- , and Coauthors, 2018: Stratospheric ozone loss over the Eurasian continent induced by the polar vortex shift. *Nat. Commun.*, **9**, 206, <https://doi.org/10.1038/S41467-017-02565-2>.
- , F. Xie, Z. C. Ma, C. Y. Zhang, M. Xu, T. Wang, and R. H. Zhang, 2019a: Seasonal evolution of the quasi-biennial oscillation impact on the Northern Hemisphere polar vortex in winter. *J. Geophys. Res. Atmos.*, **124**, 12 568–12 586, <https://doi.org/10.1029/2019JD030966>.
- , W. Tian, F. Xie, W. Sang, D. Guo, M. Chipperfield, W. Feng, and D. Hu, 2019b: Zonally asymmetric trends of winter total column ozone in the northern middle latitudes. *Climate Dyn.*, **52**, 4483–4500, <https://doi.org/10.1007/s00382-018-4393-y>.