

Co-Editor Decision: Publish subject to minor revisions (review by editor) (23 Jan 2020)
by [Thomas Birner](#)

Comments to the Author:

Dear Drs. Lindgren, Sheshadri,

your answers to the (mostly minor) reviewer comments look fine, but I did want to follow up on a few points and also wanted to bring up some other points that I noticed. Please make appropriate revisions, after which this paper should be acceptable for publication in WCD.

We thank Dr. Birner for taking the time to review the paper. The paper has been edited based on the suggestions of the editor. Line numbers refer to lines in the tracked-changes version of the manuscript.

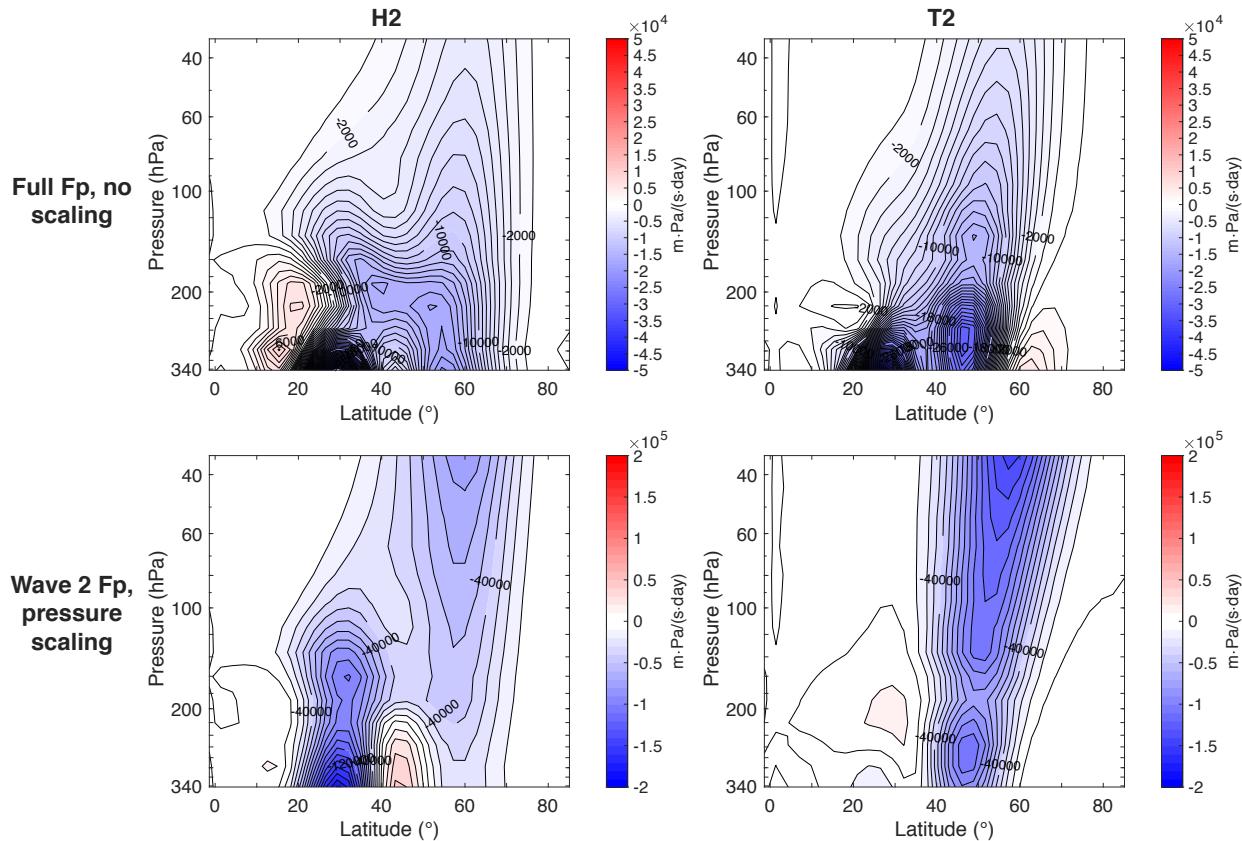
1. heating perturbation reaching 200 hPa: I don't find the figures comparing H2 and T2 in the response particularly convincing as they show the vertical EP flux in linear scale, which emphasises the troposphere (btw, is this total EP flux or just wave 2? the latter would be the appropriate quantity I think). The structure of vertical EP flux around the tropopause is almost impossible to tell from this scaling and because the strength of EP flux there is at the lowest contour level it is difficult to tell how big the difference between the two runs really is there (from a relative measure perspective the difference could be order 1). It does seem to me, as the reviewers both suggested, that with the setup of the heating perturbations you are putting some of the wave forcing into the (lowermost) stratosphere (this part therefore does not have to propagate through the tropopause). There is also evidence for this in Fig. 6b, where one can see a local minimum of wave 2 upward EP flux near 300 hPa (60 N). It'd be good to mention and discuss the implications of this setup a bit more.

The figure below shows F_p for H2 and T2 between 340 hPa and 35 hPa. The contour intervals for H2 and T2 are identical (the contour interval for the bottom row is greater than for the top row). The top row shows the full F_p for H2 and T2 without any scaling, while the bottom row only shows the wave 2 component with a p_0/p scaling, where p_0 is surface pressure (the same pressure scaling as in Figures 5 and 6). The top row shows the same thing as the figure in the response to referee file, but with a different contour interval and shorter vertical extent.

The top row shows that the magnitude of the flux around the tropopause is comparable between the two runs, although the structures of the fluxes are somewhat different. The bottom row shows only the wave 2 components, and in that case the fluxes look more different. However, the absolute majority of vertical wave flux is in the wave 2 component for T2, while H2 has more flux in wave 1. If one wants to compare the vertical wave fluxes of H2 and T2 we therefore think it is more reasonable to compare the overall fluxes rather than the wave 2 components by themselves.

Nevertheless, we have added some discussion about the implications of the vertical extent of the heating perturbation in Section 2 (L193-198). We believe that the vertical extent of the heating perturbation does not affect our results, since the transition between allowing and not allowing WWIs occurs between 50 and 30 hPa (well above

the highest extent of the heating perturbation). This location for the transition region (as opposed to the tropopause) was chosen just because it was well above regions of high wave activity around the tropopause and lower stratosphere. The rather high vertical location of the transition region is why the effects of WWIs are discussed separately in the troposphere and lower stratosphere, and middle and upper stratosphere.



2. meaning of EEIs (reviewer 1, specific comment 2): should you perhaps refer to this as wave-wave interactions (WWIs)? At least to me (and perhaps also to reviewer 1) the term "eddy-eddy interactions" provokes the connotation of smaller scale phenomena such as within a spectrum of scales in turbulence (more appropriate for tropospheric dynamics). In the stratospheric context here the interaction is mainly between two waves as you point out, so WWI may be a bit more to the point.

We have changed "eddy-eddy" to "wave-wave" and "EEIs" to "WWIs" throughout the manuscript, and renamed the model runs accordingly (NE1 to NW1, etc.) The figure labels in the Supporting information document have been edited as well.

More importantly about the removal of EEIs/WWIs: the fact that you only remove these interactions in the zonal direction, and not in the meridional, would be good to emphasise more. Right now this is only mentioned in passing (beginning of section 2) and then briefly near line 370. Otherwise it's surprising why you should get high wavenumber structures in the meridional direction (those barotropically unstable vorticity "ripples"). I do wonder to what extent these high meridional wavenumbers,

which occur much less pronounced in the full simulations (and the real atmosphere), may lead to artefacts?

We have emphasized the fact that we are removing zonal WWIs. It is mentioned in the Abstract (L3), Introduction (L125), Section 2 (L156, L158-159, L164-168), Section 4 (L384-385), and the Discussion and conclusions (L461).

We have not investigated the effects of meridional waves on the dynamics in these model runs, but we think it is unlikely that they have a large effect on the overall climatology since they only occur during vortex breakdowns, and not when the vortex is more stable. We have added this statement in the manuscript (L387-389, L464-465).

3. barotropic instability: have you also checked for symmetric/inertial instability? In any case, one question I had on this is that if you find that the necessary condition for barotropic instability is met, why do you not observe barotropically unstable waves develop? Or do you? Perhaps because the non-linear breakdown of the instability requires wave-wave interactions it doesn't end up kicking off? Please discuss somewhere in paper.

