Reviewer #1:

We thank the reviewer for the assessment of our work and the help to connect it better to previous studies. Below we reply point by point, first showing the reviewers comments in italic and blue followed by our response. To avoid confusion, we refer to graphics shown in this document as Images and graphics shown in the paper manuscript as Figures.

In this paper, simplified model experiments are carried out to investigate the impact of ozone loss induced by energetic particle precipitation on atmospheric temperatures and dynamics from the mesosphere down to the surface. The topic is highly relevant at the moment, as energetic particle precipitation is recommended as part of the solar forcing for the upcoming CMIP-6 model experiments (Matthes et al., ACP, 2017). The results therefore are of great interest, and the paper is also very clearly structured and well written.

However, there are three points which need to be addressed before the paper can be published in ACP: a) the setup of the model experiments does not reflect the temporal and spatial structure of the direct and indirect particle impact as it is known from observations; b) some observation of the temperature response of the winter-time stratosphere to geomagnetic activity exist (e.g., Lu et al., JGR, 2008; Seppaelae et al., JGR, 2013) but are not used here to compare the results of this model run (actually the observed amplitude is much larger than the results shown here). This comparison needs to be included as it provides ground truth to estimate how realistic the modeled response of the troposphere is; c) the estimation of significance using a t-test is not applicable to the high-latitude Northern hemisphere winter, where due to the occurence of strong sudden stratospheric warmings the underlying distribution is bimodal.

These as well as a few more minor points are discussed in more detail below.

Page 1, lines 11 to page 2, line 8: the impact of energetic particle precipitation on the middle and lower atmosphere has been investigated since the 1970th, and a lot more has been published than referenced here. In particular there are two recent review papers which summarize the state of the art (Sinnhuber et al., Sur Geo, 2012; Mironova et al., Space Sci Rev, 2015), as well as reports on observations of a) the temporal and spatial structure of the indirect effect in different trace species (e.g., Hendrickx et al., JGR, 2015; Fytterer et al., JGR, 2015; Sinnhuber et al., JGR, 2016; Friederich et al., ACP, 2014); b) the temporal and spatial structure of the indirect effect in Noy (e.g., Funke et al., JGR, 2014a, b) and ozone (e.g., Fytterer et al., ACP, 2015; Damiani et al., GRL, 2016; Kazutoshi et al., ACP, 2017), c) the impact of the indirect effect on stratospheric temperatures and winds in the Northern hemisphere winter and spring (e.g., Lu et al., JGR, 2008; Seppaelae et al., JGR, 2013), and d) the response of tropospheric weather patterns to geomagnetic activity (e.g., Seppaelae et al., JGR, 2009; Maliniemi et al., JGR, 2014). Observations provide the ground truth your model study has to compare to, so should be summarized here. We followed the suggestion of the reviewer and included a summary of the observational record on the ozone loss due to EPP as well as the impact on the stratospheric temperature and zonal wind to the introduction. We included a large number of the suggested references.

Page 3, lines 22-25, description of model experiments with reduced ozone loss: the scenarios differ quite substantially from what is known about particle induced ozone loss from observations of the direct and indirect impact. They are very much simplified, and of course there is justification for carrying out very simple model studies. However, you should be aware how they differ from reality (as provided by observations), and discuss this carefully. Direct impact, mesospheric ozone: the direct impact has been shown to occur in sporadic events which are mostly short-lived (one day to a few days), but can occur in a periodicity related to solar rotation (27 days, 13.5 days, 18 or 9 days). It is restricted clearly to geomagnetic latitudes corresponding to the auroral oval (about 60-75 ◦

geomagnetic latitude). Implying this impact onto the whole polar cap should lead to an overestimation of this impact (see e.g., Hendrickx et al., JGR, 2015; Fytterer et al., JGR, 2015; Sinnhuber et al., JGR, 2016; Friederich et al., ACP, 2014). The indirect effect has been observed in every winter where observations in polar night have been available (Funke et al., 2014a,b). The impact of ozone is characterized by a downwelling negative anomaly starting in the upper stratosphere in mid-winter, and moving downwards to below 30 km in spring; it is restricted to the polar vortex (e.g., Fytterer et al., ACP, 2015; Damiani et al., GRL, 2016; Kazutoshi et al., ACP, 2017). Amplitudes are generally less than 20%, however it should be pointed out that observations show the difference of years with high to years with low geomagnetic activity; as the indirect effect occurs in every winter, see above, this is something different to the model experiments, which compare years with high activity to years with no activity, something that in reality doesn't happen even during deep solar minimum. We agree with the reviewer that our description of the experiments was too brief. We added two paragraphs to Section 2.1, also taking into account the comment of Reviewer #2. We still believe that our experimental design is justified, because this allows us a clear signal-to-noise ratio and long simulation periods in order to gain as much insights in the processes governing the climate impact of EPP. But we now added a discussion on how our experiments differ from the observational record. In particular, the lack of a vertical propagation of the signal, the shift of the polar vortex and the EPP restricted to the auroral oval are now discussed. We would also like to thank the reviewer for providing such an extended list of references, from which several are now cited in the manuscript.

Page 4, lines 5-9, determination of statistical significance: using a t-test implies a distribution of temperatures which is random around a mean state. However, in the Northern hemisphere polar winter, this is obviously not the case: years with sudden stratospheric warmings are not outliers of the mean atmospheric state distribution, they belong to a different distribution: the distribution of temperatures do not approach a normal distribution (as student's t-distribution), but is bimodal, with one mode for the years without, and one mode for the years with warmings. Therefore, you can only use the t-test separately for years with and without warmings (if the distribution of those years is indeed symmetric, which maybe you should check before doing a statistical test); it is definitely not applicable, and therefore meaningless, for the whole sample of winters with and without warmings. We analyzed the probability density functions (pdfs) for the variables shown in the paper. Image 1 shows exemplary the pdfs for winter polar mean (left) temperature averaged over (60 – 90N) and (right) zonal wind at 60N between 1 – 10 hPa. This corresponds to Figure 2 in the paper. Although the distribution is not smooth, it resembles more a normal distribution than a bi-modal distribution. We don't think that the notion of two distinct states, with and without an SSW is correct. There is a spectrum of major SSWs of very different peak intensities and durations which smoothly transitions into winter states without major SSWs which however often include minor SSW events of again different characteristics. This is also reflected by the fact that different SSW definitions may identify different cases. We think it is sufficiently justified to use Student's t-Test as it has been done in many earlier publications, e.g., the Student's t-Test is widely used for boreal polar winter, e.g., by Seppälä et al. (2009), Arsenovic et al. (2016) and Gray et al. (2012). We decided to stick to the Student's t-test.

Page 4, line 25-27: I eventually understood what you did there, but the sentence was difficult to follow. Maybe you can clarify it. Done.

Page 5, lines 12-13: there is one publication in ACPD at the moment which shows the same impact on heating rates (Sinnhuber et al., 2017) using a slightly different approach to yours. The results seem comparable, and I would encourage you to discuss/compare those results to yours. Thank you for pointing us to this paper. We added a comparison to this paper in Section 3.1.

Page 5, line 28: a change in the heating rate of 10% as for your stratospheric ozone experiment means a change of 0.1-0.2 K/day (see Figure 1). Observations and also the model study by Sinnhuber et al., ACPD, 2017, imply that this change in the stratosphere is not sporadic, but persists for several weeks, implying a warming during mid-winter of a few K. That is actually not a small change, and also in line with observations of the temperature response due to high geomagnetic activity in the high-latitude upper stratosphere (e.g., Lu et al., 2008; Seppaelae et al., 2013). We added the absolute values of the change in heating rates to the manuscript and compare it with the model study of Sinnhuber et al. (2017).

Page 5-8, discussion of statistical significance: a t-test is just not applicable if you combine years with and years without SSWs, see my comment above. I think you should study the change in years with and without warmings separatedly; then you can provide a robust measure of the significance. Also, this would make the results more comparable to the observations shown in Seppaelae et al., 2013, for the stratospheric response, as they also analyze years without warmings. We followed the suggestion of the reviewer and redid Figures 2-4 from the paper separately for SSW and no-SSW (see Images 2-4 only for no-SSW). For each SSW event the according season was marked as "with SSW". If a SSW occurs in February or November, also the next season was marked as "with SSW" (i.e., for February MAM and for November DJF). This method ensures the consistency of each season and prevents an influence of early or late SSWs on the next season. In total, we obtained 74 (71) winters without SSW for piControl (strato-O3). Comparing Images 2-4 to Figures 2-4 of the paper, we see that they are very similar and our conclusions still hold. For strato-O3 the signal, especially in the late winter, even weakens. We understand that the inclusion of SSW winters in the context of EPP forcing is somewhat problematic as such events, depending on their time of occurrence, may in reality or coupled chemistry models (not in our idealized setting) influence the forcing (i.e., polar ozone depletion) itself. On the other hand, discarding SSW winters might actually remove a big part of the signal, as a forcing may also change the timing of SSW occurrence (Gray et al., 2013). The QBO dependence of solar UV effects on the polar winter stratosphere, as shown e.g., by Labitzke et al. (2006), is e.g., strongly dependent on SSW occurrence. Additionally, see above, we don't think that the notion of a bimodal distribution of winter states is correct. Therefore, we strongly prefer to keep the figures showing all years (SSW+no-SSW). But we added information on the changes in temperature and zonal wind if only no-SSW seasons are considered.

Page 8, line 4: the impact in the winter-time high latitude upper stratosphere temperatures you show in Figure 2 has a similar structure to observed temperature and wind field changes for years with high geomagnetic activity (Lu et al., 2008; Seppaelae et al., 2013). However, the amplitude of the warming is much smaller (about one order of magnitude?) than in the observations. This comparison to observations needs to be discussed here. It is true that our results match qualitatively very well the results (for temperature) obtained from reanalysis data (Lu et al, 2008, Seppälä et al., 2013). Whereas for the zonal wind response, the two studies differ from each other. Seppälä et al. (2013) showed a strengthening of the polar vortex with enhanced equatorward planetary waves, whereas Lu et al. (2008) showed a weakening of the polar vortex. We added this information to the manuscript.

