The effect of interactive ozone chemistry on weak and strong stratospheric polar vortex events

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Abstract. Modeling and observational studies have reported effects of stratospheric ozone extremes on Northern Hemisphere spring climate. Recent work has further suggested that the coupling of ozone chemistry and dynamics amplifies the surface response to midwinter sudden stratospheric warmings (SSWs). Here we study the importance of interactive ozone chemistry in representing the stratospheric polar vortex and Northern Hemisphere winter surface climate variability. We contrast two simulations from the interactive and specified chemistry (and thus ozone) versions of the Whole Atmosphere Community Climate Model, which is designed to isolate the impact of interactive ozone on polar vortex variability. In particular, we analyze the response with and without interactive chemistry to midwinter SSWs, March SSWs, and strong polar vortex events (SPVs). With interactive chemistry, the stratospheric polar vortex is stronger and more SPVs occur, but we find little effect on the frequency of midwinter SSWs. At the surface, interactive chemistry results in a pattern resembling a more negative North Atlantic Oscillation following midwinter SSWs but with little impact on the surface signatures of late winter SSWs and SPVs. These results suggest that including interactive ozone chemistry is important for representing North Atlantic and European winter climate variability.

1 Introduction

The climate impacts of stratospheric ozone extremes, particularly Antarctic ozone depletion, have been widely studied (Previdi and Polvani, 2014, and references therein). While the effects are clearer and larger in the Southern Hemisphere, ozone extremes have also been shown to be associated with springtime surface anomalies in the Northern Hemisphere (Smith and Polvani, 2014; Calvo et al., 2015; Ivy et al., 2017).

Polar cap ozone anomalies are strongly related to interannual variability in stratospheric polar vortex strength, which is larger in the Northern Hemisphere than the Southern Hemisphere. This is a result of the larger amplitudes of upward propagating planetary waves which perturb the stratospheric circulation. Years with low wave activity tend to correspond to a stronger vortex and a weaker Brewer–Dobson circulation (BDC), resulting in weaker ozone transport from the tropics into the poles and decreased mixing across the vortex edge, as well as the enhanced formation of polar stratospheric clouds, which contribute to increased springtime destruction of ozone. Years with high wave activity correspond to a weaker vortex and a stronger BDC with stronger ozone transport from the tropics and increased mixing (Newman et al., 2001).

These processes are well represented in fully interactive chemistry–climate models (Strahan and Douglass, 2004). However, such models are computationally expensive compared to the more common ones in which stratospheric ozone is simply prescribed. A number of studies have explored the importance of interactive ozone chemistry on model representations of coupled stratosphere–troposphere variability. Smith and Polvani (2014) and Karpechko et al. (2014) found little impact of stratospheric ozone extremes on surface climate in the Northern Hemisphere using prescribed zonal mean monthly mean ozone fields. However, Calvo
et al. (2015) found robust surface impacts associated with stratospheric ozone extremes using an interactive chemistry–climate model, suggesting the potential importance of this coupling. Further model studies are needed to disentangle the effects of ozone from those of polar vortex variability.

While the effect of polar stratospheric clouds on ozone is mainly seen in the spring when sunlight returns to the region, the variability of the polar vortex can result in wintertime ozone anomalies which may have surface impacts. The most extreme states of the polar vortex are sudden stratospheric warmings (SSWs) and strong polar vortex events (SPVs). We define these precisely in Sect. 2 based on extreme values of zonal mean zonal wind. Leading up to an SSW, dynamical forcing disrupts the stratospheric circulation, eventually resulting in a reversal of zonal mean zonal wind throughout much of the polar stratosphere. SSWs have surface effects for the 2 months following in particular a negative North Atlantic Oscillation (NAO) and cold anomalies over much of northern Eurasia. Conversely, SPVs in which abnormally strong westerly zonal mean zonal winds occur are the result of anomalously weak planetary wave activity over a protracted period. As such, they are not rapid dynamical events in the same way as SSWs, but they may still have surface impacts, which are typically a positive North Atlantic Oscillation (Baldoand Dunkerton, 2001).