Barotropically unstable waves of wave number 2 likely develop in all model runs. We mean that barotropic instability is the most likely reason for the presence of wave 2 in the heating wave 1 model run without wave-wave interactions (NW1). We do not know of any other plausible explanation for the presence of wave 2 in this model run. Barotropically unstable waves of wavenumber 2 develop and grow in the presence of wave 1 forcing, and this is the reason why so many “splits” (wave 2 disturbances) are observed in a model run with pure wave 1 forcing and no wave-wave interactions. Hartmann (1983) and Manney et al. (1991) both showed that barotropically unstable waves of shorter wave numbers (and particularly wave number 2) develop in the presence of wave 1 forcing. This is discussed in the paper (L361-372, L401-404, L476-482).

We have not checked for symmetric/inertial instability.

4. tuning of GCMs to produce realistic SSW frequency: in the case of WACCM the Charlton et al. paper doesn't mention anything about tuning in WACCM, only that its SSW frequency is too low. So unless you have a reference where the tuning in WACCM to produce realistic SSW frequency is explicitly mentioned you should remove this part of the statement and only refer to idealised modelling studies.

The sentence has been edited to refer only to idealized modeling studies (L46-49).

The Role of **Eddy-Eddy** **Wave-Wave** Interactions in Sudden Stratospheric Warming Formation

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Abstract. The effects of **eddy-eddy** **wave-wave** interactions on sudden stratospheric warming formation are investigated using an idealized atmospheric general circulation model, in which tropospheric heating perturbations of zonal wave numbers 1 and 2 are used to produce planetary scale wave activity. **Eddy-eddy** **Zonal wave-wave** interactions are removed at different vertical extents of the atmosphere in order to examine the sensitivity of stratospheric circulation to local changes in **eddy-eddy** **wave-wave** interactions. We show that the effects of **eddy-eddy** **wave-wave** interactions on sudden warming formation, including sudden warming frequencies, are strongly dependent on the wave number of the tropospheric forcing and the vertical levels where **eddy-eddy** **wave-wave** interactions are removed. Significant changes in sudden warming frequencies are evident when **eddy-eddy** **wave-wave** interactions are removed even when the lower stratospheric wave forcing does not change, highlighting the fact that the upper stratosphere is not a passive recipient of wave forcing from below. We find that while **eddy-eddy** **wave-wave** interactions are required in the troposphere and lower stratosphere to produce displacements when wave number 2 heating is used, both splits and displacements can be produced without **eddy-eddy** **wave-wave** interactions in the troposphere and lower stratosphere when the model is forced by wave number 1 heating. We suggest that the relative strengths of wave numbers 1 and 2 vertical wave flux entering the stratosphere largely determine the split and displacement ratios when wave number 2 forcing is used, but not wave number 1.

15 1 Introduction

Sudden stratospheric warmings (SSWs) are dynamical events that can occur during hemispheric winter and which result in a collapse of the stratospheric polar vortex. During SSWs the temperature of the middle and upper polar stratosphere increases by over 30 K over a period of a few days (Butler et al., 2015), and the strongest SSWs, often called major SSWs, are usually defined by a reversal of the zonal mean westerlies at 60° and 10 hPa (Charlton and Polvani, 2007). Accurate simulations 20 of SSWs in models are crucial to capture the variability of surface climate in midlatitudes, since the effects of SSWs can migrate down to the troposphere and can impact surface weather up to two months after onset (e.g., Baldwin and Dunkerton, 2001). During SSWs the stratospheric polar vortex is either displaced from the pole as a single entity or split into two daughter vortices. These two types of SSWs are known as displacements and splits, and they are dominated by zonal wave number 1 and 2 disturbances, respectively. It has been suggested that splits and displacements are dynamically distinct (Charlton and 25 Polvani, 2007; Matthewman et al., 2009), and some studies claim that they may have different surface impacts (e.g., Mitchell

et al., 2013; Seviour et al., 2013; Lindgren et al., 2018), while others have found that differences in surface impacts are not consistent for different split and displacement classifications, and that large numbers of events are required to distinguish any differences (Maycock and Hitchcock, 2015). Major SSWs occur about every other year in the Northern Hemisphere and are roughly equally distributed between splits and displacements (Charlton and Polvani, 2007; Lindgren et al., 2018, and others).
30 Only one major SSW has been observed in the Southern Hemisphere (e.g., Allen et al., 2003).

Despite their importance for Northern Hemispheric climate variability, the dynamics behind SSW generation ~~remains~~ remain poorly understood. SSWs can occur when waves propagate from the troposphere to the stratosphere where they break and deposit momentum (e.g., Matsuno, 1971), and from this perspective SSWs can be considered wave-mean flow interactions. It has long been known that SSW-like zonal mean wind variations can occur in model setups as simple as one-dimensional
35 β -plane models (e.g., Holton and Mass, 1976; Yoden, 1990). SSW generation in general circulation models (GCMs) and the observed atmosphere, however, is likely much more complicated. Although anomalously strong tropospheric wave forcing can produce SSWs, SSW-like events have been found to occur in idealized GCMs with suppressed tropospheric variability (Scott and Polvani, 2004, 2006), and recent research has shown that only about a third of SSWs and SSW-like events are associated with anomalous tropospheric wave forcing. This has been shown in reanalysis data (Birner and Albers, 2017; de
40 la Cámara et al., 2019), chemistry-climate models (de la Cámara et al., 2019; White et al., 2019) and one idealized GCM (Lindgren et al., 2018). Birner and Albers (2017) argued that the dynamical processes responsible for most SSWs occur just above the tropopause, even though the wave forcing responsible for the events originates near the surface. The importance of stratospheric processes in SSW dynamics was also emphasized by Hitchcock and Haynes (2016), who found that the evolution of the stratospheric mean state during SSWs is crucial in determining the wave flux during the warming. Given the complexity
45 of SSW generation it is not possible to predict how frequently, or even if, SSWs will occur in a given model setup *a priori*. ~~, and idealized~~ GCMs are therefore often extensively tuned to produce realistic SSW frequencies ~~, both in idealized models~~ (e.g., Gerber and Polvani, 2009; Sheshadri et al., 2015; Lindgren et al., 2018) ~~and advanced chemistry-climate models such as the Whole Atmosphere Community Climate Model (WACCM), the first version of which only produced 0.10 SSWs per year (2).~~

50 Exactly which dynamical processes are responsible for vortex splits and displacements is also unclear. Since displacements and splits are zonal wave number 1 and 2 disturbances, respectively, one could expect that the zonal wave number of the wave flux propagating from the troposphere will determine the type of SSW produced. Large-scale topography of a single, zonal wave number has often been used to produce Northern Hemisphere winter-like stratospheric variability in idealized GCMs, and in such cases the wave number of the topography does indeed strongly influence the type of SSW produced, with wave number
55 1 (wave 1) topography favoring displacements (Martineau et al., 2018) and wave number 2 (wave 2) topography producing mostly or only splits (Gerber and Polvani, 2009; Sheshadri et al., 2015; Martineau et al., 2018; Lindgren et al., 2018). Recently, however, Lindgren et al. (2018) showed that splits and displacements occur in comparable amounts when an idealized GCM is forced with wave 1 or wave 2 tropospheric heating perturbations. Since the tropospheric forcings are of pure wave 1 or 2 format these results indicate that some wave-wave ~~(eddy-eddy)~~ interaction could be taking place somewhere between the waves being
60 forced in the troposphere and the waves breaking in the stratosphere.

Idealized models are useful tools when investigating specific dynamical processes, and simple models have frequently been used in previous studies of SSW dynamics. After one-dimensional β -plane models (e.g., Holton and Mass, 1976; Yoden, 1990) the next step in the model hierarchy includes the effects of ~~eddy-eddy interactions (EEIs)~~wave-wave interactions (WWIs), and research about the role of ~~EEIs-WWIs~~ in stratospheric dynamics dates back to the 1980s. Lordi et al. (1980) applied wave 1 and 2 geopotential height forcings at the lower boundary of a primitive equation spectral model of the stratosphere to simulate SSWs. They used spectral truncation to remove interactions between wave numbers, and only allowed the wave number of the forcing to interact with the mean flow. They concluded that nonlinear interactions are important for the evolution of the mean flow and temperature fields in the middle and upper stratosphere, especially when wave 1 forcing was used. Austin and Palmer (1984) used a primitive equation model of the stratosphere and mesosphere to investigate the role of nonlinear effects in setting up the monthly mean wave amplitudes in the stratosphere during December 1980, and concluded that ~~EEIs-WWIs~~ cannot be ignored for accurate simulations of the middle atmosphere.

O'Neill and Pope (1988) carried out an extensive study of linear and nonlinear disturbances in the stratosphere resulting from tropospheric perturbations, with the same primitive equations model that Austin and Palmer (1984) used. They found that linear theories of wave propagation and wave-mean flow interaction were useful when the imposed lower boundary forcing was weak, but that the stratospheric flow was highly nonlinear in the latter stages of simulations with strong forcing. Their results suggested that major stratospheric warmings evolve from highly asymmetric states through complicated nonlinear interactions, rather than simple wave-mean flow interactions.