Page 8, line 8: the interhemispheric coupling is evident in both the meso-O3 and the strato-O3 experiments as a "statistically significant" change in the summertime upper mesosphere. However, this is more likely an affect of SSWs? We checked this for only winter without a SSW and still found a warming in the summer upper mesosphere (see Image 2). This suggests that the signal is not an effect of SSWs. While we find this strong change very interesting, we think that further analysis is beyond the scope of this paper.

Page 8, line 25-30: The patterns and amplitudes you observe here should be compared to observations (Seppaelae et al., 2009; Maliniemi et al., 2014). However, as the amplitudes of your stratospheric warming appears to be much lower than observed, I would expect the impact on the troposphere also to be low compared to observations. Another point: Seppaelae et al., 2009 show that the impact on surface temperatures is different, with larger amplitudes, when years with SSWs are not considered. You should separate years with and years without warmings here as well. Can you reproduce their result regarding the impact of warmings? Again, a t-test is not applicable if you use years with and without warmings. We added a comparison of the surface temperature response to observations. In addition to the analysis of the full sample we analyzed the impact restricted to winters without SSW (see Image 4). A more detailed description on how this subset is calculated is given in the comment "Page 5-8". We obtained larger amplitudes in the surface temperature for winters without SSW. However, still much smaller than in Seppälä et al. (2009) and Baumgaertner et al. (2011). The cooling over Northern America agrees qualitatively with the aforementioned studies, but we obtained no warming over Eurasia. We added the behavior for winters without SSW to the manuscript.

Page 11, 11: "Our results suggest that the climate impact of an ozone loss due to EPP is small" considering that the impact of particle precipitation in your analysis is masked by the strong variability implied on the Northern hemisphere winter atmosphere by sudden stratospheric warmings, and your results of the stratospheric impact strongly underestimate the observed response of the stratosphere, you can not draw this conclusion at this point. We rewrote the whole paragraph and encourage now more research to clarify the effects of EPP. We think it is important to point out that the climate impact of EPP is not as clear as often thought.

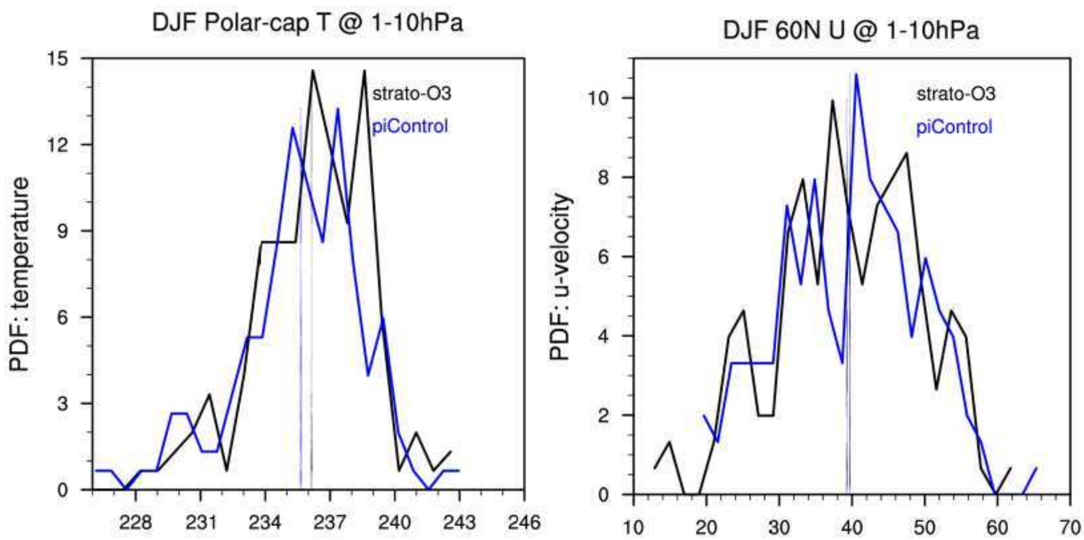

Image 1: Probability density functions for winter polar mean (left) temperature averaged over 60 – 90N and (right) zonal wind at 60N between 1 and 10 hPa. Two experiments are depicted: (black) strato-O3 and (blue) piControl.

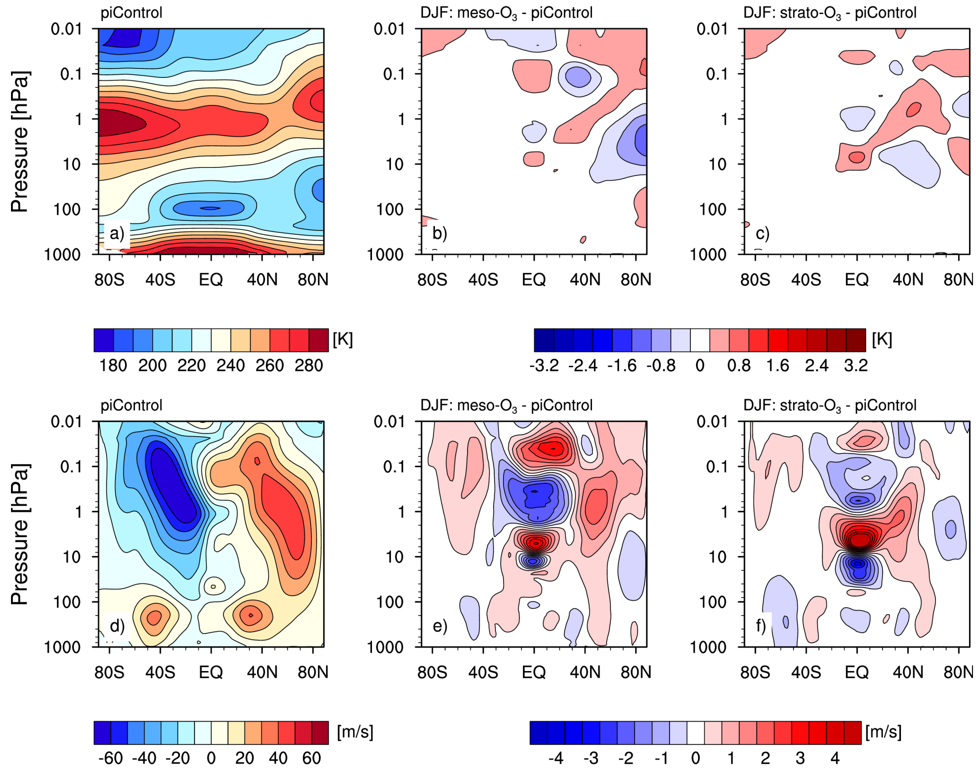

Image 2: Same as Figure 2 (in paper) but only for winters without SSW.

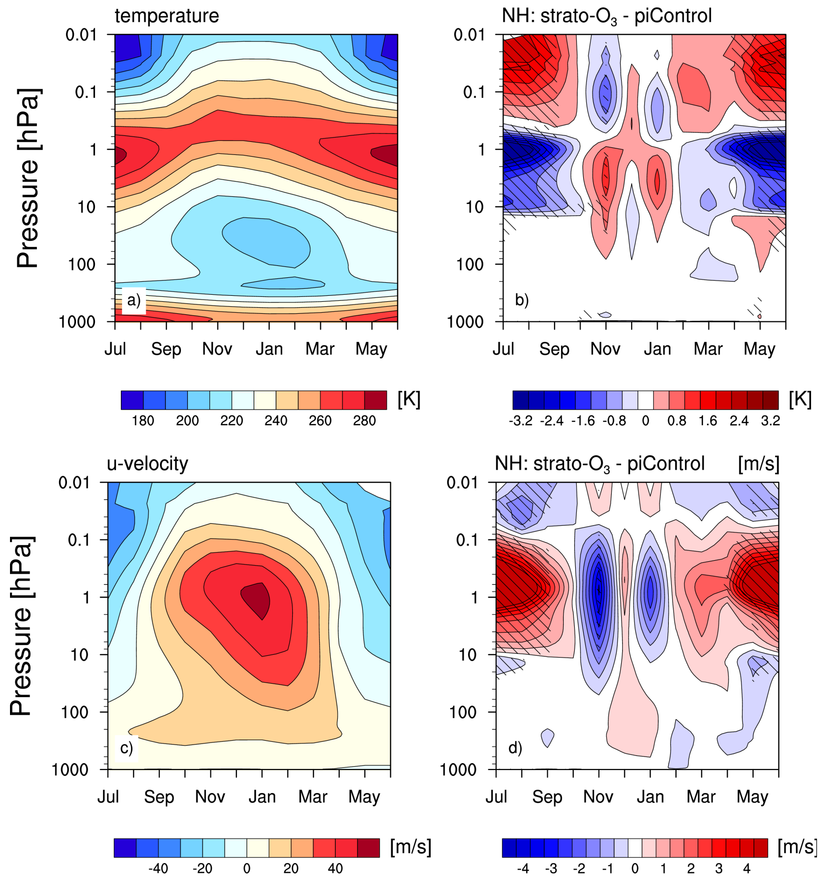

Image 3: Same as Figure 3 (in paper) but only for seasons without SSW.

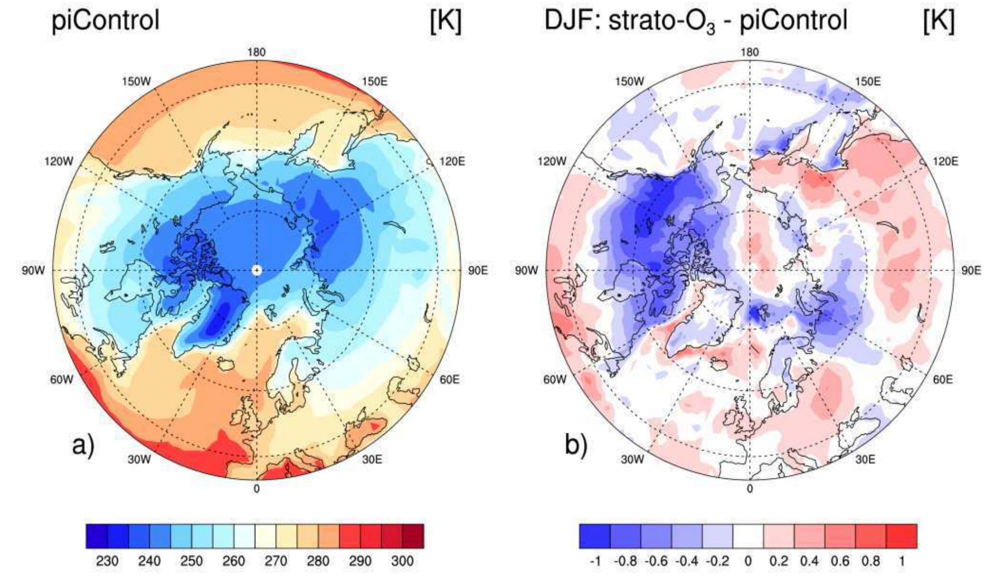

Image 4: Same as Figure 4 (in paper) but only for winters without SSW.