For the dynamical reasons described above, SSWs and SPVs tend to be associated with the occurrence of positive and negative stratospheric ozone anomalies, respectively. About 2 weeks prior to an SSW, the BDC accelerates, resulting in adiabatic warming of the stratosphere and enhanced isentropic eddy transport of ozone and thus increased ozone concentration over the pole (de la Cámara et al., 2018). SPVs are similarly accompanied by an anomalously weak BDC because of the lack of planetary wave activity and thus an anomalously low transport of ozone as well.

Because they affect both stratospheric ozone and the NAO in the troposphere, extreme vortex events offer an ideal case in which to study wintertime surface impacts of ozone chemistry. Haase and Matthes (2019) studied the impact of interactive versus prescribed ozone on SSWs, as well as their surface effects, in simulations of the recent past (1955–present) in an earth system model. They compared results of a simulation with interactive ozone to those of a simulation with prescribed ozone. This prescribed ozone was given daily (with no averaging or climatology) from a single historical integration using the “specified chemistry” version of WACCM, known as SC-WACCM (Smith et al., 2014). We refer to this pre-
scribed chemistry simulation as the NOCHEM simulation in the analysis. In the NOCHEM simulation, ozone concentrations (and other radiatively active atmospheric constituents, including chlorofluorocarbons) are prescribed using zonally symmetric, monthly mean, seasonal climatology computed from the WACCM integration. These zonally symmetric monthly ozone fields are read into SC-WACCM and interpolated linearly to the day of the year. More details can be found in Smith et al. (2014). Hence, both CHEM and NOCHEM strictly impose identical year 2000 forcings for all radiatively active species, and only differ in their treatment of ozone. The use of climatological ozone fields in NOCHEM removes the effect of extreme ozone variations on the climate system. One might consider specifying non-zonally symmetric ozone (Haase and Matthes, 2019), but that comes at the cost of a major physical inconsistency between the polar vortex and the ozone field; in other words, the extreme ozone years in the model will not correspond with the unperturbed vortex years. More importantly, the vast majority of climate models in the Coupled Model Intercomparison Project (CMIP) specify zonally symmetric stratospheric ozone, including within CMIP6 (Keeble et al., 2020); hence, the zonally symmetric specified ozone case is the one of most interest in terms of evaluating the impact of interactive ozone chemistry.

We identify SSWs in the model output following the definition in Charlton and Polvani (2007a) (see the corrigendum Charlton-Perez and Polvani, 2011). We define an SSW as a reversal of zonal mean zonal wind at 60° N and 10 hPa from westerly to easterly from November through March, with the central date being the first day of easterly zonal mean zonal winds. No later date can be a central date until the winds have been westerly again for at least 20 d, and the winds must return to westerly for at least 10 consecutive days before 30 April (thus discarding stratospheric final warmings). This definition is optimal for identifying SSWs, as described by Butler and Gerber (2018). We focus on SSWs occurring in December–February and in March. We consider March events separately from December–February events due to different shortwave heating behavior, model bias in March SSW frequency (too frequent SSWs in our model), and different NAO structure in early spring compared to winter.

To the best of our knowledge, there is no standard definition of an SPV. Different methods have been used in the literature (Limpasuvan et al., 2004; Tripathi et al., 2015; Scaife et al., 2016; Beerli and Grams, 2019). We here follow the definition used in Scaife et al. (2016) and Smith et al. (2018), which is designed to be analogous to the SSW definition of Charlton and Polvani (2007a) and to result in a similar number of events in reanalysis. We define an SPV as zonal mean zonal wind at 60° N and 10 hPa reaching 48 m s\(^{-1}\) or higher (westerly) from November through March, with the central date being the first day of zonal mean zonal winds above 48 m s\(^{-1}\). No later date can be a central date until the winds return below 48 m s\(^{-1}\) for at least 20 consecutive days. We focus on SPVs occurring in December–February due to low event frequency in November and March. A separate analysis reveals that results are not sensitive to using a 41.2 m s\(^{-1}\) threshold as in Tripathi et al. (2015).