Robinson (1988) simulated a minor wave 1 stratospheric warming in a primitive equation model, while running the model in a fully nonlinear mode as well as in a quasi-linear mode, with waves of only one wave number and the zonal flow. The model runs were 60 days long following the spin-up period. He found that irreversible modifications of the potential vorticity field were stronger and more localized when ~~EEIs-WWIs~~ were allowed, and that the differences between the nonlinear and quasi-linear model runs come from modifications of the interactions between wave 1 and the zonal flow by shorter waves. The interactions between shorter waves and the mean flow were found to be of much less importance when accounting for the differences between the experiments.

There is also observational evidence to suggest that ~~EEIs-WWIs~~ are largely responsible for modulating the relative strengths of wave 1 and wave 2 disturbances in the stratosphere. Smith et al. (1983) used satellite data to investigate the importance of wave-mean flow interactions versus ~~EEIs-WWIs~~ in the Northern Hemisphere during the 1978-79 winter, and found that oscillations between wave 1 and wave 2 in the geopotential height field could be largely attributed to ~~EEIs-WWIs~~ in the stratosphere, rather than forcing from the tropopause region.

The authors mentioned above have shown that ~~EEIs-WWIs~~ play an important role in stratospheric dynamics in general and SSW dynamics in particular, but many questions regarding the role of ~~EEIs-WWIs~~ in SSW dynamics remain unanswered. One major restriction of the previous studies is that spectral truncation was used to eliminate ~~EEIs-WWIs~~, while keeping only the interactions between a single wave number and the mean flow (Lordi et al., 1980; Robinson, 1988). The climatological wave flux in the observed stratosphere contains both wave 1 and wave 2 components (e.g., Birner and Albers, 2017), both of which can interact with the mean flow. Given that displacements and splits are wave 1 and wave 2 disturbances, respectively,

removal of either of the major stratospheric wave numbers along with **EEIs-WWIs** will enable only one type of SSW to form, even though SSWs in the observed atmosphere are roughly equally distributed between splits and displacements (Charlton and Polvani, 2007; Lindgren et al., 2018, and others). Other limitations of previous work include the short temporal extents and coarse resolutions of the model runs, due to the limited computing resources of the 1980s. It is known that **EEIs-WWIs** are
100 important both for SSW generation (O'Neill and Pope, 1988) and the evolution of the stratospheric mean state during SSWs (Hitchcock and Haynes, 2016) but the effect that **EEIs-WWIs** have on SSW frequencies has not been investigated, and it is not clear *a priori* whether **EEIs-WWIs** act to enhance or diminish SSW generation.

Furthermore, previous work has removed **EEIs-WWIs** throughout the vertical extent of the models even though the dynamics of the troposphere, tropopause region, and stratosphere are very different. Birner and Albers (2017) emphasized the importance
105 of dynamics in the 300-200 hPa region (just above the tropopause) for SSW formation. Polvani and Waugh (2004) found that anomalous wave fluxes at 100 hPa and further down in the troposphere precede SSWs and deduced that the origin of SSWs can be found in the troposphere (although they acknowledged the fact that the stratosphere may play a role in modulating the events). These results indicate that dynamical processes in the troposphere and lower stratosphere are crucial in SSW formation, which raises the question as to whether or not the role of **EEIs-WWIs** in SSW dynamics varies accordingly. The
110 recent results of Lindgren et al. (2018) show that both splits and displacements can form with a tropospheric forcing of a single wave number, indicating that **EEIs-WWIs** could act to transfer energy from the wave number of the forcing to the other major stratospheric wave number somewhere between the troposphere and the stratosphere. In order to understand the importance of **EEIs-WWIs** for SSW generation as well as split and displacement distributions at different vertical levels in the atmosphere, a different method of removing **EEIs-WWIs** must be used.

115 More recently, the effects of **EEIs-WWIs** on atmospheric dynamics has been investigated in idealized models by calculating the advection of eddy fluctuations by eddy winds, and replacing them with their zonal mean values. Unlike the spectral truncation mentioned in the papers above, this method allows climatological wave flux of all wave numbers to interact with the mean flow, and it retains much of the zonal mean climatology of (nonlinear) control runs. The method has been successfully used to investigate tropospheric dynamics by several authors: O'Gorman and Schneider (2007) removed **EEIs-WWIs** in an idealized
120 GCM and found that the atmospheric kinetic energy spectrum retained its wave number dependence when **EEIs-WWIs** were removed, even though this dependence was previously thought to be determined by **EEIs-WWIs**. Chemke and Kaspi (2016) showed that **EEIs-WWIs** in an idealized GCM actually decrease the number of eddy-driven jets in the atmosphere by narrowing the latitudinal region where zonal jets can appear. This method has also proven useful when applied to theories of jet dynamics in beta-plane models (e.g., Srinivasan and Young, 2012; Tobias and Marston, 2013; Constantinou et al., 2014).

125 In this paper we investigate the role of **EEIs-WWIs** in SSW formation by removing the effects of zonal **EEIs-WWIs** in an idealized GCM, using the method of O'Gorman and Schneider (2007). We use the model output produced with heating wave 1 (H1) and wave 2 (H2) tropospheric forcing from Lindgren et al. (2018), both of which produce splits and displacements in comparable amounts even though the forcings are of a single wave number. This makes the model setups ideal for studying the role of **EEIs-WWIs** in SSW generation in general, and split and displacement formation in particular. We perform model runs
130 under three additional settings for each forcing: one without **EEIs-WWIs** anywhere, one without **EEIs-WWIs** in the troposphere

and lower stratosphere, and one without ~~EEIs~~~~WWIs~~ in the middle and upper stratosphere. The latter two, hereafter referred to as the mixed runs, allow us to investigate the effects of removing ~~EEIs~~~~WWIs~~ below (above) the vertical levels that Polvani and Waugh (2004) and Birner and Albers (2017) highlighted as crucial for SSW generation, while still allowing ~~EEIs~~~~WWIs~~ above (below). By comparing the results of the mixed runs to the fully nonlinear control runs and the model runs without ~~EEIs~~~~WWIs~~ 135 anywhere, we can investigate the importance of ~~EEIs~~~~WWIs~~ in the middle and upper stratosphere when the climatology (and hence mean wave forcing) in the troposphere and lower stratosphere remains unchanged. We use the model runs to answer three questions related to ~~EEIs~~~~WWIs~~ and SSW formation that have not been investigated before:

1. To what extent do ~~EEIs~~~~WWIs~~ affect SSW frequencies?
2. To what extent are the numbers of splits and displacements, and their ratios, affected by ~~EEIs~~~~WWIs~~?
- 140 3. How do ~~EEIs~~~~WWIs~~ affect SSWs frequencies and split and displacement ratios when the lower stratospheric wave forcing does not change?

We find that the effects of ~~EEIs~~~~WWIs~~ on SSWs are strongly dependent on the wave number of the tropospheric forcing. We show that the absence of ~~EEIs~~~~WWIs~~ can significantly alter SSW frequencies, and that whether ~~EEIs~~~~WWIs~~ increase or decrease SSW frequencies depends on the tropospheric forcing that is used and the vertical levels where ~~EEIs~~~~WWIs~~ are 145 removed. Significant changes in SSW frequencies between model runs with and without ~~EEIs~~~~WWIs~~ in the middle and upper stratosphere are obtained even though the lower stratospheric wave forcings of the model runs do not change, highlighting the fact that the stratosphere is not a passive recipient of lower stratospheric wave forcing. We further find that while ~~EEIs~~~~WWIs~~ are required in the troposphere and lower stratosphere to produce displacements when wave 2 forcing is used, both splits and displacements can be produced without ~~EEIs~~~~WWIs~~ in the troposphere and lower stratosphere when the model is forced by 150 wave 1 heating.

Section 2 describes the method used to remove the effects of ~~EEIs~~~~WWIs~~, and the way it was implemented in the model runs. Section 3 describes changes in climatology caused by removal of ~~EEIs~~~~WWIs~~. Section 4 compares the SSW frequencies, split and displacement ratios, and polar vortex strength variabilities of the eight different model runs. A discussion of the results and conclusions can be found in Sect. 5.