Reviewer #2:

We thank the reviewer for the assessment of our work and the useful suggestions for improvements. Below we respond point by point, first showing the reviewer's comments in blue and italic followed by our response. To avoid confusion, we refer to graphics shown in this document as Images and graphics shown in the paper manuscript as Figures.

The manuscript presents the response of the atmosphere and surface temperature to the introduced permanent decrease of the ozone concentration in the mesosphere and upper stratosphere simulated with the MPI-ESM model. The forcing was designed to mimic the ozone depletion by hydrogen and nitrogen oxides formed by the precipitating energetic particles. The subject of the manuscript is appropriate for ACP because it addresses widely discussed during the last decade question about possible influence of the energetic particles on the atmosphere, ozone and surface air temperature. The manuscript is well written, the most of relevant publications are cited, the figures are clear. However, the manuscript does not look mature because the bold conclusions cannot really be supported by the presented results. It seems obvious for the authors because in the summary they formulate why the results are not convincing and what to do to make them better. Therefore, I cannot recommend publication in the present form.

Main issues:

1. The experimental design is too simplified. It resembles the ozone loss due to EPP obtained from the observations and models however substantially differs in the time evolution and distribution in space. Application of realistic ozone depletion scenarios could lead to very different results. If the authors do not know the implications of the chosen scenario (as it is said in the summary) what potential readers could learn from the paper? There are several aspects of the problem such as shift of the vortex from the pole and intensified ozone influence on solar radiation heating or interaction of the propagating disturbance with internal variability modes like PJO. These effects are automatically taken into account in the models considered all relevant to EPP processes, but they are missed if too simplified approach is applied. The simplest way to avoid the problem is to eliminated connection with EPP. Actually, the introduced ozone depletion scenario in the upper stratosphere is closer to the influence of halogens. We agree with the reviewer that our description of the experiments was too brief. However, its simplistic nature is intended and, we think, useful. We added two paragraphs to Section 2.1, also taking into account the comment of Reviewer #1. Earlier studies (see introduction for references) consider a mix of stratospheric and mesospheric ozone losses. The sole impact of a mesospheric ozone loss due to the direct EPP effect as suggested by Andersson et al. (2014) remains unclear. Additionally, a stratospheric warming due to EPP was identified in reanalysis data (Lu et al. 2008; Seppälä et al. 2013), whereas model studies obtained a stratospheric cooling either of dynamical origin (Baumgaertner et al. 2011) or of radiative origin (Arsenovic et al. 2016). In this sense, we believe that our experimental design is justified, because a) we can separate the climate impact of stratospheric and mesospheric ozone loss due to EPP; and b) the simplified approach allows us to gain insights in the processes governing the climate impact of EPP. Prescribing complex ozone reductions that vary in space, interseasonally and interannually, or simulating the ozone reduction interactively, might enable more realism but doesn't facilitate the identification of potential mechanisms. We think a reader can learn from our study that a) a significant climate impact of a mesospheric ozone change as suggested by Andersson et al. (2014) seems unlikely; and b) the interplay of dynamical cooling and radiative warming is complex and the climate impact of stratospheric ozone losses due to EPP is not as clear as often thought. In our simulations, we obtained a radiative warming in November and January. But in December, when the polar night is shortest and, hence, the radiative warming is strongest, a dynamical cooling is found. Therefore, additional research is needed to clarify the role of wave reflection for the dynamical feedback and for the coupling mechanism between stratosphere and troposphere. Furthermore, we now added a discussion on how our experiments differ from the observational record. In particular, the lack of downward propagation of the signal, the shift of the polar vortex and the EPP restricted to the auroral oval are now discussed.

2. I found interesting a large disagreement between the results of 80 and 150-year long runs. I guess, this phenomenon should be understood and explained with more details. I am not convinced that it is just the results of inter-annual variability. If so all modeling community is in a huge trouble. Did the authors check the presence of any model drift? We agree with the reviewer that this large disagreement is interesting. Following your recommendation, we show different quantities that one might assume to influence EPP signals if they were drifting (Image 1). We do not find any drift in the model. The maximum difference (highest value – lowest value) in the sea surface temperature is 0.2 K for piControl and 0.17 K for strato-O3. This agrees with the internal variability in global mean surface temperature estimated by Sutton et al. (2015) for CMIP5 pre-industrial control experiments. We added a sentence to the manuscript and stated that no model drift is found.

3. The authors frequently discuss not statistically significant responses. I have noticed that almost all results presented in Figure 2 and 4 are not significant. It is rather interesting why the applied model is not sensitive to 20% decrease of the ozone in the polar upper stratosphere. There were several publications (mentioned in the introduction) claiming significant response of the atmosphere to the observed ozone depletion in the last decades of 20th century and the ozone depletion scenario is close to what is used in the manuscript. Some discussion of this issue is necessary. It is true that most signals in Figures 2 and 4 are not significant at the 95% level. Nevertheless, it makes sense to analyze if the signals could have a physical explanation and not be purely accidental. Additionally, we want to emphasize that even if DJF averages are not significant, this can be different for individual months, as we show in Figure 3. Graf et al. (2007) and Langematz et al. (2003) used observed ozone changes to analyze the role of ozone for climate change. In both studies, the ozone is mostly reduced in the lower stratosphere, in contrary to the upper stratosphere in our study. Additionally, they used a rather short simulation period (10 years in Graf et al. (1997) and 20 years in Langematz et al. (2003)). Analyzing different simulation periods we obtain mesospheric warming and cooling of apparent significance. However, also compared to observational records of temperature and zonal wind responses due to EPP (Lu et al. (2008) and Seppälä et al. (2013)), the amplitude of our responses are smaller. We now added a comparison to the above mentioned studies.

4. Section 3.1: The use of 75N should be better motivated if the authors would like to wire these results with ozone depletion due to EPP. If the ozone depletion occurs inside polar vortex then 75N is not representative because huge ozone influence on solar heating rate outside polar night area will dominate over very small longwave effect. It should be also considered that in the Northern hemisphere the vortex is not stable and tends to move from the pole out of the polar night area. Figure 1 is only an illustrative example of polar ozone heating rates and it is not thought to be representative. At other latitudes the polar night would be, of course, shorter or longer. Additionally, we agree with the reviewer that the length of the polar night exposure of an air parcel depends on altitude and the actual dynamics (e.g., movement of the air parcel). The pure radiative response to ozone loss should be a warming in mid-winter and get weaker towards early and late winter. However, our Figure 3 shows a warming in November and January/February, but not in December. The December cooling is of dynamical origin. We now discuss the missing shift of the polar vortex in Section 2.1 and added the above mentioned information to Section 2.2.

Minor issues:

1. Page 2, line 2: if –> of. Done.

2. Page 2, line 4: Langematz et al. (2003) showed tiny direct LW warming (Fig.7) , but the resulting stratosphere is cooler (Fig.8). Graf et al., (1998) showed the response in the lower stratosphere (70 hPa). Thank you for pointing this out. We changed the sentence to: "During polar night reduced ozone slightly decreases the infrared cooling of the polar stratosphere resulting in a net (small) stratospheric warming (Graf et al., 1998; Langematz et al., 2003). However, both studies prescribed an ozone loss in the lower stratosphere."

3. Page 3, line 23-25, line 31: The ozone depletion scenario is too simplified. We extended the description of the applied ozone losses and discuss now differences to observed changes. See also reply to major comment 1.

4. Section 2.2: The radiation code is not described. The references do not provide satisfactory information about the treatment of solar (e.g., spectral range coverage, spherical) and infrared (e.g., LTE treatment) radiation. The standard version of the RRTMG does not include wavelengths shorter 200 nm and therefore the heating rate in the mesosphere should be heavily underestimated due to the absence of Lyman-alpha line and Schumann-Runge bands. How it is treated in Psrad? The solar and infrared radiation is treated in Psrad in the same way as in RRTMG. Hence, wavelengths shorter than 200 nm are not included. However, the absorption of ozone takes primarily place in three spectral regions: Hartley band (200 – 310 nm), Huggins band (310 – 350 nm) and Chappius band (410 – 750 nm) (Brasseur and Solomon, 2005). All of those bands are considered in RRTMG and, hence, also in Psrad. The Schumann-Runge bands are of great importance for the mesosphere, but primarily due to absorption of molecular oxygen (and not ozone). In this sense, we underestimate the total heating rate in the mesosphere. However, in our setup we compare two radiative transfer calculations. The difference between both calculations is not (at least not strongly) affected by the underestimated total heating rate. We extended the description of Psrad in the manuscript.

5. Page 4, lines 16-18: I do not understand what means "separately . . .and then combined". Why CO2 is not in the input list. Is it not included in Psrad? We rewrote this sentence to make it clear that the optical properties are calculated for shortwave and longwave separately, but then combined to estimate the total heating rate. CO2 and O2 are set to fixed values invariant with height in Psrad. We added this information.

6. Page 4, line 24: Actually, the length of the polar night depends on the altitude and at 80 km it could well be shifted by one month relative to the surface. In Figure 1 this effect is absent, which affects the results in the mesosphere. Thank you for pointing this out. Please see also comment to major point 4. We added a discussion on the representativeness of 75N to the manuscript.

7. Page 5, line 4: The maximum of the ozone VMR is normally around 6 hPa for this location. What ozone profiles were used? We used ozone profiles averaged over the late 20th century provided by the general circulation and chemistry model HAMMONIA. In this profile the maximum ozone VMR is also around 6 hPa. The strongest heating occurs around the stratopause, which agrees e.g., with Brasseur and Solomon, 2005). We adjusted the sentence accordingly.