The results we present here are based on composites of daily model output for climate variables with composites centered around SSW or SPV central dates. For composites from either CHEM or NOCHEM simulations, we calculate significance using a Monte Carlo test based on 5000 randomly chosen central dates. We also consider the difference in CHEM or NOCHEM composites, which is denoted CHEM-NOCHEM; for these, we calculate significance from a two-sided, two-sample \(t\) test.

3 Impact of interactive chemistry

3.1 Stratospheric mean state and extreme events

We first consider the effect of interactive chemistry on the mean state of the stratosphere by examining the climatological Northern Hemisphere 10 hPa zonal mean zonal wind (\(U\); Fig. 1). We find stronger westerlies in CHEM than in NOCHEM in the vortex formation stage (September and early October) and in the latter half of winter (January–April) between 60 and 80° N. In line with this, we also find weaker downwelling in winter in the upper latitudes in CHEM than in NOCHEM (not shown). This relative strength in CHEM in late winter also corresponds to a delayed final warming by 7 d on average. These results are in agreement with those of Haase and Matthes (2019). We also found similar results in six 1955–2005 historical integrations of WACCM and SC-WACCM (with ozone specified monthly or daily from the WACCM climatology) from Neely et al. (2014) (not shown), further indicating that this feature is robust.

This is not the case in Smith et al. (2014) in which the vortex is of similar strength with interactive and prescribed chemistry.
ozone under constant year 1850 conditions. The difference
between that study and ours is the level of chlorofluorocar-
bons (CFCs); these are 0 in Smith et al. (2014), which simu-
lates preindustrial conditions, but they are substantial in our
study, which simulates year 2000 conditions. They are simi-
larly substantial in the historical (1955–present day) simula-
tions in Haase and Matthes (2019) and Neely et al. (2014).
Because the differences between interactive and specified
ozone simulations depend on the level of CFCs, a precise
understanding of the mechanisms for the difference will
require disentangling the dynamics and chemistry. Higher
ozone variability in the presence of CFCs (Calvo et al., 2015)
might increase the effects of the ozone-dynamic feedbacks,
rendering this a very difficult problem. There are indications
that these differences may be related to the zonal asymme-
try of ozone (Haase and Matthes, 2019), further complicating
the relationship. Albers and Nathan (2012) have proposed a
complex mechanism to detail the coupling of zonally asym-
metric ozone and dynamics in the context of a highly ideal-
ized linear model. In their model, zonal asymmetries in
ozone precondition the waves, causing a reduction in plan-
etary wave drag and a colder polar vortex. However, deter-
mining whether this mechanism is operative in our compre-
hensive model would be quite difficult as the mechanism
relies on many assumptions that are likely inapplicable in
the presence of highly nonlinear, time-dependent breaking
waves which are observed in the winter polar stratosphere
in a fully interactive model.

Because we identify extreme stratospheric events using
zonal mean zonal winds at 10 hPa and 60° N (U1060) (Charl-
ton and Polvani, 2007a; Butler and Gerber, 2018), we next
examine the mean state and variability of this quantity in
CHEM and NOCHEM. Figure 2 shows the two distributions
of U1060 from December through March (DJFM). The av-

erage difference in DJFM between CHEM and NOCHEM is
about 1.7 m s⁻¹. To determine whether this is statistically
significant, we consider the average zonal mean zonal winds
over each winter and treat the winters as independent. A two-
tailed, two-sample Welch’s t test of DJFM average winds in
CHEM and NOCHEM yields a p value of 0.023, so the
difference, though small, is significant at a 95% level. The
CHEM distribution also has a longer right tail, which is con-
sistent with the polar vortex being stronger overall with in-
teractive chemistry. It also indicates that we should expect
more SPVs in CHEM than in NOCHEM. While there are
fewer days of weak westerlies (0–20 m s⁻¹) in CHEM than
in NOCHEM, the number of days of easterlies is similar, so
we expect less of a difference in SSW frequency between the
two simulations.