155 2 Removal of ~~eddy-eddy~~~~wave-wave~~ interactions

Following O’Gorman and Schneider (2007), we calculated the tendency due to ~~EEIs~~~~(the zonal WWIs (the *eddy-eddy* wave-wave tendency)~~, subtracted it from the total tendency of horizontal wind and temperature, and added the zonal mean value of the ~~eddy-eddy~~~~wave-wave~~ tendency (the *mean tendency*) to the total tendency equation. This method removes interactions be-

tween zonal waves. This substitution was described by O’Gorman and Schneider (2007) by using the equation for temperature

160 tendency as an example. In the control run the evolution is

$$\begin{aligned}\frac{\partial T}{\partial t} &= -v \frac{\partial T}{\partial y} + \dots, \\ &= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v} \frac{\partial T'}{\partial y} - v' \frac{\partial \bar{T}}{\partial y} - v' \frac{\partial T'}{\partial y} + \dots,\end{aligned}\quad (1)$$

where overbars denote zonal means while primes show deviations from the zonal mean (eddies). Only the terms related to meridional advection of temperature have been written out. The last term in Eq. (1) describes the advection of temperature eddies due to meridional wind eddies. In the model runs where ~~EEIs-WWIs~~ are not allowed Eq. (1) becomes

$$\begin{aligned}\frac{\partial T}{\partial t} &= -v \frac{\partial T}{\partial y} + \dots, \\ &= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v} \frac{\partial T'}{\partial y} - v' \frac{\partial \bar{T}}{\partial y} - \overline{v' \frac{\partial T'}{\partial y}} + \dots,\end{aligned}\quad (2)$$

where the contribution due to ~~EEIs-WWIs~~ has been replaced by its zonal mean value. It should be noted that this method only removes interactions between zonal waves, and that meridional WWIs are still allowed. However, it is zonal flow over the continents and oceans as well as large scale topography that produces the planetary scale waves which propagate into the stratosphere, and the WWIs important for stratospheric dynamics therefore occur between zonal waves. Unless otherwise noted, WWIs hereafter refer to zonal WWIs.

Two model runs from Lindgren et al. (2018) are used in this paper. The GCM is a dry, hydrostatic, global primitive equation 170 model with T42 resolution in the horizontal and 40 vertical σ levels, where $\sigma = p/p_s$. The model setup is based on that of Polvani and Kushner (2002). There are no convection or radiation schemes in the model, and temperature is relaxed towards a zonally symmetric temperature profile through Newtonian relaxation. The temperature profile is symmetric about the equator in the troposphere and identical to that of Held and Suarez (1994), but set to perpetual Northern Hemisphere midwinter conditions in the stratosphere. The transition between tropospheric and stratospheric equilibrium temperature profiles occurs at 200 hPa. 175 Tropospheric diabatic heating perturbations are used to produce Northern Hemisphere winter-like stratospheric variability. The reader is referred to Lindgren et al. (2018) for more information about the model and heating perturbations. In addition to the H1 and H2 runs from Lindgren et al. (2018) another three runs were performed for each forcing wave number, where each of the three additional runs removed ~~EEIs-WWIs~~ in different parts of the atmosphere. All model set ups were run under Northern Hemisphere winter conditions for 31,100 days, and the last 30,000 days were used for the analysis. The vertical structures of 180 where ~~EEIs-WWIs~~ are permitted and removed can be found in Figs. 1 and 2.

In the control runs (H1 and H2; black line in Figs. 1 and 2) ~~EEIs-WWIs~~ are allowed everywhere. In the no ~~EEIs-anywhere runs (NE1 or NE2)~~ WWIs-anywhere runs (NW1 or NW2) depending on the wave number of the forcing; red line) the effects of ~~EEIs-WWIs~~ were removed at each pressure level. In the mixed runs the model is set up to switch between allowing and not allowing ~~EEIs-WWIs~~ linearly with pressure. The transition occurs between 50 and 30 hPa. In the case of the runs with no ~~EEIs~~

185 WWIs allowed in the troposphere and lower stratosphere (hereafter shortened as NEt1 or NEt2NWt1 or NWt2; green line) the following substitution is made:

$$\frac{\partial T}{\partial t} = \begin{cases} \frac{\partial T}{\partial t} - v' \overline{\frac{\partial T'}{\partial y}} + v' \frac{\partial T'}{\partial y}, & p > p_1, \\ \frac{\partial T}{\partial t} + \left(1 - \frac{p_1-p}{p_1-p_2}\right) \cdot \left(-v' \overline{\frac{\partial T'}{\partial y}} + v' \frac{\partial T'}{\partial y}\right), & p_2 \leq p \leq p_1, \\ \frac{\partial T}{\partial t}, & p < p_2. \end{cases} \quad (3)$$

In the above equation the temperature tendency from Eqs. 1 and 2 has been used as an example. $p_1 = 50$ hPa and $p_2 = 30$ hPa. Similarly, when EEIsWWIs are not allowed in the upper stratosphere (hereafter shortened as NEs1 or NEs2NWs1 or NWs2; blue line) the equation describing the substitution is

$$\frac{\partial T}{\partial t} = \begin{cases} \frac{\partial T}{\partial t}, & p > p_1, \\ \frac{\partial T}{\partial t} + \frac{p_1-p}{p_1-p_2} \cdot \left(-v' \overline{\frac{\partial T'}{\partial y}} + v' \frac{\partial T'}{\partial y}\right), & p_2 \leq p \leq p_1, \\ \frac{\partial T}{\partial t} - v' \overline{\frac{\partial T'}{\partial y}} + v' \frac{\partial T'}{\partial y}, & p < p_2. \end{cases} \quad (4)$$

190 The mixed runs enable an investigation of the effects of EEIsWWIs in different regions of the atmosphere. Although it may seem an obvious choice to put the transition region around the tropopause this alters the climatologies of the mixed runs significantly compared to the control runs, something that is likely caused by substantial changes in the strong climatological wave convergence in the tropopause region just above the tropopause (see Figs. 3 and 4). Instead, the Furthermore, the heating perturbation from Lindgren et al. (2018) reaches up to 200 hPa, which puts a small part of the perturbation in the lower stratosphere. Therefore, the 50 hPa and 30 hPa levels were chosen as start and end points of the transition region. This choice has two three strong advantages over the tropopause region: first, it is an unusually calm region of the atmosphere in terms of wave activity and changes in wave interactions do not affect the climatology as strongly. Second, the it is located well above the extent of the heating perturbation, so no part of the wave forcing crosses the transition region. Third, the pressure levels where lower stratospheric wave forcing have been highlighted as important for SSWs range from 300-200 hPa (Birner and 195 Albers, 2017) to 100 hPa (Polvani and Waugh, 2004). The choice of a transition region between 50 and 30 hPa as opposed to the tropopause means that we can keep what is thought to be the most important wave forcing identical between model runs 200 that allow (remove) EEIsWWIs everywhere and turn EEIsWWIs off (on) in the middle and upper stratosphere.

3 Climatology in the absence of EEIsWWIs

Figure 2 shows the climatological zonal mean zonal wind for the four model runs with wave 2 heating, along with panels 205 indicating the pressure levels where EEIsWWIs are allowed. Zonal mean zonal winds for model runs with wave 1 heating

can be found in the supporting information (Fig. S1). A comparison of H2 (Fig. 2a) to NE2-NW2 (Fig. 2b) shows that the model retains much of the climatological zonal mean zonal wind structure even in the absence of EEIs-WWIs, although with some notable exceptions. For one, the polar night jet is much more separated from the tropospheric jet when EEIs-WWIs are not allowed in the troposphere and lower stratosphere (panels b and d). The jet strength, however, is largely unaffected 210 in NE2-NW2. This is not the case in all model runs, especially NEs2-NWs2 (Fig. 2c) compared to the other model runs with wave 2 heating. The area where the zonal mean zonal wind approaches zero ms^{-1} found in the equatorial lower stratosphere in the control runs is not reproduced when EEIs-WWIs are not allowed in the troposphere and lower stratosphere (panels b and d), indicating that EEIs-WWIs play an important role in this area. Furthermore, there are two tropospheric jets in the 215 Southern Hemispheres of these runs. O’Gorman and Schneider (2007) and Chemke and Kaspi (2016) also obtained additional tropospheric jets in their models when removing EEIs-WWIs, and Chemke and Kaspi (2016) showed that EEIs-WWIs decrease the number of eddy-driven jets in the atmosphere.

Comparisons between the mixed runs and H2 or NE2-NW2 show that changes in the middle and upper stratosphere have very little influence on the climatology of the troposphere and lower stratosphere. When EEIs-WWIs are allowed in the troposphere and lower stratosphere only (NEs2-NWs2; Fig. 2c) the zonal mean zonal winds at vertical levels below 50 hPa are very similar 220 to those of the control run (H2; Figs. 2a). Similarly, when EEIs-WWIs are not allowed in the troposphere and lower stratosphere (NEt2-NWt2; Fig. 2d) the climatology at vertical levels below 50 hPa is very similar to that of the model run where EEIs-WWIs are not allowed anywhere (NE2-NW2; Fig. 2b). The same is true when wave 1 heating is used; see Fig. S1.