8. Page 5, line 33: I guess, Langematz et al. (2003) showed the same. We added Langematz et al. (2003) to the references.

9. Page 6, line 9: 75N is not really representative (see above). Please see the comment to major point 4.

10. Page 8, line 5: 75N is not really representative (see above). This result disagrees with Langematz et al. (2003, see their Figure 7 and 8). Langematz et al. (2003) showed a dynamical cooling in the polar winter stratosphere but expected also a warming from the radiative transfer modeling. In contrast, Lu et al. (2008) and Sepppälä at al. (2013) showed a warming in the polar winter upper stratosphere due to EPP in re-analysis data, but the magnitude is much stronger (~5 K) than in our simulations. Furthermore, we found a small dynamical cooling in December, which is caused – as in Langematz et al. (2003) – by a reduction of waves entering the stratosphere. In this context, we discuss the differences to Langematz et al. (2003). We added the comparison to Lu et al. (2008) and Seppälä et al. (2013) to the manuscript.

11. Page 8, line 15: statein –> state in. Done.

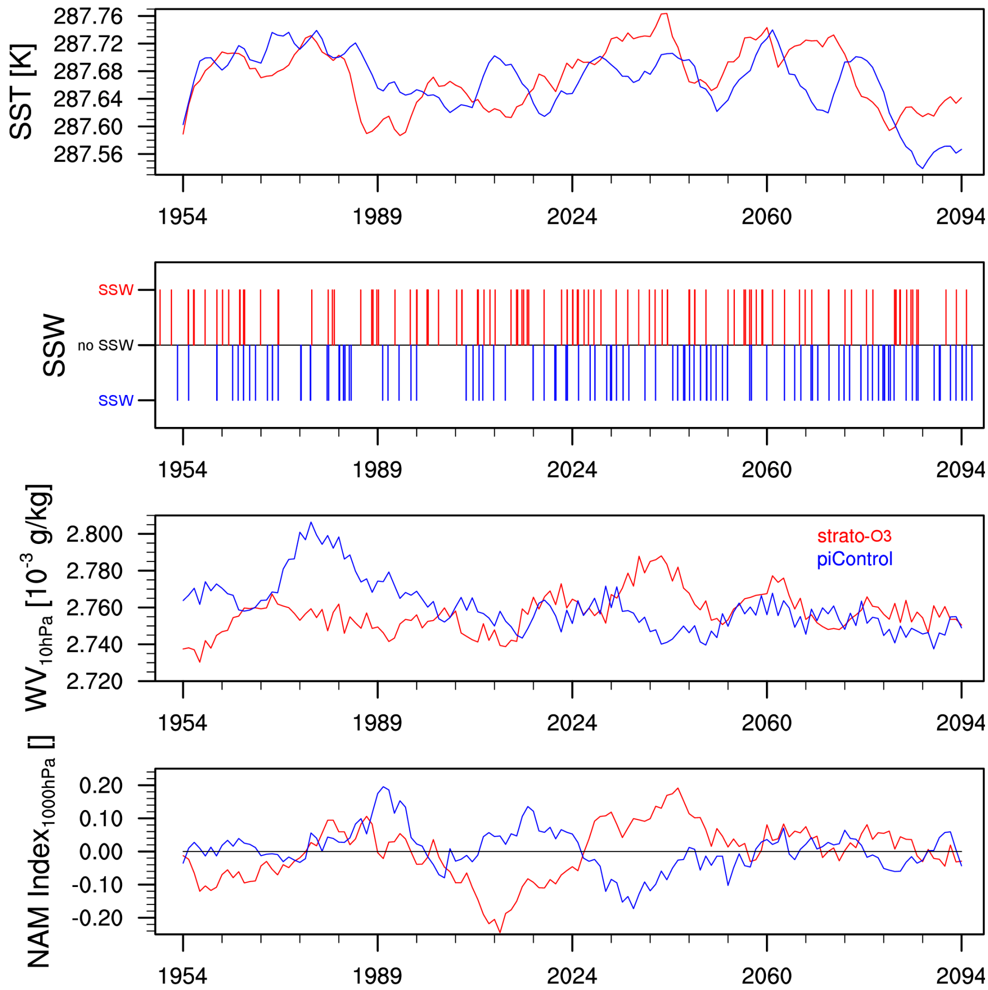

Image 1: Temporal evolution of different quantities for piControl (blue) and strato-O3 (red). (from top to bottom) Global and 10-year running mean of sea surface temperature; occurrence of sudden stratospheric warming events; global and 10-year running mean of water vapour at 10 hPa and 10-year running mean of Northern Annual Mode index at 1000 hPa.

<u>Additional references:</u>

Sutton, Rowan, Emma Suckling, and Ed Hawkins. “What Does Global Mean Temperature Tell Us about Local Climate?” Phil. Trans. R. Soc. A 373, no. 2054 (2015): 20140426. doi:10.1098/rsta.2014.0426.

Brasseur, Guy P., and Susan Solomon. Aeronomy of the Middle Atmosphere: Chemistry and Physics of the Stratosphere and Mesosphere. Springer Science & Business Media, 2005.

Lu, Hua, Mark A. Clilverd, Annika Seppälä, und Lon L. Hood. „Geomagnetic Perturbations on Stratospheric Circulation in Late Winter and Spring“. Journal of Geophysical Research: Atmospheres 113, Nr. D16 (2008): D16106. doi:10.1029/2007JD008915.

Seppälä, A., H. Lu, M. A. Clilverd, und C. J. Rodger. „Geomagnetic Activity Signatures in Wintertime Stratosphere Wind, Temperature, and Wave Response“. Journal of Geophysical Research: Atmospheres 118, Nr. 5 (2013): 2169–83. doi:10.1002/jgrd.50236.

List of relevant changes to the manuscript

- Added references to satellite observations and reanalysis data to the introduction
- Extended the description of the experiments (e.g., comparison with satellite observations, what is missing)
- Added treatment of solar and infrared radiation in the radiative transfer model
- Compared the heating rates to the recent paper of Sinnhuber et al. (2017)
- Added information on behavior without SSWs to the results
- Compared our results with earlier studies using re-analysis data or model data
- Rewrote the limits and perspectives of our study

Climate Impact of Polar Mesospheric and Stratospheric Ozone Losses due to Energetic Particle Precipitation

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Abstract. Energetic particles enter the polar atmosphere and enhance the production of nitrogen oxides and hydrogen oxides in the winter stratosphere and mesosphere. Both components are powerful ozone destroyers. Recently, it has been inferred from observations that the direct effect of energetic particle precipitation (EPP) causes significant long-term mesospheric ozone variability. Satellites observe a decrease in mesospheric ozone by up to 34 % between EPP maximum and EPP minimum. Here, we analyze the climate impact of polar mesospheric and polar stratospheric ozone losses due to EPP in the coupled climate model MPI-ESM. Using radiative transfer modeling, we find that the radiative forcing of a mesospheric ozone loss during polar night is small. Hence, climate effects of a mesospheric ozone loss due to energetic particles seem unlikely. A stratospheric ozone loss due to energetic particles warms the winter polar stratosphere and subsequently weakens the polar vortex. However, those changes are small, and few statistically significant changes in surface climate are found.

1 Introduction

Energetic particles enter the Earth's atmosphere near the magnetic poles altering the chemistry of the middle and upper atmosphere. Energetic particle precipitation (EPP) is the major source of nitrogen oxides (NO_x) and hydrogen oxides (HO_x) in the polar middle and upper atmosphere (Crutzen et al., 1975; Solomon et al., 1981). Both chemical components catalytically deplete ozone; NO_x mainly below and HO_x mainly above 45 km.

HO_x is short-lived in the middle atmosphere and depletes mainly ~~the~~ ozone in the mesosphere. In contrast, NO_x persists up to several months in the polar winter middle atmosphere. Inside the polar vortex, NO_x can be transported downward from the lower thermosphere to the stratosphere, where it depletes ozone ~~(Funke et al., 2007; Sinnhuber et al., 2014) .~~ (e.g., Funke et al., 2017; Sinnhuber et al., 2014; Hendrickx et al., 2015) . Observational evidence of polar winter stratospheric ozone loss due to EPP is still limited. Only recently, long-term satellite observations with good temporal and spatial coverage became available. In austral polar winter EPP causes an ozone loss of about 10 – 15 % descending from 1 hPa in early winter to 10 hPa in late winter (Fytterer et al., 2015; Damiani et al., 2016) . Extensive information on the current knowledge of energetic particle precipitation can be found in Sinnhuber et al. (2012) and Mironova et al. (2015) .

Ozone loss influences stratospheric temperature and the polar vortex. The Northern Annual Mode (NAM) index is often used to describe the strength of the polar vortex, with positive NAM values indicating a strong polar vortex and negative NAM values indicating a weak polar vortex. Observations indicate that anomalous weather regimes associated with the NAM index can propagate from the stratosphere down to the surface (Baldwin and Dunkerton, 2001). Hence, energetic particle precipitation may provide a link from space weather to surface climate. Here, we study the impact of an ozone loss due to EPP on the circulation and subsequently on climate. Discussed are both a polar mesospheric and a polar stratospheric ozone loss.

Since the discovery of the ozone hole in the mid-1980s, the climate impact of a stratospheric ozone loss has been intensively studied (e.g., Shine, 1986; Randel and Wu, 1999; Lubis et al., 2016). Most studies concentrated on the climate impact of the ozone hole during austral spring and reported a cooling in the spring Southern Hemispheric stratosphere due to reduced absorption ~~if~~ of solar radiation and a strengthening of the polar vortex. In contrast, our study concentrates on an ozone loss during the boreal polar night. During polar night reduced ozone slightly decreases the infrared cooling of the polar stratosphere resulting in a net (small) stratospheric warming (Graf et al., 1998; Langematz et al., 2003). However, both studies prescribed an ozone loss in the lower stratosphere.