Indeed, this is what we find when we calculate the fre-
cuencies of weak and strong vortex events in the CHEM and
NOCHEM simulations (Table 1). We consider December–
February (DJF; midwinter) and March (late winter) sepa-
rately for two reasons. First, the ozone impacts in midwinter
are different from those in late winter/early spring, as short-

![Figure 2. Histogram of daily values of zonal mean zonal wind at 10 hPa and 60° N in December–March for CHEM and NOCHEM.](image)

**Table 1.** Summary of sudden stratospheric warming (SSW) and strong polar vortex (SPV) events in 200-year year 2000 time slices with and without interactive chemistry (CHEM and NOCHEM, re-

spectively). We separately consider the events occurring from De-

cember through February and those occurring in March. Reported

<table>
<thead>
<tr>
<th></th>
<th>NOCHEM</th>
<th>CHEM</th>
<th>Percent difference</th>
<th>p value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total winters</td>
<td>200</td>
<td>200</td>
<td>0.0 %</td>
<td>0.00</td>
</tr>
<tr>
<td>DJF SSW events</td>
<td>75</td>
<td>67</td>
<td>-10.7 %</td>
<td>0.45</td>
</tr>
<tr>
<td>DJF SPV events</td>
<td>58</td>
<td>74</td>
<td>+29.3 %</td>
<td>0.13</td>
</tr>
<tr>
<td>March SSW events</td>
<td>28</td>
<td>39</td>
<td>+39.3 %</td>
<td>0.14</td>
</tr>
<tr>
<td>March SPV events</td>
<td>7</td>
<td>5</td>
<td>-28.6 %</td>
<td>0.58</td>
</tr>
</tbody>
</table>

wave effects become important in spring. Second, our model
is biased in March with too many SSWs compared to reanal-
ysis, a feature also seen in more recent versions of this model
(Gettelman et al., 2019). We see 1.4 March SSWs per decade
in NOCHEM and 1.95 March SSWs per decade in CHEM
compared to 0.87–1.1 per decade in the reanalysis (But-
ler et al., 2017).

The stronger vortex in midwinter in the CHEM simulation
might lead us to expect fewer DJF SSWs in CHEM than in
NOCHEM. We do see a decrease of about 10% in DJF SSWs
with interactive chemistry compared to specified chemistry,
but this decrease is far from being statistically significant.
In contrast, in March, we see more SSWs in CHEM than in
NOCHEM, potentially related to the later breakdown of the
vortex.

Haase and Matthes (2019) consider the overall (November–March) number of SSWs. They report a
the central date, but in the CHEM simulation the negative
have comparable NAM anomalies in the stratosphere around
Fig. 4. The CHEM and NOCHEM composites around SSWs
appendix A. We show the results of the NAM calculations in
ber and Martineau (2018); the detailed procedure is in Ap-
a method similar to that of Gerber et al. (2010) and Ger-
Northern Annular Mode (NAM) for each simulation. We use
ferences originating in the stratosphere, we calculate the
low and more persistent in CHEM than in
the surface signature
significant and strongly projects onto the NAO 30–60 d after
stronger signal. However, the difference is statistically sig-
first 30 d, with the CHEM simulation having only a slightly
is minimal difference between the two simulations in the
SSWs in CHEM (W ACCM; a) and NOCHEM (SC-WACC; b)
simulations, as well as the difference in the CHEM and NOCHEM
composites (c). Significance at the 95 % level using a Monte Carlo
test (a, b) or a two-sided t test (c) is indicated by stippling.
The number of events included in each composite is noted in brackets
above the figures.

3.2 Midwinter sudden stratospheric warmings

We start by focusing on the surface impacts of SSWs, seek-
ing to document any differences between the CHEM and
NOCHEM simulations. After noting the impact of the events
on the surface, we then consider how any differences in those
impacts arise.