To investigate the changes in wave activity caused by removing EEIs-WWIs we calculated the climatological wave 1 and wave 2 components of vertical Eliassen-Palm (EP) flux (F_p) and divergence of EP flux for the eight model runs. The calculations were based on Edmon et al. (1980), and are identical to those found in Lindgren et al. (2018). Although waves of higher 225 wave numbers also interact with each other and the mean flow (especially in the troposphere), we focus our attention on wave numbers 1 and 2 since only large-scale waves can propagate into the stratosphere (e.g., Charney and Drazin, 1961), and since these are the wave numbers of the tropospheric forcings. Figures 3 through 6 show the wave 1 and wave 2 components of the two quantities. The most important result of the EP flux figures is that, just like the zonal mean zonal wind in Fig. 2, the 230 divergence of EP flux and F_p at vertical levels below 50 hPa depend on whether or not EEIs-WWIs are allowed at these levels: NE1 and NE2-NW1 and NW2 look very similar to NEt1 and NEt2-NWt1 and NWt2, while H1 and H2 look like NEs1 and NEs2-NWs1 and NWs2 in this region. This indicates that changing the conditions for EEIs-WWIs above 50 hPa does not affect the climatological wave forcing from lower levels, and it will enable us to shed light on the importance of the middle and upper stratosphere for SSW generation compared to the 300-200 hPa (Birner and Albers, 2017) and 100 hPa (Polvani and Waugh, 235 2004) levels highlighted by previous authors.

There are substantial wave 1 and wave 2 EP flux divergence components in the stratosphere of both H1 and H2 (panels a and b in Figs. 3 and 4), which shows that there is a large amount of both wave 1 and wave 2 activity in the two control runs. This results in large numbers of both splits and displacements in both runs (Lindgren et al., 2018). In contrast, removal of EEIs-WWIs everywhere results in an EP flux divergence completely dominated by the wave number of the forcing, with practically 240 only wave 1 and no wave 2 EP flux convergence in NE1-NW1 (Fig. 3c and d) while the opposite is true for NE2-NW2 (Fig.

4c and d). This result is not surprising: removal of ~~EEIs-WWIs~~ means that waves can only interact with the mean flow, which excludes the possibilities of energy transfer between wave numbers. In contrast, there are areas of EP flux divergence of a wave number different from that of the tropospheric forcing just above the transition region in ~~NEt1 and NEt2~~ ~~NWt1 and NWt2~~ (Figs. 3f and 4e), and areas of EP flux convergence in the same region in the wave number of the forcing (Figs. 3e and 4f).

245 This suggests that once ~~EEIs-WWIs~~ are allowed some of the wave activity is transferred from the wave number of the forcing to the other of the two major stratospheric wave numbers. Another aspect worth noting is that the areas of strongest F_p and EP flux convergence occur further poleward when ~~EEIs-WWIs~~ are allowed in the stratosphere compared to when they are not. This does not seem to be a result of changing zonal wind climatology and hence a shift in the structure of the wave guide, since the latitudinal polar night jet shifts between the model configurations are modest (Figs. 2 and S1). Robinson (1988) 250 found that removal of all waves but wave 1 resulted in a more equatorward wave flux compared to his nonlinear model run. He deduced that this difference came from the removal of interactions between wave 1 and shorter waves, and that interactions between shorter waves and the mean flow had a comparably small effect on the wave flux. Our results confirm the conclusion of Robinson (1988) since our experiments allow interactions between all waves and the mean flow, and the equatorward shift persists.

255 4 Impacts on SSWs and polar vortex strength

Table 1 shows the SSW frequencies, time mean and variability of zonal mean zonal wind at 10 hPa and 60° N, and split and displacement ratios for the eight runs. A SSW is defined here as a reversal of the zonal mean westerlies at 60° N and 10 hPa with at least 20 days of consecutive westerlies between events (Charlton and Polvani, 2007). The SSW frequency is one way to characterize the variability of the stratospheric polar vortex, but it is a binary definition and not wholly representative of 260 stratospheric variability. The time mean and variability of zonal mean zonal wind at the same pressure level and latitude therefore provide further metrics for how changes in ~~EEIs-WWIs~~ affect the polar vortex strength. The wave amplitude classification (WAC) introduced by Lindgren et al. (2018) was used to classify the SSWs as splits or displacements. We use a Monte Carlo approach to assess the statistical significance of differences in SSW frequencies and split and displacement ratios, at a 95 % confidence level; the method is described in detail in the supporting information.

265 Removal of ~~EEIs-WWIs~~ affects SSW frequencies significantly when the model is forced by wave 1 heating. The SSW frequency in H1 is 0.66 SSWs per 100 days, but in ~~NEt1~~ ~~NWt1~~ the frequency is increased to 0.82 (a 24 % increase). The frequencies are lower in the mixed runs: 0.44 in ~~NEt1~~ ~~NWt1~~ and 0.31 in ~~NEs1~~ ~~NWs1~~ (decreases of 34 % and 53 % compared to the control run, respectively). The differences between these model runs are all statistically significant. It is reasonable that the SSW ~~frequency-frequencies~~ in the wave 1 runs should be affected by removal of ~~EEIs~~ ~~since EEIs-WWIs, since WWIs~~ affect 270 the climatological wave forcing quite strongly with this heating perturbation. Figure 5 shows that there are strong tropospheric wave 1 and wave 2 components of F_p in H1, but that the wave 2 component disappears when ~~EEIs-WWIs~~ are not allowed in the troposphere. However, the results from the mixed runs are more surprising: removal of ~~EEIs-WWIs~~ in the middle and upper stratosphere only (~~NEs1~~ ~~NWs1~~) decreases the SSW frequency by over 50 % compared to the control run. Similarly,

allowing ~~EEIs-WWIs~~ in the middle and upper stratosphere only (~~NEt1-NWt1~~) decreases the SSW frequency by 45 % compared
275 to ~~NE1-NW1~~. As was mentioned in the previous section, the tropospheric and lower stratospheric wave forcing depends on whether or not ~~EEIs-WWIs~~ are allowed at these levels, and not on the conditions in the middle and upper stratosphere. The fact that the wave forcings and climatologies below 50 hPa in H1 and ~~NEs1-NWs1~~ as well as ~~NE1-and NEt1-NW1 and NWt1~~ are practically identical while their SSW frequencies are very different shows that ~~EEIs-WWIs~~ in the stratosphere play a major role in SSW generation. Supporting the conclusions of Hitchcock and Haynes (2016), it also highlights the fact that the upper
280 stratosphere is not a passive recipient of wave forcing from below, even though the importance of tropospheric and lower stratospheric wave forcing for SSW generation has often been emphasized (Polvani and Waugh, 2004; Birner and Albers, 2017). Like O'Neill and Pope (1988), we find that SSWs cannot simply be considered forced by wave-mean flow interactions from a lower boundary. However, there is no clear answer to how ~~EEIs-WWIs~~ in the middle and upper stratosphere influence SSW frequencies with wave 1 forcing: a comparison between H1 and ~~NEs1 suggests that EEIs-NWs1 suggests that WWIs~~ are
285 necessary in the middle and upper stratosphere to get high SSW frequencies, while the results for ~~NE1-and NEt1-NW1 and NWt1~~ indicate that allowing ~~EEIs-WWIs~~ in the middle and upper stratosphere decreases the SSW frequency.

In contrast to the runs with wave 1 forcing, removal of ~~EEIs-WWIs~~ in the troposphere and lower stratosphere does not significantly alter the SSW frequency when wave 2 forcing is used: ~~NE2-and NEt2-NW2 and NWt2~~ have SSW frequencies of 0.51 and 0.45, compared to 0.48 in the control run. This is not surprising considering the fact that almost all tropospheric and
290 lower stratospheric wave forcing in H2 is in the wave number of the forcing (Fig. 6), so removal of ~~EEIs-WWIs~~ does not affect the climatological forcing as strongly as when wave 1 forcing is used. However, when ~~EEIs-WWIs~~ are removed in the middle and upper stratosphere only (~~NEs2-NWs2~~) the SSW frequency increases by 37 % compared to H2. The differences between this frequency and those of H2, ~~NE2-and NEt2-NW2 and NWt2~~ are statistically significant. This increase in SSW frequency could be the reason for the weakened climatological polar night jet seen in Fig. 2c, although another explanation is that removal of
295 ~~EEIs-WWIs~~ weakens the polar night jet. This weakened jet would then require less wave forcing to create SSWs, which would increase the SSW frequency. Like the case of the mixed runs with wave 1 forcing, the increase in SSW frequency shows that middle and upper stratospheric ~~EEIs-WWIs~~ play a major role in SSW generation. Interestingly, this change in SSW frequency is completely different from the one between ~~NE1-and NEs1-NW1 and NWs1~~: with wave 2 forcing the SSW frequency increases without ~~EEIs-WWIs~~ in the middle and upper stratosphere, while the opposite is true with wave 1 forcing.