Several studies suggested a significant influence of EPP on ~~the surface air temperature during winter.~~ climate. Seppälä et al. (2013) and Lu et al. (2008) used reanalysis data to investigate the dependence of stratospheric temperature and zonal wind to the AP-Index. They found a stratospheric warming up to 5 – 10 K for strong energetic particle precipitation descending from the stratopause to the mid-stratosphere. However, for the zonal wind response the two studies differ from each other. Seppälä et al. (2013) found a strengthening of the polar vortex, wheras Lu et al. (2008) showed a weakening of the polar vortex. Moreover, Seppälä et al. (2009) analyzed surface air temperature changes in reanalysis data for years with various strengths of EPP. They found a warming over Eurasia and a cooling over Greenland for winters with enhanced EPP, but could not rule out that the estimated changes are induced by NAM variability independent of EPP.

Other studies relied on atmospheric chemistry models, which showed similar surface temperature change patterns as found in the reanalysis data (e.g., Rozanov et al., 2005; Baumgaertner et al., 2011; Arsenovic et al., 2016). They reported a small cooling in the polar winter stratosphere due to EPP~~, although~~ . However, the radiative effect of a polar night ozone loss should ~~be a warming. This~~ lead to a warming, which can also be found in reanalysis data (Lu et al., 2008; Seppälä et al., 2013) . The simulated stratospheric cooling is attributed to a dynamical, adiabatic cooling caused by a decrease in the mean meridional circulation (Schoeberl and Strobel, 1978; Christiansen et al., 1997). Langematz et al. (2003) suggested that the weaker mean meridional circulation is caused by a decrease in midlatitude tropospheric wave forcing. The aforementioned model studies analyzing the climate impact of EPP relied on relatively few simulation years and applied complex forcings. Instead of prescribing ozone, these studies simulated EPP effects by changing the production of NO_x and HO_x and modeling the effects on ozone interactively. This could potentially be more realistic than simulations with prescribed ozone anomalies but introduces uncertainties related to the representation of chemistry and transport in the model, and renders the understanding of the effects more complicated as the ozone forcing varies in space and time. To avoid these difficulties and to obtain a clear signal-to-noise ratio, we use an idealized ozone forcing and a long simulation period.

Commonly, the effects of EPP are classified into direct and indirect effects (Randall et al., 2006, 2007). Direct effects are the effects of the local production of NO_x and HO_x, whereas indirect effects are the effects of the NO_x transport from the thermosphere to the stratosphere. Whereas most of the above mentioned studies discuss a mainly stratospheric ozone loss due to the indirect EPP effect, Andersson et al. (2014) suggested a potential climate influence of a mesospheric ozone loss due to the direct EPP effect. By using satellite observations they showed that HO_x causes long-term variability in mesospheric ozone up to 34 % between EPP maximum and EPP minimum. Arsenovic et al. (2016) were the first to include the direct effect of HO_x local production due to EPP in a chemistry-climate model. They found a similar mesospheric ozone loss as Andersson et al. (2014) and ultimately, reported a cooling over Greenland and a warming over Eurasia. However, Arsenovic et al. (2016) also considered the indirect effect of the NO_x descent. Hence, the sole impact of a mesospheric ozone loss due to the direct EPP effect as suggested by Andersson et al. (2014) remains unclear.

This paper studies the circulation and climate impact of idealized mesospheric and stratospheric ozone losses that could be attributed to energetic particle precipitation. We use simulations with the Max Planck Institute Earth System Model (MPI-ESM) applying an idealized ozone forcing in either the mesosphere or the stratosphere. The idealized mesospheric ozone loss that we prescribe may be considered to be mostly a direct EPP effect, whereas the prescribed stratospheric ozone loss should be considered indirect. ~~We use simulations with the Max Planck Institute Earth System Model (MPI-ESM) applying an idealized ozone forcing in either the mesosphere or the stratosphere.~~ Additionally, we use a radiative transfer model to quantify the radiative forcing of ozone at different altitudes and months. Ultimately, we discuss whether an ozone loss in the middle atmosphere due to EPP has the potential to significantly alter the surface climate. Section 2 describes the MPI-ESM as well as the radiative transfer model. Section 3 links mesospheric and stratospheric ozone losses to changes in the atmospheric temperatures and winds. Finally, Section 4 summarizes and discusses the main outcomes and limitations of this study.

2 Models and numerical experiments

2.1 MPI-ESM: The Max Planck Institute Earth System Model

The Max Planck Institute Earth System Model (MPI-ESM; Giorgetta et al. (2013)) consists of the coupled atmospheric and ocean general circulation models, ECHAM6 (Stevens et al., 2013) and MPIOM (Jungclaus et al., 2013) as well as of the land and vegetation model JSBACH (Reick et al., 2013) and of the model for marine bio-geochemistry HAMOCC (Ilyina et al., 2013). We use the 'mixed-resolution' configuration of the model (MPI-ESM-MR). The ocean model uses a tripolar quasi-isotropic grid with a nominal resolution of 0.4° and 40 vertical layers. ECHAM6 is run with a triangular truncation at wave number 63 (T63), which corresponds to 1.9° in latitude and longitude. The vertical grid contains 95 hybrid sigma-pressure levels resolving the atmosphere from the surface up to 0.01 hPa. The vertical resolution is nearly constant (700 m) from the upper troposphere to the middle stratosphere and less than 1000 m at the stratopause. The time steps in the atmosphere and ocean are 450 and 3600 s, respectively.

The model has been used for many simulations within the CMIP5 (Coupled Model Intercomparison Project Phase 5) framework (Taylor et al., 2012). An overview of the dynamics of the middle atmosphere in these simulations is given by Schmidt et al. (2013). In this study, the preindustrial CMIP5 simulation (piControl) is used as reference. The forcing is constant in time and uses pre-industrial conditions (1850 AD) for the greenhouse gases. Solar irradiance and ozone concentrations are averaged over a solar cycle (1844 – 1856 for the solar irradiance and 1850 – 1860 for ozone concentrations). No volcanic forcing is applied. A period of 150 years of this simulation is used.

In order to analyze the impact of ozone changes on the model climate, two additional experiments with reduced ozone concentration are carried out. In one experiment, the mesospheric ozone is reduced by 40 % between 0.01 hPa and 0.1 hPa polewards of 60° N (this is called "meso-O_3"). In the other experiment, stratospheric ozone is reduced by 20 % between 1 hPa and 10 hPa polewards of 60° N (this is called "strato-O_3"). ~~The~~ We perform on-off experiments, whereas in reality EPP causes a constant (but variable) ozone loss. However, the magnitude of the prescribed ozone losses is based on ~~the results from previous studies for boreal winter . Andersson et al. (2014)~~ satellite observations for winter conditions between years with high geomagnetic activity and years with low geomagnetic activity. In general, the impact of energetic particles is sporadic in the mesosphere, Andersson et al. (2014) , however, showed that the direct HO_x effect induces a long-term variability in mesospheric ozone up to 34 % from November to February in satellite data. ~~Baumgaertner et al. (2011) simulated polar winter losses of up to 20~~Fytterer et al. (2015) and Damiani et al. (2016) revealed an upper stratospheric ozone loss between 10 – 15 % ~~in the upper stratosphere due to EPP in a model.~~ due to energetic particles for the Antarctic high latitudes for 1979 – 2014. Note that the applied ozone losses are slightly larger than the EPP influence diagnosed from observations. We use the stronger forcing to obtain a clear ~~signal from which to diagnose the sensitivity of the climate to an ozone loss due to EPP~~signal-to-noise ratio. However, this implies a potentially overestimated climate response.

To facilitate the experiment design, we applied the ozone losses constant over time. Although we concentrate our analysis on boreal winter high latitudes, this allows us to gain insights on boreal spring (i.e., the transition time from polar night to polar day). Observed ozone losses in summer are in general smaller than during winter, but this idealized setting allows an easy comparison of potential effects during the different seasons. It is unlikely that signals in summer affect the climate of the next winter.

Both experiments are forced by the same conditions as the piControl experiment. Moreover, the simulations are restarted from the same year in the piControl experiment. This ensures that the ocean state is similar in all experiments. For both simulations 150 years are simulated.

The simplistic nature of our experiments is intended and, we think, useful. We chose this idealized experimental design in order to separate the climate impact of stratospheric and mesospheric ozone loss due to EPP and to gain insights in the processes how EPP affects the climate. Prescribing complex ozone reductions that vary in space, interseasonally and interannually, or simulating the ozone reduction interactively, might enable more realism but do not facilitate the identification of potential mechanisms. However, due to the simplification we cannot consider all features associated with EPP. In particular, three main effects are not taken into account: a) energetic particles enter the atmosphere only over the auroral oval regions (Hendrickx et al., 2015; Fytterer et al., 2015) ; b) the negative ozone signal due to EPP propagates from the stratopause in mid-winter to the lower stratosphere in spring within the polar vortex (Funke et al., 2017; Damiani et al., 2016) ; and c) the polar vortex can be shifted off the pole to regions with more solar radiation. We, instead, apply a constant ozone reduction between the stratopause and mid-stratosphere (1 – 10 hPa) over the whole polar cap. The climate response in our simulations is likely overestimated as we reduce ozone over a larger latitudinal and altitude region than observations suggest.

In the Sections 3.2 and 3.3 the differences between the experiments and the control simulation (i.e., piControl) are analyzed. Statistical significance is calculated using the 95% confidence intervals assuming normally-distributed regression errors and using the 0.975 percentile of Student's t-distribution with the appropriate degrees of freedom. Properties of two simulations are considered statistically significantly different if the mean value of the control simulation is outside 95% confidence interval of the experiment.

2.2 The radiative transfer model PSrad

The radiative transfer scheme of MPI-ESM is based on the rapid radiation transfer suite of models optimized for general circulation models (RRTMG; (Mlawer et al., 1997; Iacono et al., 2008)). The RRTMG is widely used and its ability to calculate radiative forcing has been evaluated by Iacono et al. (2008). In its stand-alone version, it is used here to study the impact of ozone on heating rates. It is divided into sixteen bands in the longwave (1000 – 3 µm) and fourteen bands in the shortwave (12195 – 200 nm) (Clough et al., 2005) . The spectral bands are chosen to include the major absorption bands of active gases. The major ozone absorption bands – Hartley band (200-310 nm), Huggins bands (310-350 nm), and Chappuis bands (410-750 nm) – are considered. However, absorption of oxygen at shorter wavelengths than 200 nm is missing, which could lead to an underestimation of the total heating rate in the mesosphere. The radiative transfer scheme is further described in Pincus and Stevens (2013) and Stevens et al. (2013) and onwards we will refer to it as the radiative transfer model "PSrad".