Figure 3 shows composite surface level pressure anom-
lies in the first and second months (top and bottom, respec-
tively) following December–February SSWs in CHEM (Fig.
3a; 75 events) and NOCHEM (Fig. 3b; 67 events), as well
as the difference between the two (Fig. 3c). We see a strong
and significant pattern resembling a negative North Atlantic
Oscillation in the first month following SSWs in both CHEM
and NOCHEM, and in both cases this negative annular mode
persists through the second month following the event. There
is minimal difference between the two simulations in the
first 30 d, with the CHEM simulation having only a slightly
stronger signal. However, the difference is statistically sig-
ificant and strongly projects onto the NAO 30–60 d after
the central date. This indicates that the surface signature
of SSWs is stronger and more persistent in CHEM than in
NOCHEM.

To determine whether the differences at the surface fol-
lowing SSWs in CHEM and NOCHEM are a result of dif-
ferences originating in the stratosphere, we calculate the
Northern Annular Mode (NAM) for each simulation. We use
a method similar to that of Gerber et al. (2010) and Ger-
ber and Martineau (2018); the detailed procedure is in Ap-
pendix A. We show the results of the NAM calculations in
Fig. 4. The CHEM and NOCHEM composites around SSWs
have comparable NAM anomalies in the stratosphere around
the central date, but in the CHEM simulation the negative
anomaly persists more strongly in the lower stratosphere be-
yond 40 d after the central date. The CHEM-NOCHEM dif-
ference shows that this change in persistence with interactive
chemistry is significant at the 95 % level. There is also more
descent of the anomaly to the surface in the CHEM simula-
tion especially at about 30 d after the central date.

This difference in descent is also seen in the CHEM-
NOCHEM temperature anomalies (Fig. 5a). The warming
in the stratosphere associated with the onset of the SSW is
larger with interactive chemistry. This stratospheric tempera-
ture anomaly then descends more strongly through the strato-
sphere and troposphere in the CHEM simulation than in the
NOCHEM simulation.

We investigate the processes leading to these changes in
more detail by examining the dynamical, longwave, and
shortwave heating terms. The greater warming through-
out the stratosphere is due to increased dynamical heating
(Fig. 5b) in CHEM compared to NOCHEM, as the higher
temperature with interactive chemistry is also associated with
a longwave cooling response (Fig. 5c). The higher strato-
spheric temperatures result in greater longwave emission.
The increase in dynamical forcing also corresponds to in-
creased ozone transport. Ozone is a longwave emitter, so the
increased dynamical forcing could directly account for part of
this longwave cooling difference as well.

The increased dynamical heating in CHEM could be re-
lated to the greater wave activity necessary for an SSW to
occur with a stronger mean vortex state. Figure 6 shows the
eddy heat flux over 40–80°N over time in CHEM
and NOCHEM. This is stronger by about 2 mK s$^{-1}$ around the central date in CHEM than in NOCHEM, indicating a slightly stronger wave forcing in CHEM. The CHEM and NOCHEM means are at the upper and lower bounds of the other’s confidence intervals, respectively. Further, the zonal mean zonal winds at 10 hPa and 60° N around the central date of the SSW (shown in Fig. 7) are both stronger prior to the event and more easterly following the central date in CHEM than in NOCHEM. However, the residual vertical velocity anomalies leading up to SSWs are nearly identical for CHEM and NOCHEM (not shown), so the increased dynamical heating in CHEM might be a result of a stronger vertical temperature gradient related to the stronger vortex in this simulation (associated with a colder polar stratosphere).

In DJF, the dynamical heating and the longwave heating are the dominant temperature tendency terms. There is also a significant shortwave heating response (Fig. 5d), but in midwinter it is 1 order of magnitude smaller than the other terms owing to the absence of incoming solar radiation to polar night. The structure in height and time is related to integrated effects of the ozone anomalies following the SSW, which show a similar structure (Kiesewetter et al., 2010). The importance of the shortwave response increases the later in winter the SSW events occur. We illustrate this in Fig. S1 in the Supplement, which shows much stronger differences in CHEM and NOCHEM shortwave anomalies for February SSWs than for December or January events.