300 Unlike the SSW frequencies, some changes in mean polar vortex strength and variability, as measured by the standard deviation of polar vortex strength, are consistent for model runs with both wave 1 and wave 2 forcing. In both cases the standard deviation of polar vortex strength is highest in the control run (H1 and H2), and lowest when ~~EEIs-WWIs~~ are removed in the troposphere and lower stratosphere (~~NEt1-and NEt2-NWt1 and NWt2~~). As can be expected, low mean polar vortex strengths and high standard deviations are correlated with high SSW frequencies. More revealing than these numbers are the
305 time evolutions of polar vortex strength, seen in Fig. 7 for all eight model runs during 2000 days of the model simulations. The 2000 days are typical for the model runs. The data has been smoothed with a 10 day filter for clarity, and Fig. S2 in the supporting information shows the same data unfiltered. Figure 7a shows how removal of ~~EEIs-WWIs~~ affect the time evolution of the polar vortex strength when wave 1 heating is used. The strength of the polar vortex is highly variable in H1 (black), and

has maximum and minimum strengths higher and lower than any of the other three model runs. Table 1 indicated that the polar
310 vortex strength in **NE1-NW1** (red) is both lower and less variable than H1, and Fig. 7a confirms this. The variations also seem
to be of shorter duration in **NE1-NW1** compared to H1. Based on the figure, the polar vortex strength in **NEs1-NWs1** (blue)
is much more similar to that in H1 than to the one in **NE1-NW1**. The polar vortex in **NEs1-NWs1** exhibits the same long term
variability as H1, but does not become strongly negative as frequently as the one in H1. In Fig. 7a the polar vortex in **NEt1**
315 **NWt1** (green) exhibits little variability. From the figure it may seem like SSWs are infrequent in **NEt1-NWt1** compared to the
other model runs, but it actually has a higher SSW frequency than **NEs1-NWs1**. Instead, the polar vortex **NEt1-NWt1** exhibits
a lot of short term variability that is filtered out in Fig. 7a, but is visible in Fig. S2a.

The latitude of maximum polar vortex strength changes slightly between the model runs (from about 63° N to 71° N; see
Figs. 2 and S1), and the relative mean values and variabilities of polar vortex strength therefore change when the latitude of
320 interest is changed. However, the timescales of variability are qualitatively similar for given model runs with modest changes
in latitude and pressure. Figures 7 and S2 can therefore be thought of as reasonable representations of overall polar vortex
behavior.

As in the case of H1, the polar vortex strength in H2 has a larger variability than in any of the model runs with **EEIs-WWIs**
removed (Fig. 7b). The polar vortex strength in H2 varies on timescales up to 1000 days; much longer than any other model
runs. As in the case of **NE1-NW1** and H1, the polar vortex in **NE2-NW2** has a lower mean strength and lower variability than
325 that in H2, and the timescale of the variability is much shorter than in H2. Unlike the case with wave 1 heating, the polar vortex
in **NEs2-NWs2** has a mean strength and variability that resembles that of **NE2-NW2** more than that of H2. Just like in the
case of **NEt1-NWt1**, the variability of polar vortex strength in **NEt2-NWt2** is low, and much of the variability happens on short
timescales which are filtered away in Fig. 7. However, unlike the case of **NEt1-NWt1** the polar vortex strength in **NEt2-NWt2**
does not seem to vary on long timescales.

330 Figure 7 tells us much about how **EEIs-WWIs** affect polar vortex strength and, by extension, SSW frequencies. The structure
of the polar vortex variability in **NEs1-NWs1** is similar to that of H1, indicating that **EEIs-WWIs** in the troposphere and lower
stratosphere are important for much of the long term (a few hundred days) variability of polar vortex strength when wave 1
heating is used. The fact that the SSW frequency is much lower in **NEs1-NWs1** compared to H1 could indicate that middle and
335 upper stratospheric **EEIs-WWIs** are important in order to strongly disturb the polar vortex, at least when the lower stratospheric
wave forcing is obtained with wave 1 heating in the presence of **EEIs-WWIs**. The similarities in polar vortex behavior between
NE2 and NEs2-NW2 and NWs2, on the other hand, indicate that **EEIs-WWIs** in the troposphere and lower stratosphere are
not as important when it comes to variability of the polar vortex (although the SSW frequency is higher in **NEs2 compared
to NE2-NWs2 compared to NW2**). This is likely due to the fact that most of the tropospheric wave forcing in all model runs
with wave 2 heating is in wave 2 (Fig. 6). This is not the case with wave 1 heating, where the tropospheric wave forcing
340 has substantial wave 1 and wave 2 components in H1 and **NEs1 but not NE1 and NEt1-NWs1 but not NW1 and NWt1** (Fig.
5). The differences between H2 and **NEs2 show that EEIs-NWs2 show that WWIs** in the middle and upper stratosphere are
crucial in setting the long term behavior of the polar vortex when wave 2 heating is used. **EEIs-WWIs** seem to strongly alter
the structure of the stratosphere, which may alter the amount of wave forcing that can propagate up from below. The fact that

the polar vortices in ~~NEt1 and NEt2~~ NWt1 and NWt2 exhibit the lowest amounts of variability, and variability on the shortest timescales, could indicate that much of any anomalous tropospheric and lower stratospheric wave flux is damped away in the transition region between 50 and 30 hPa. Comparisons of total EP flux convergence (not shown) support this: compared to ~~NEt1 and NEt2~~ NWt1 and NWt2, there is stronger EP flux convergence in the transition region of ~~NEt1 and NEt2~~ NWt1 and NWt2, and less EP flux convergence further up in the stratosphere. The wave convergence in the transition region could produce low frequency variability in the polar vortex strength and leave less wave flux to converge further up in the stratosphere, where it would likely affect the polar vortex more strongly.

Table 1 also contains the numbers and fractions of splits in the model runs. 59 % of SSWs in H1 are splits, and this number is increased to 80 % in ~~NEt1~~ NWt1 (a statistically significant difference). This result seems counterintuitive: as was mentioned above there is strong climatological tropospheric wave 1 and wave 2 flux in H1, while the tropospheric forcing is almost pure wave 1 when ~~EEIs~~ WWIs are not allowed. A possible explanation for this is that much of the wave activity is transferred from wave 1 to wave 2 in the upper stratosphere, where ~~EEIs~~ WWIs are allowed. Panels e and f in Fig. 3 show the wave 1 and wave 2 EP flux divergence for ~~NEt1~~ NWt1. The panels show that while the EP flux convergence in the upper stratosphere is certainly dominated by the wave 1 component (panel e), much of this wave 1 convergence is overlapped by large wave 2 divergence (panel f). These areas, which can be found above the transition region, likely show regions of wave 1 to wave 2 energy transfer. This energy transfer from wave 1 to wave 2 could supply enough wave 2 forcing to produce splits. The wave 2 vertical flux and flux convergence is lower than its wave 1 counterparts, which shows that the split and displacement ratios are not simply results of the relative climatological forcings. Another explanation for the wave 2 structures is that they arise through barotropic instability, which has been shown to induce wave 2 growth in the stratosphere. Hartmann (1983) investigated the barotropic instability of the polar night jet and suggested that the presence of wave 1 forcing may enhance the growth rates of shorter waves of similar phase speeds. Motivated by observations of wave 2 growth confined to the Southern Hemisphere winter stratosphere, Manney et al. (1991) investigated the possibility of barotropic instability as a mechanism for wave 2 growth using a barotropic model as well as a zonally symmetric three-dimensional model. They found that both wave 2 and 3, but in particular wave 2, were destabilized when the basic flow in the barotropic model contained stationary wave 1 and zonal flow. Unstable modes of wave 1, 2 and 3 were found in the three-dimensional model, and the authors noted that the wave 2 modes were usually the most unstable. To verify that barotropic instability could be the reason for the presence of wave 2 structures in these model runs, we calculated the quasi-geostrophic potential vorticity in the 20° N to 90° N region of the stratosphere. A condition for barotropic instability is that the meridional derivative of potential vorticity changes sign in the domain (Charney and Stern, 1962), and this condition is fulfilled for the absolute majority of days in all eight model runs.