The shortwave and longwave components are calculated separately. Furthermore, optical properties for gases, clouds and aerosols are computed separately ~~and then~~ for longwave and shortwave and, finally, combined to compute the total heating rates. PSrad expects profiles of gases (H_2O, N_2O, CH_4, CO, O_3), profile of cloud parameters as well as additional parameters (e.g., albedo and zenith angle) as input. ~~For~~ Additionally, CO_2 and O_2 are set to fixed values invariant with height. For all other gases, we use multi-year monthly means representative for the late 20th century provided by the atmospheric and chemistry model HAMMONIA (Hamburg Model for Neutral and Ionized Atmosphere; Schmidt et al. (2006)). For the albedo and cloud properties (e.g., cloud fraction, cloud water/ice content), multi-year monthly means from the piControl experiment are used. All quantities are extracted for 75° N. The zenith angle is calculated for 12 UTC at 75° N/0° E for the 15th of each month. The latitude of 75° N is chosen ~~as it represents a mean~~ exemplary for a polar latitude. The results are insensitive to the actual latitude, the main difference at other polar latitudes is the length of the polar night. Note that the length of the polar night for an air pocket depends also on the altitude and on atmospheric dynamics (e.g., movement of the polar vortex). Both effects are omitted in this study. In our simulations we reduce ozone not depending on actual dynamics but over the whole polar cap (60 – 90 °).

To quantify the impact of ozone on the heating rates, we perform multiple runs in which for each run the ozone concentration of a single layer is set to 0 once. Then we take the differences between a control run and each single run~~and a control run arecalculated and finally, added up~~. The differences of each run are, finally added up for the estimation of the total heating rate. This method allows us to consider that layers of reduced ozone will lead to increased absorption of shortwave radiation in the layers directly below.

3 Results

3.1 Ozone effects on the heating rates

An ozone loss directly alters the atmospheric energy transfer. Before analyzing circulation and climate impacts due to ozone losses, we study the heating rate response using the radiative transfer model PSrad. The heating rates are calculated for the polar latitude of 75° N (see Figure 1). As the effect of EPP is most important at the winter polar cap, we will concentrate our analysis on boreal winter high latitudes.

In the shortwave part of the spectrum, ozone strongly absorbs solar radiation and heats the whole atmosphere. The strongest heating (about 12 K/day) occurs ~~at the altitudes near the maximum volume mixing rate of ozone~~in the uppermost stratosphere around 1 hPa. An ozone loss would, hence, result in a relative cooling due to reduced heating. The ozone heating and, hence, the cooling caused by an ozone reduction are getting smaller for larger zenith angles and vanish in polar night.

In the longwave part of the spectrum, the radiative effect of ozone is highly temperature dependent. Ozone cools the atmosphere via infrared emission in the stratosphere and in warm regions of the mesosphere below 0.1 hPa (see Figure 1b). The strongest cooling (about -2 K/day) occurs at the stratopause. In the troposphere and in the cold regions of the mesosphere above 0.1 hPa, the absorption of outgoing radiation exceeds the infrared emission resulting in a heating of the atmosphere due to ozone.

In total, the shortwave heating dominates all sunlit months. During polar night, ozone cools the atmosphere between 0.1 and 100 hPa and, hence, an ozone loss in the stratosphere and lower mesosphere results in a warming. Near the terminator (e.g., at 75° N in November and February), the net influence of ozone is more complex: At some altitudes ozone heats and at some it cools the atmosphere. The net radiative forcing of an ozone loss depends on when and where ozone is reduced. For example, in November, a stratospheric (1 hPa) ozone loss leads to a heating, but a mesospheric (0.1 hPa) ozone loss to a cooling.

These results are in line with previous work. It is widely accepted that an ozone loss in spring and summer leads to a stratospheric cooling (e.g., Shine, 1986; Randel and Wu, 1999). Some studies analyzed the radiative forcing of a winter stratospheric ozone loss. Graf et al. (1998) showed that the observed stratospheric ozone loss in the late 20th century led to a winter warming and a summer cooling in a GCM. Using a radiative transfer model with fixed dynamical heating, Langematz et al. (2003) confirmed that a stratospheric ozone loss over the winter pole results in a ~~stratospheric warming~~small stratospheric radiative warming and a dominating stratospheric dynamical cooling. Shine (1986) showed that the shortwave cooling of the stratosphere due to an ozone loss dominates in all sunlit months the infrared heating due to an ozone loss. Recently, Sinnhuber et al. (2017) simulated a warming in mid-winter and a cooling in late winter and spring in the upper stratosphere for ozone losses explicitly induced by EPP.

The above stated results are confirmed by the actual heating rate anomalies induced by the applied ozone losses in the experiments 'meso-O_3' and 'strato-O_3' (not shown). The heating rates are calculated at the first time step of the model at which the radiation is updated (1 January) excluding any feedbacks occurring only at later time steps. Compared to the total heating rates of piControl, the changes in heating rates caused by a 40 % reduction of mesospheric ozone in polar night is very small (below ~~-2~~-1 % and below 0.1 K/day) and of a 20 % reduction of stratospheric ozone small (~~about -10~~below -5 % ~~).~~ and between

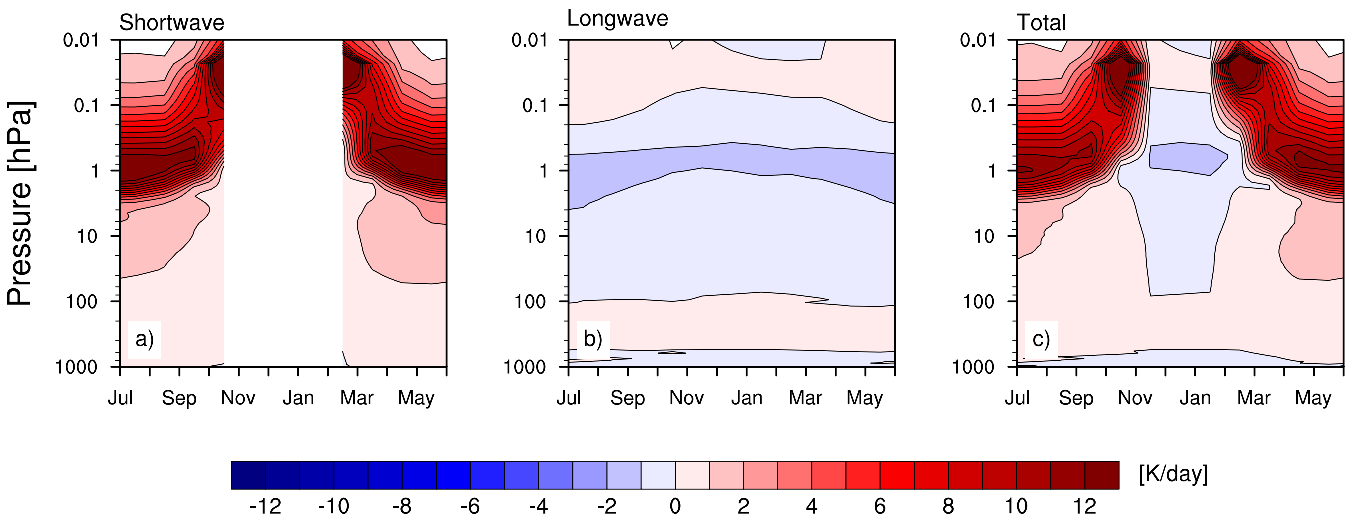

Figure 1. Monthly mean heating rates of ozone [K/day] for 75° N calculated by the radiative transfer model PSrad for (a) shortwave, (b) longwave and (c) total (shortwave + longwave) radiation.

0.1 - 0.3 K/day). This agrees with the estimate of Sinnhuber et al. (2017) , who simulated a change of 0.1 K/day in the winter stratospheric heating rate due particle-induced ozone loss.

3.2 Climate effects of a mesospheric ozone loss

As changes in heating rates due a reduced ozone during polar night are small, one might reason that climate impact of a winter polar ozone loss is small. But large effects may occur in regions slightly outside the polar night. Furthermore, several studies suggested that changes in the heating rates due to a winter polar ozone loss leads to a dynamical cooling ~~(e.g., Baumgaertner et al., 2011; Arsenovic et al., 2016)~~ (e.g., Langematz et al., 2003; Baumgaertner et al., 2011; Arsenovic et al., 2016) , whereas the initial radiative forcing suggests a warming. Hence, we further analyze the climate impact of a winter polar ozone loss. As large variations in the polar vortex can propagate downward and affect the surface climate, we first concentrate on the circulation changes of the middle atmosphere due to an ozone loss, which are a prerequisite for a potential climate impact of EPP. In the following, we analyze the climate effect of an idealized polar mesospheric ozone loss, while in Section 3.3 we analyze the climate effect of an idealized polar stratospheric ozone loss.

Figures 2a and 2d show the zonal mean temperature and zonal ~~mean~~ wind simulated for boreal winter (December - February). Main observed characteristics of the zonal mean temperature, e.g., the stratopause tilt from the summer towards the winter pole, are well reproduced. The changes in the zonal mean zonal wind are consistent with the temperature changes via the thermal wind balance. In most regions, the difference between meso-O_3 and piControl is very small (see Figures 2b and 2e).

Near the winter pole, a dipole structure emerges with cooling in the upper stratosphere and warming in the mesosphere. According to our radiative transfer calculations a mesospheric winter polar ozone loss should lead to a cooling. However, the temperature differences are small (below 1 K) and not significant at the 95 % level. As the applied forcing is very small, small and low significant values are expected. At the winter pole, the polar vortex slightly weakens, whereas the mesospheric winds strengthen: these differences are not significant. The signal is only slightly stronger but still insignificant if winters with major sudden stratospheric warmings (SSW) are excluded (not shown). As stated above, large variations in the winter polar vortex can propagate to the surface influencing the surface climate. However, the changes reported here are small. The anomalies reaching the troposphere are statistical artifacts. Indeed, the surface temperature reveals no statistically significant change (not shown).