Finally, we examine the anomaly in the total ozone column around the central date of the SSW (Fig. 7) in the CHEM simulation. We see a sharp increase in ozone in the 15 d leading up to the central date, reaching a peak of on average about 40 Dobson units above climatology just after the central date, similar to that seen in reanalysis and a similar model by de la Cámara et al. (2018). This ozone anomaly results from transport due to the greater dynamical forcing in CHEM, as noted earlier. Following the central date, anomalies of about 20 Dobson units persist for up to 3 months following the central date. This ozone anomaly is consistent with the total ozone column in analyses and a similar model (de la Cámara et al., 2018) and the smaller ozone depletion in years with early SSWs observed by Strahan et al. (2016).

In summary, DJF SSWs are preceded by larger wave forcing in CHEM than in NOCHEM partially because of the stronger mean state of the polar vortex. This then results, on average, in more intense SSWs, stronger stratosphere–troposphere coupling, a more negative NAO-like pattern at the surface, and long-lasting stratospheric ozone anomalies.

### 3.3 March sudden stratospheric warmings

We now turn to the March SSWs. Figure 8 shows the composite sea level pressure anomalies for CHEM and NOCHEM, as well as the CHEM-NOCHEM difference, for each of the first 2 months following the central date. Both simulations again show a negative NAO-like pattern in the...
2 months following the SSW. There are some regions with a significant difference between CHEM and NOCHEM in the first 30 d, but the pattern does not project strongly onto the NAO. Also, there is very little difference between the two composites in the second 30 d after the central date.

The surface responses seen following March SSWs in both models are weaker and less persistent than those following DJF SSWs, and the areas of strong or significant low or high anomalies are smaller. Three factors could contribute to this: weaker SSWs, weaker stratosphere–troposphere coupling, and a shorter NAM decorrelation timescale in March than in DJF (Baldwin et al., 2003; Simpson et al., 2011), resulting in weaker anomalies at the surface when averaged over several weeks. The differences between surface impacts of SSWs in CHEM and NOCHEM are also weaker for March SSWs. Thus, interactive ozone seems much less important for the surface effects of March SSWs than for DJF SSWs.

Considering the NAM in these simulations as shown in Fig. S2, we see negative NAM anomalies at the surface in both the CHEM and NOCHEM simulations, consistent with the negative NAO-like pattern seen in Fig. 8. There is a stronger signal in the troposphere in the CHEM compared to NOCHEM March SSW simulations at around 15–20 d after the central date, which may correspond to the surface pressure differences.

The NAM anomalies suggest that March SSWs in both CHEM and NOCHEM are weaker overall than the DJF SSWs; the stratospheric NAM anomalies are smaller and less significant. The eddy heat flux shown in Fig. S3, however, shows weaker wave forcing preceding only the CHEM (not the NOCHEM) March SSWs compared to those in DJF. Stratosphere–troposphere coupling also seems weaker compared to that seen for DJF SSWs. Further, the difference in the NAM descent between CHEM and NOCHEM is less
Figure 6. Eddy heat flux (in mK s$^{-1}$) over 40–80$^\circ$ N from $-60$ to $+30$ d around the DJF SSW central dates. The CHEM average is in blue with confidence intervals shown in pale blue. The NOCHEM average is in black with confidence intervals shown in gray.

Figure 7. Composite of total column polar cap (over 60–90$^\circ$ N) ozone anomalies in Dobson units in the CHEM simulations and composites of zonal mean zonal wind at 60$^\circ$ N and 10 hPa (in m s$^{-1}$) from $-60$ to 90 d around the central date of DJF SSWs in CHEM and NOCHEM. The black line shows the mean total ozone column; 1σ from the mean is shaded. The solid and dashed blue lines show the mean U1060 in CHEM and NOCHEM, respectively.

Strong and persistent than the difference seen after midwinter SSWs.