The fractions of splits for ~~NE1, NEs1, NE2 and NEs2~~ NW1, NWs1, NW2 and NWs2 are in brackets to signify that the SSWs in these runs do not look like typical splits and displacements, even though our SSW classification sorts these SSWs into one of the two categories. Figure 8 shows the absolute vorticity at 10 hPa and 80 hPa on the central dates of typical SSWs in the eight model runs. Corresponding videos showing 60 days surrounding the SSWs can be found in the supporting information. The 10 hPa level is of interest since that is where SSWs are usually defined, and the 80 hPa level was chosen because it is below the transition region. Panels a and e show a displacement in H1. The displacement of the polar vortex from the pole can clearly be

seen at 10 hPa, while there is little to no suggestion of a displacement at 80 hPa. A clear split in H2 can be seen in panels i and 380 m, and the split of the vortex extends all the way down to the lower stratosphere. In contrast, the 10 hPa levels in **NE1**, **NEs1**, **NE2** and **NEs2** **NW1**, **NWs1**, **NW2** and **NWs2** (panels b, d, j and l) do not show either splits or displacements, even though wave 1 and wave 2 zonal structures can be seen. Instead it seems SSWs are followed by meridionally oriented waves when the 385 effects of zonal **EEIs** **WWIs** are removed. These structures arise when the meridional shear of the flow $(u_\phi) \partial u / \partial \phi$ interacts with zonal wave numbers. Even though these are wave-wave interactions they are allowed to occur, since our method only 390 removes interactions between zonal waves. Shepherd (1987) argued that interactions between stationary and transient waves in the observed atmosphere could be understood to first order through processes like these, where meridional shear transfers enstrophy along lines of constant zonal wave numbers. We have not investigated the effects of the polar meridional waves on the dynamics in these model runs, but we think it is unlikely that they have a significant effect on the overall climatology since they only occur during vortex breakdowns, and not when the vortex is less disturbed.

390 Even though typical splits and displacements do not occur when **EEIs** **WWIs** are not allowed in the middle and upper stratosphere, SSWs in **NE1**, **NEs1**, **NE2** and **NEs2** **NW1**, **NWs1**, **NW2** and **NWs2** are still classified as "splits" and "displacements" 395 since the SSW classification of Lindgren et al. (2018) is based on wave amplitudes of geopotential height. The numbers in brackets in Table 1 therefore show that wave 2 amplitudes completely dominate SSWs when wave 2 forcing is used, with **NE2** and **NEs2** **NW2** and **NWs2** producing 100 % and 90 % wave 2 dominated SSWs ("splits;" statistically significant increases 400 compared to H2). These numbers can be explained by the fact that most of the stratospheric wave flux is of wave 2: the upper stratospheric wave 2 flux in **NE2** **NW2** is an order of magnitude stronger than its wave 1 counterpart (Figs. 6c and d), and although the differences are not as large in **NEs2** **NWs2** wave 2 forcing still dominates (Figs. 6g and h). Just like H1 and **NEt1** **NWt1**, SSWs with wave 1 forcing are also mostly wave 2 dominated, with 57 % and 80 % "splits" for **NE1** and **NEs1** **NW1** and **NWs1**; the difference between the two is statistically significant. As in the case of **NEt1** **NWt1**, this cannot be explained 405 by the dominant wave number of the stratospheric wave flux since wave 1 is the dominant wave number (Figs. 5c, d, g and h). That **NE1** **NW1** produces mostly wave 2 dominated events even though the tropospheric forcing is wave 1 and **EEIs** **WWIs** are not allowed suggests that enhanced growth of wave 2 in the presence of wave 1 forcing through barotropic instability, as hypothesized by Hartmann (1983) and shown by Manney et al. (1991), could be a major factor in the wave 2 structures seen in the wave 1 runs.

405 Figure 8 also shows that **EEIs** **WWIs** are only needed locally for typical splits and displacements to form. First, true splits and displacements do not occur in **NEs1** or **NEs2** **NWs1** or **NWs2** even though the 80 hPa structures look similar to those of the control runs (panels e and m, compared to h and p). As has already been established, the climatological tropospheric and lower stratospheric wave forcing is very similar between these two runs. Second, **NEt1** and **NEt2** **NWt1** and **NWt2** show that 410 splits and displacements do occur when **EEIs** **WWIs** are removed in the troposphere and lower stratosphere, just as long as **EEIs** **WWIs** are allowed above. The structures of the 80 hPa levels in **NE1** and **NEt1** **NW1** and **NWt1** (panels f and g) as well as **NE2** and **NEt2** **NW2** and **NWt2** (panels n and o) are very similar while the 10 hPa levels (panels b versus c, and j versus k) are completely different, with **NEt1** **NWt1** producing a displacement and **NEt2** **NWt2** a split.

The fraction of splits in H2 was 74 %. When the effects of ~~EEIs-WWIs~~ are removed in the troposphere and lower stratosphere (NEt2NWt2) the model only produces splits. While almost all climatological tropospheric F_p in H2 is in wave 2, the small amount of wave 1 that does exist (Fig. 6a) is apparently enough to make about every fourth SSW a displacement. Without ~~EEIs-WWIs~~ in the troposphere and lower stratosphere this low level wave 1 forcing disappears (Fig. 6e). There is almost no climatological wave 1 EP flux convergence in ~~NEt2-NWt2~~ (Fig. 4e), suggesting that there is very little wave 1-mean flow interaction.

5 Discussion and conclusions

In this paper we have investigated the effects of ~~EEIs-WWIs~~ on SSW formation in an idealized GCM, and found that removal of ~~EEIs-WWIs~~ can change the SSW frequency dramatically. While SSWs can be considered wave-mean flow interactions to first order, our results show that removing or adding ~~EEIs-WWIs~~ alter the conditions for SSW generation in non-predictable ways. While removing ~~EEIs-WWIs~~ everywhere and below 50 hPa with wave 2 forcing (~~NE2 and NEt2~~ and ~~NW2 and NWt2~~) does not change the SSW frequency drastically, removal of ~~EEIs-WWIs~~ in the upper stratosphere only (~~NEs2Nws2~~) increases the SSW frequency by 37 % compared to the control run. Since we showed that the wave forcing and climatology of the troposphere and lower stratosphere was dependent on whether or not ~~EEIs-WWIs~~ were allowed in that same region, this 37 % increase can be attributed entirely to changes in nonlinear interactions in the upper stratosphere. The SSW frequencies with wave 1 forcing are strongly dependent on ~~EEIs-WWIs~~: even though H1 and ~~NEs1-Nws1~~ as well as ~~NE1 and NEt1-NW1 and NWt1~~ have almost identical tropospheric and lower stratospheric wave forcings their SSW frequencies are very different, with a 53 % decrease in ~~NEs1-Nws1~~ compared to H1 and a 45 % decrease in ~~NEt1 compared to NE1~~ ~~NWt1 compared to NW1~~. The results from these mixed runs can be contrasted to previous work which has emphasized the importance of tropospheric and lower stratospheric wave flux for SSW generation (Polvani and Waugh, 2004; Birner and Albers, 2017). The results in this paper confirm previously published results showing that the upper stratosphere is not a passive recipient of tropospheric and lower stratospheric wave forcing (Hitchcock and Haynes, 2016), and that stratospheric nonlinear processes are important for SSW generation (O'Neill and Pope, 1988).

While previous authors have investigated the effects of nonlinear interactions in stratospheric dynamics and found them to be important, this work is the first that has explicitly investigated the effects of ~~EEIs-WWIs~~ on SSW frequencies in a global primitive equation model. We find that removal of ~~EEIs-WWIs~~ does not simply increase or decrease the SSW frequency: even with the same tropospheric forcing removal of ~~EEIs-WWIs~~ in the upper stratosphere can decrease SSW frequencies (H1 compared to ~~NEs1Nws1~~) or increase it (~~NE1 compared to NEt1NW1 compared to NWt1~~). To better understand the effects of ~~EEIs-WWIs~~ on the polar vortex we also investigated the variability of polar vortex strength. We find that ~~EEIs-WWIs~~ in the troposphere and lower stratosphere determine much of the long term stratospheric polar vortex variability when wave 1 heating is used. In contrast, ~~EEIs-WWIs~~ in the troposphere and lower stratosphere have a small impact on polar vortex variability when wave 2 heating is used, while middle and upper stratospheric ~~EEIs-WWIs~~ are crucial in determining the long term variability of polar vortex strength. Furthermore, the model runs where ~~EEIs-WWIs~~ are removed in the troposphere and lower stratosphere

but allowed above (~~NEt1 and NEt2~~~~NWt1 and NWt2~~) produce high frequency variability in polar vortex strength, but little long term variability. It seems that much of the wave forcing from below converges in the transition region between ~~no-EEIs and EEIs-allowed~~~~no-WWIs and WWIs-allowed~~, which could be the source of the high frequency variability. The frequent dissipation of wave forcing at lower levels could result in less wave forcing in the middle and upper stratosphere, causing lower 450 amounts of low frequency variability in polar vortex strength.