Although, the temperature and wind signals are not statistically significant after 150 simulated years, nevertheless, it makes sense to analyze if the signals could have a physical explanation and not be purely accidental. Note that with fewer simulation years apparently very different results can be obtained. Analyzing different simulation periods we obtain mesospheric warming and cooling of apparent significance. Particularly, we calculated a statistically significant weakening of the polar vortex when using only the first 80 simulation years. We can not identify a model drift in the experiments, which could explain the disagreement between the 150-year and 80-year runs. However, the model simulates variability on time-scales up to multi-decadal, which is common in many climate models (Sutton et al., 2015) , and might cause the apparently different responses to ozone reduction in different sub-periods of the 150-year simulation. The high degree of internal variability of the winter polar stratosphere can obviously create wrong apparent signals. The most dramatic demonstration of this variability are major sudden stratospheric warmings (SSW), which occur on average about 6 times per decade in the Northern Hemisphere (see Charlton and Polvani (2007) for more information on SSW). A short simulation period may lead to an over-representation or under-representation of SSWs. Over our whole simulation period (150 years) the number of major SSWs is balanced in all three experiments. In total, there are 102 events in piControl, 99 events in meso-O_3 and 109 events in strato-O_3 (using a reversal of the zonal wind at 60° N and 10 hPa as criterion of a major SSW occurrence).

3.3 Climate effects of a stratospheric ozone loss

In this section, we analyze the climate effect of an idealized polar stratospheric ozone loss. Figures 2a and 2d show the zonal mean temperature and zonal wind simulated for boreal winter (December - February) for piControl, and Figures 2c and 2f the difference between strato-O_3 and piControl. The winter stratosphere warms due to an ozone loss as expected from the calculations with the radiative transfer model. As a consequence of the warming, the stratospheric winds weaken. The small mesospheric cooling likely results from enhanced eastward momentum deposition from gravity waves as shown by Lossow et al. (2012). Our results are in line with earlier studies. Seppälä et al. (2013) and Lu et al. (2008) identified a warming in the polar winter upper stratosphere due to EPP in reanalysis data, but their magnitude is much stronger (5 K) than in our simulations. Regarding the zonal wind response, the two studies differ from each other. Seppälä et al. (2013) analyzed a strengthening of the polar vortex with enhanced equatorward planetary waves, whereas Lu et al. (2008) analyzed a weakening of the polar vortex. The statistically significant warming of the summer mesopause is an indication of inter-hemispheric coupling as discussed by Karlsson and Becker (2016) and also persists for winters without a sudden stratospheric warming event.

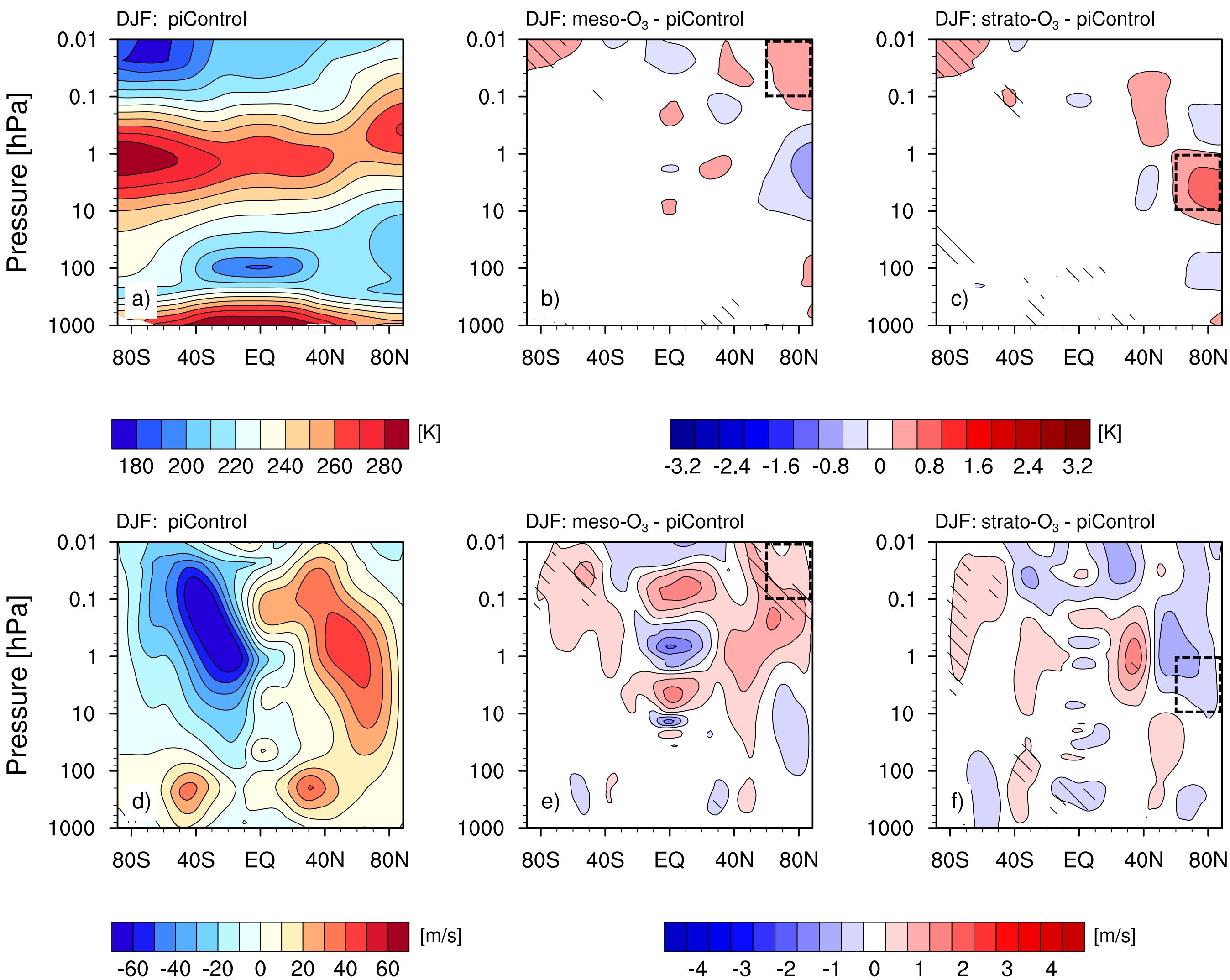

Figure 2. (upper row) Zonal mean temperature [K] and (lower row) zonal mean zonal wind [m/s] averaged over December - February (DJF) for (a,d) piControl, (b,e) the difference between meso-O_3 and piControl and (c,f) the difference between strato-O_3 and piControl. Shaded areas are statistically significant at the 95 % confidence interval. The black, dashed boxes highlight the regions where ozone is reduced.

Figure 2 shows only changes for the mean over December to February, while the radiative transfer model suggests that the month-to-month variability of the forcing is large. To study whether the impact of a stratospheric ozone loss differs over the course of the winter, we analyze the monthly means of temperature and zonal wind (see Figure 3). An ozone loss during most of the polar night (except December) leads to a warming, whereas at all other times and locations it leads to a cooling. This agrees with the calculations of the radiative transfer model and with our assumption that the winter cooling is not affected by a strong summer warming. However, the cooling in December is unexpected from the radiative transfer modeling. Kodera and Kuroda (2002) argued that the polar winter atmosphere transits from a radiatively controlled ~~statein early~~ state in early winter to a dynamically controlled state in late winter. Given the opposite sign of the diabatic forcing, the simulated cooling must be dynamically caused already in December. This is in agreement with early model studies which showed that uniform ozone losses lead to dynamical cooling at the boreal winter polar latitudes (e.g., Schoeberl and Strobel, 1978; Kiehl and Boville, 1988). Langematz et al. (2003) suggested that the dynamical cooling is due to a weakening of the mean meridional circulation related to reduced wave forcing caused by a reduction of mid-latitude wave flux into the stratosphere. Similarly, in our simulations we find a (albeit not significant) reduction of the zonal mean eddy heat flux at 100 hPa in the midlatitudes from December to March (not shown). ~~However, the origin of the changes in the wave activity remains unexplained~~This may be caused by enhanced wave reflection as suggested by Lu et al. (2017) for the dynamical response to 11-year solar irradiance forcing. Also Baumgaertner et al. (2011) reported a dynamical cooling in the winter polar stratosphere due to EPP. However, in their model the cooling dominates the winter (DJF) signal, whereas we obtain a small warming for the DJF average (see Figure 2). The magnitude of the signal decreases in our simulations, especially in late winter, if we exclude all seasons with a SSW (not shown).

The zonal wind changes consistently with the temperature changes via the thermal wind balance. Simultaneously with the warming (cooling), the polar wind weakens (strengthens). Anomalies in the polar vortex occasionally reach the troposphere (e.g., the strengthening in November or the weakening in December or February). Although, most of those changes are not significant, some disturbances in the polar vortex may still force the surface temperature (see Figure 4). In our simulations for boreal winter, stratospheric ozone loss cools large parts of the northern high latitudes from northern Europe to Eurasia and over northern America. Excluding all winter with a SSW strengthens the cooling over northern America (not shown). Over Greenland and the pole, the surface warms. This is consistent with the weakening of the zonal wind in December (see Figure 3 d). However, most changes are small and not significant. Seppälä et al. (2009) and Baumgaertner et al. (2011) analyzed statistically significant changes in surface temperature: A warming over Eurasia of about 1.5 K and a cooling over northern America of about -1 K. Compared to both studies the amplitude of our signal is much smaller. The weaker signal also persists if we exclude all winters with a SSW (not shown). However, Baumgaertner et al. (2011) based their study on only nine simulated years and we have shown that the large variability in the polar winter stratosphere can cause wrong apparent signals if the ensemble is not large enough. Seppälä et al. (2009) could not rule out that their results are by chance induced by the Northern Annual Mode.