Soon after the central date for March SSWs, the NAM signal in the stratosphere is weaker with CHEM than NOCHEM in contrast to the midwinter SSW case. This difference appears to arise from the temperature and heating anomalies (Fig. S4). The lower stratosphere is only briefly and weakly warmer in CHEM compared to NOCHEM. Shortwave heating seems to be dominant in the temperature response to March SSWs, with the CHEM-NOCHEM difference in temperature anomalies (Fig. S4a) largely following the difference in shortwave heating anomalies (Fig. S4d). This is in contrast to the DJF SSWs, in which the shortwave heating had little effect and dynamical heating was dominant.

Finally, we note that unlike the DJF SSW case, the ozone anomaly for March SSWs does not persist after the event (Fig. 9). This is related to the seasonal breakdown of the vortex, as seen in the wind curves. Because these are late winter SSWs, the second month following the central date is near the expected stratospheric final warming date; the winds return to easterly about 50 d after the March SSW central date. The ozone anomaly returns to 0 Dobson units as the vortex breaks down. The maximum ozone anomaly is also about half the size of the maximum anomaly seen in DJF, which is consistent with the weaker nature of the March SSW events overall.

3.4 Midwinter strong polar vortex events

Finally, we turn our attention to strong polar vortex (SPV) events in DJF. While less extensively studied than SSWs, SPVs also impact surface climate. Baldwin and Dunkerton
(2001) suggest that strong polar vortex events can have surface signals comparable to but opposite in sign to those following SSWs, and Smith et al. (2018) found effects of Northern Hemisphere SPVs on spring and summer Arctic sea ice.

In the 30 d following the SPV central date, we see a pattern reminiscent of a weakly positive NAO in both CHEM and NOCHEM (Fig. 10). This positive NAO-like pattern appears stronger in CHEM than in NOCHEM but not significantly so. There is very little difference from climatology at the surface in the second month after the event in either of the simulations. This minimal difference using interactive versus specified ozone compared to the difference seen with SSWs may be related to the more zonal nature of SPVs. We specify ozone in a zonally symmetric way, which is much more consistent with the vortex seen in an SPV than in a SSW.

The NAM anomalies following SPVs in CHEM and NOCHEM (Fig. S5) have a similar strength (and opposite sign) in the stratosphere to those following midwinter SSWs, but they have much weaker downward propagation, which is consistent with an only weakly positive NAO. The difference between the NAM anomalies in CHEM and NOCHEM confirms a more positive NAM in the middle to lower troposphere in the first month following the SPV central date with interactive chemistry, but again, this difference is not significant and does not reach the surface.

These minimal differences in surface pressure and NAM are consistent with the similarity in the evolution of stratospheric temperature and heating rates in CHEM and NOCHEM, as shown in Fig. S6. The only large and significant difference is in stratospheric temperature 40–60 d following the SPV central date when the stratosphere is colder with interactive chemistry. This is after zonal mean zonal winds have returned to typical levels and is thus likely related to the stronger mean state of the stratospheric polar vortex with interactive chemistry compared to specified chemistry.

The zonal mean zonal winds in CHEM and NOCHEM around the SPV central dates further confirm that there is little difference in the strength of these events between CHEM and NOCHEM; the winds follow nearly identical trajectories from 30 d before to 30 d after the central date. We also see a weaker ozone anomaly following SPVs than following SSWs with a maximum absolute anomaly of about 30 Dobson units compared to 40 (Fig. 11). The ozone decrease following SPVs is also much more gradual than the increase seen in DJF SSWs. This is consistent with the fact that SPVs are not strong and sudden dynamical events in the way that SSWs are. As with DJF SSWs, though, the anomaly does persist for 3 months after the central date.

4 Conclusions

The climate model results presented here show an important relationship between interactive ozone, the climatological state of the stratospheric polar vortex, and the Euro-Atlantic surface impacts of midwinter SSWs. However, ozone chemistry has a minimal impact on the surface effects of March SSWs and of midwinter SPVs despite long-lasting total ozone column anomalies in the latter case. Furthermore, in contrast to the results reported by Haase and Matthes (2019), we do not find significantly fewer SSWs with interactive chemistry despite the stronger climatological polar vortex. However, we do find more frequent SPVs.