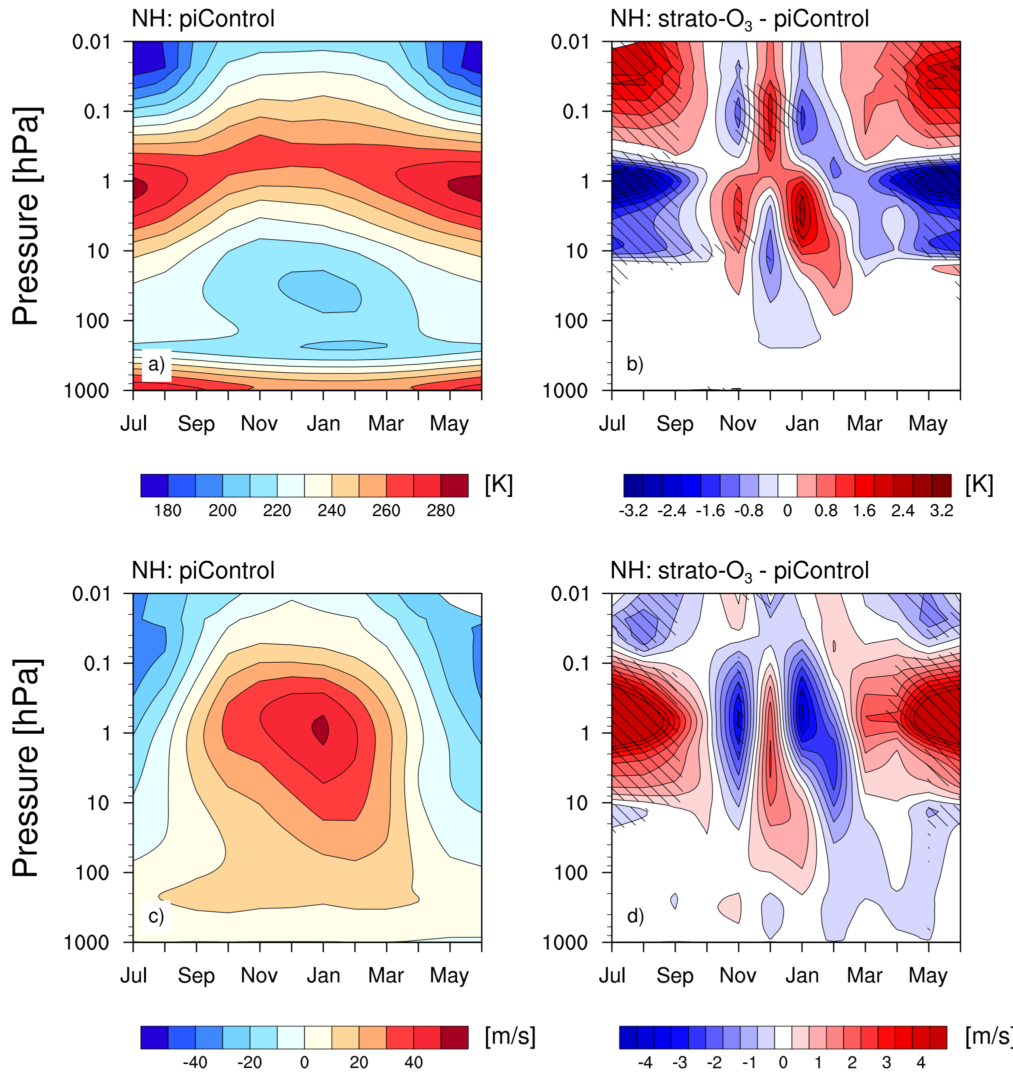

Figure 3. Monthly mean (upper row) temperature averaged between 60° N and 90° N [K] and (lower row) zonal wind [m/s] for 60° N for (a,c) piControl and (b,d) the difference between strato-O_3 and piControl. Shaded areas are statistically significant at the 95 % confidence interval.

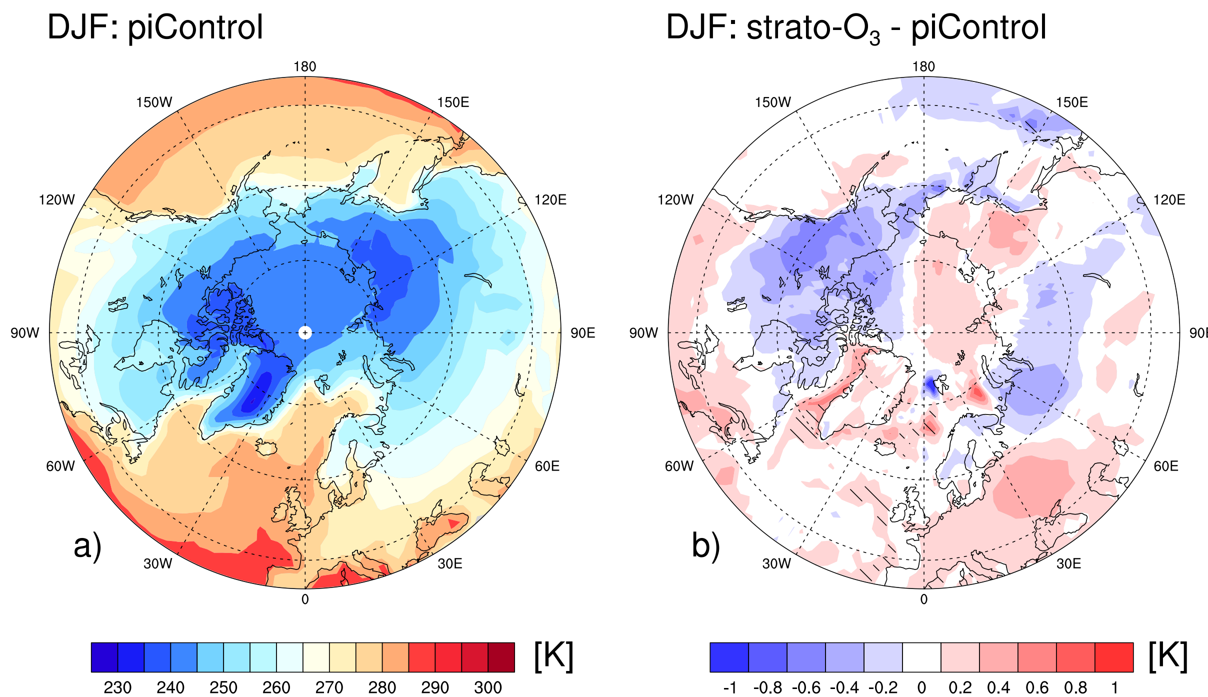

Figure 4. Surface temperature [K] for the difference between strato-O_3 and piControl (a) averaged over December - February (DJF) for the Northern Hemisphere and (b) averaged over September - November (SON) for the Southern Hemisphere. Shaded areas are statistically significant at the 95 % confidence interval.

4 Summary and conclusion

In this study, we analyzed the climate impact of mesospheric and stratospheric ozone losses. Although this study is motivated from the enhancement of NO_x due to energetic particle precipitation (EPP), the results presented here could also be applied to other processes causing ozone destruction. The radiative forcing of polar ozone is calculated by the radiative transfer model PSrad. In sensitivity studies with the Max Planck Institute Earth System Model (MPI-ESM), we reduced ozone either by 40 % in the winter polar mesosphere or by 20 % in the winter polar stratosphere.

Recently, Andersson et al. (2014) showed that the direct EPP-HO_x effect induces large long-term variability in winter mesospheric ozone. They suggested that these large changes may have an impact on climate. Following their idea, we analyzed the atmospheric response to a mesospheric ozone loss. We found that the winter atmospheric changes due to a mesospheric ozone loss in our model are negligible. Calculations with a radiative transfer model showed that the radiative forcing of mesospheric ozone is very small during polar night, which makes the small dynamic response plausible.

Several studies analyzed the climate effect of a stratospheric ozone loss due to EPP. Seppälä et al. (2009) calculated a correlation of the winter surface temperature and energetic particle precipitation in reanalysis data. However, they could not rule out an accidental occurrence of the correlation. Since then several model studies tried to establish a physical link between EPP and climate (Baumgaertner et al., 2011; Rozanov et al., 2012; Arsenovic et al., 2016). In all these model studies, a dynamical cooling of the winter polar stratosphere due to energetic particle precipitation was simulated. In our model, a stratospheric ozone loss during polar night (except December) results in a warming, whereas at all other times and locations it leads to a cooling. This agrees with the calculations of the radiative transfer model. We obtained a cooling during December due to stratospheric ozone loss caused by a reduced vertical wind. However, the changes in the polar winter stratosphere are small and not significant in our model. Consequently, also the impact on the simulated winter surface temperature is weak. In contrast to the above mentioned studies, in our experiment the dynamical feedback leading to the stratospheric cooling is not dominant throughout the boreal winter. However, the earlier model studies were based on only a few simulation years. Using only the first 80 years of our simulations we obtained false positives. The high degree of internal variability of the polar vortex can create wrong apparent signals.

~~Our results suggest that the climate impact of an ozone loss due to EPP is small.~~ As the radiative forcing of our prescribed mesospheric ozone loss is negligible, a significant climate impact of a mesospheric ozone change as suggested by Andersson et al. (2014) seems unlikely. ~~However, we cannot rule out that our experimental design with an ozone loss constant in time is too simplified, and that the time and altitude dependent loss caused by the downward propagation of NO_x concentrations would create a different result~~Our experimental design would likely rather overestimate the climate impact of EPP than underestimate it. However, our simulations indicate only small changes in the stratospheric circulation and temperature and a weak impact on surface temperature. We encourage more research on the effects of EPP as the climate impact of stratospheric ozone losses due to EPP is not as clear as often thought and the underlying processes are not well understood. The upcoming CMIP6 model intercomparison may help to resolve those open points, because energetic particle forcing is recommended - for the first time - as part of the solar forcing. (Matthes et al., 2017) . Especially the role of wave reflection for the coupling mechanism between stratosphere and troposphere needs to be clarified. Furthermore, the catalytic destruction of ozone by NO_x works only effectively if sunlight is available. The influence of EPP induced NO_x may be larger near the terminator.

Finally, although previous studies have shown that MPI-ESM reproduces stratospheric temperature responses to forcings reasonably well (e.g., Bittner et al., 2016; Schmidt et al., 2013), the possibility remains that the model's sensitivity to ozone loss is biased low. To address this, we would like to encourage multi-model studies on EPP climate impact as currently suggested for the third phase of the SOLARIS-HEPPA project, which investigates solar influences on climate as part of the 'Stratosphere-troposphere Processes And their Role in Climate' (SPARC) project.

5 Code and data availability

Primary data and scripts used in the analysis and other supplementary information that may be useful in reproducing the author's work are archived by the Max Planck Institute for Meteorology and can be obtained by contacting publications@mpimet.mpg.de.

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