The stronger polar vortex mean state with interactive ozone chemistry also affects the surface signature of SSWs. A possible mechanism is that stronger wave forcing is necessary for an SSW to occur, and the resulting negative NAM propagates to the surface more strongly as well. This result is also consistent with that reported by Haase and Matthes (2019), although the effects documented here are weaker. In extending this work to consider March SSWs, we found that while the same stronger dynamical forcing is present, the influence of the shortwave heating term in late winter/early
spring results in a stratospheric temperature difference of the opposite sign, and there is little difference at the surface following March SSWs between interactive chemistry and specified chemistry simulations. We also find a minimal impact on midwinter surface effects of SPVs. However, we do see persisting negative ozone anomalies that can have an important effect in spring (Ivy et al., 2017).

Previous work (Smith and Polvani, 2014; Calvo et al., 2015; Ivy et al., 2017; Lin et al., 2017; Rieder et al., 2019) has shown the importance of ozone for the stratospheric polar vortex and surface springtime climate variability. Haase and Matthes (2019) further suggested that feedbacks among chemistry and dynamics are important for accurately capturing the response at the surface to SSWs, one of the major drivers of North Atlantic and European winter climate variability. By running longer simulations allowing for a cleaner quantification of the impact of interactive ozone, we find that these feedbacks are important for representing impacts of midwinter SSWs. However, we do not find similar importance for describing the surface response to March SSWs or DJF SPVs. Our results suggest that including interactive ozone chemistry may have a sizable impact on North Atlantic and European winter and spring climate variability in models.

Finally, we note that while we have only focused on winter SSWs and SPVs, stratospheric final warmings also have tropospheric effects (Black et al., 2006; Ayarzagüena and Serrano, 2009; Wei et al., 2007; Hardiman, 2011; Thieblemont et al., 2019; Butler et al., 2019). Those effects are dependent on the timing of the final warming, with earlier final warmings resulting in surface effects more like those seen following SSWs (Ayarzagüena and Serrano, 2009; Li et al., 2012). Interactive chemistry may thus also affect the representation and surface signature of stratospheric final warmings in models; this will be investigated in a follow-up study.
Appendix A

We calculate the NAM using a method similar to that of Gerber et al. (2010) and Gerber and Martineau (2018). The specific procedure is as follows.

1. We average model output to find a time series of daily zonal mean geopotential height $Z(t, \lambda, p)$ as a function of time $t$, latitude $\lambda$, and pressure $p$.

2. For every day and pressure level, we remove the global mean geopotential height $Z_{\text{global}}(t, p)$. This helps to remove the global changes so that the index instead mainly captures meridional differences or shifts (Gerber et al., 2010). (While not the case for the simulations used in this study, this step would remove much of the global warming signal if it were present.)

3. For each day, latitude, and pressure level, we remove the average for that calendar day over the whole period; that is, we remove the climatology to find an anomalous height.

4. For each day, latitude, and pressure, we remove the linear trend over the period.

5. For each day and pressure level, we compute a polar cap average. Here we are interested in the NAM, and we take the average from 65 to 90° N. This is a proxy for the annular mode as shown in Baldwin and Thompson (2009).

6. We multiply by $-1$ so that a positive polar cap geopotential height anomaly yields a negative NAM for consistency with the convention of Thompson and Wallace (1998).

7. We normalize the index by its standard deviation at each pressure level.
Data availability. All the model output is currently stored at the High Performance Storage System (HPSS) repository at the National Center for Atmospheric Research (NCAR). More specifically, the data can be found under the experiment tags “CO2x1SmidEmin_yBWCN” (CHEM) and “b.e10.B2000WSCCN.119_g16.control.001” (NOCHEM). Additionally, the data are available from the corresponding author upon request.

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Competing interests. The authors declare that they have no conflict of interest.

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