Some changes caused by removing ~~EEIs-WWIs~~ can be found in all model runs: the stratospheric vertical wave flux and wave flux convergence is further equatorward when ~~EEIs-WWIs~~ are not allowed in the middle and upper stratosphere. This is not a result of a shift in the stratospheric polar vortex, since latitudinal changes in the polar night jet locations are small compared to the changes in wave flux. The equatorward shift is consistent with the results of Robinson (1988).

455 We showed that in the absence of ~~EEIs-WWIs~~, meridionally oriented waves are created in the polar stratosphere, and the vortex is not displaced and does not split. We also found that splits and displacements occur even when ~~EEIs-WWIs~~ are not allowed in the troposphere and lower stratosphere, indicating that only middle and upper stratospheric ~~EEIs-WWIs~~ are necessary for split or displacement formation. These results are in contrast to those of Lordi et al. (1980), who found realistic wave 1 patterns in polar stereographic geopotential height when they used wave 1 forcing without ~~EEIs-WWIs~~. This discrepancy 460 comes from the difference in methods used to remove the effects of ~~EEIs-WWIs~~: many authors, including Lordi et al. (1980), used zonal truncation to only allow one wave number to interact with the mean flow, while we removed all zonal ~~EEIs-WWIs~~. The meridional waves arise from interactions between the meridional shear of the flow and zonal wave numbers. Lordi et al. (1980) therefore found that wave 1-mean flow interactions are enough to create a displacement in the polar region, while we find that removing ~~EEIs-WWIs~~ result in SSWs that are neither splits nor displacements. We do not believe that these meridional 465 waves significantly affect the climatology of the model runs since the waves only occur during vortex breakdowns.

With wave 2 forcing, all SSWs were splits when ~~EEIs-WWIs~~ were turned off in the troposphere and lower stratosphere (~~NEt2NWt2~~). This is likely due to the fact that all tropospheric forcing is in wave 2 when ~~EEIs-WWIs~~ are turned off in the lower levels, and the resulting SSWs are therefore splits. Even though splits and displacements do not occur without ~~EEIs-WWIs~~ in the middle and upper stratosphere the SSWs in these model runs are still dominated by zonal wave 1 or wave 2 470 geopotential height anomalies. SSWs are strongly dominated by wave 2 anomalies in ~~NE2 and NEs2~~~~NW2 and NWs2~~, with 100 % and 90 % wave 2 dominated SSWs ("splits"). As in the case of ~~NEt2NWt2~~, this can be explained by the fact that the stratospheric wave flux is mostly of wave 2 format in these model runs. The fraction of splits increased when ~~EEIs-WWIs~~ were turned off in the lower levels with wave 1 forcing: 80 % splits in ~~NEt1NWt1~~ compared to 59 % in H1. This is despite the fact that there is strong climatological tropospheric wave 1 and wave 2 wave flux in H1, while the flux is almost purely wave 1 in 475 ~~NEt4NWt1~~. The wave 2 forcing required to produce these splits could originate in the transition region between ~~no-EEIs and EEIs-allowed~~~~no-WWIs and WWIs-allowed~~, where some of the wave 1 flux is transferred to wave 2. Barotropic instability is also likely a factor: Hartmann (1983) suggested and Manney et al. (1991) demonstrated that waves of zonal wave number 1 may enhance growth rates of shorter waves. The strong wave 1 forcing in the lower stratosphere of ~~NEt1NWt1~~ could make the flow unstable to wave 2, which would contribute to the large number of splits. SSWs in ~~NE1 and NEs1~~~~NW1 and NWs1~~ are also 480 mostly wave 2 dominated with 59 % and 80 % "splits" even though the stratospheric wave flux is mostly wave 1. The fact that

~~NEI-NW1~~ produces wave 2 dominated events even though the forcing is of wave 1 and ~~EEIs-WWIs~~ are not allowed strongly suggests that barotropic instability could be responsible for the wave 2 structure around SSWs in this model run. The precise nature of interactions between waves 1 and 2 in setting stratospheric variability is the subject of an ongoing investigation.

Code availability. The Flexible Modeling System (FMS) that is used in this paper can be found at <https://www.gfdl.noaa.gov/fms/>.

485 *Data availability.* Zonal mean zonal winds, geopotential height wave amplitudes as well as 10 and 80 hPa absolute vorticity fields are archived at <https://osf.io/pr63b/>.

Video supplement. Movies of the SSWs seen in Fig. 8 are archived at <https://osf.io/3vrme/>.

Author contributions. Erik Lindgren performed the simulations and data analysis, and wrote the initial drafts of the manuscript. Aditi Sheshadri originated the idea for the project, contributed to the interpretation of the results, and improved the final manuscript.

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

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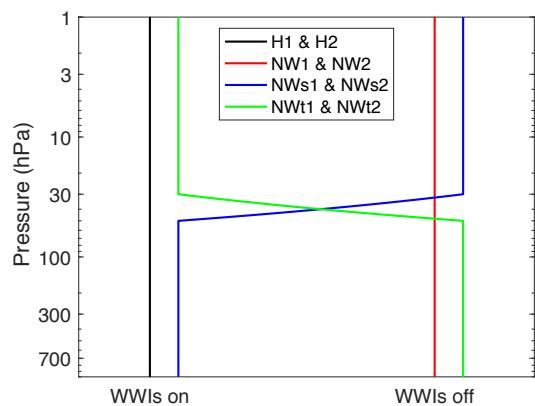


Figure 1. Vertical structure of the four runs used for each wave number (1 or 2). See text for details.

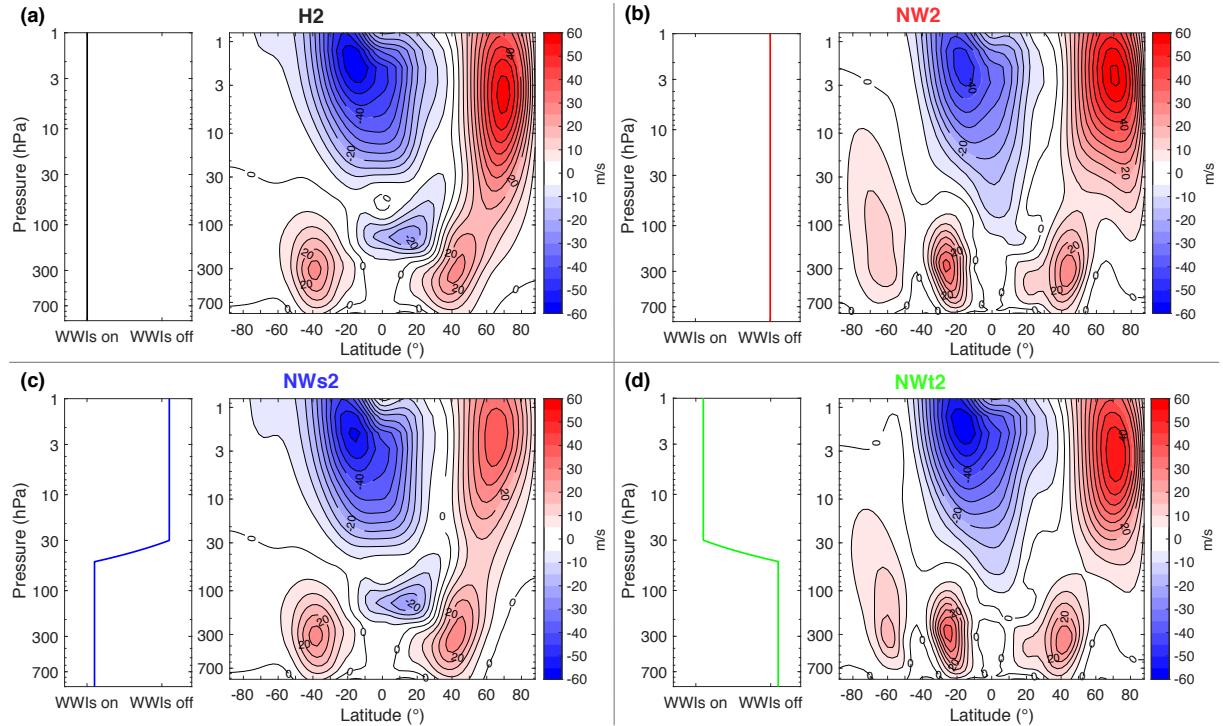


Figure 2. Zonal mean zonal winds for H2 (a), NE2-NW2 (b), NEs2-NWs2 (c) and NEt2-NWt2 (d), with panels showing the pressure levels where EEIs WWIs are allowed. The contour interval is 5 ms^{-1} .

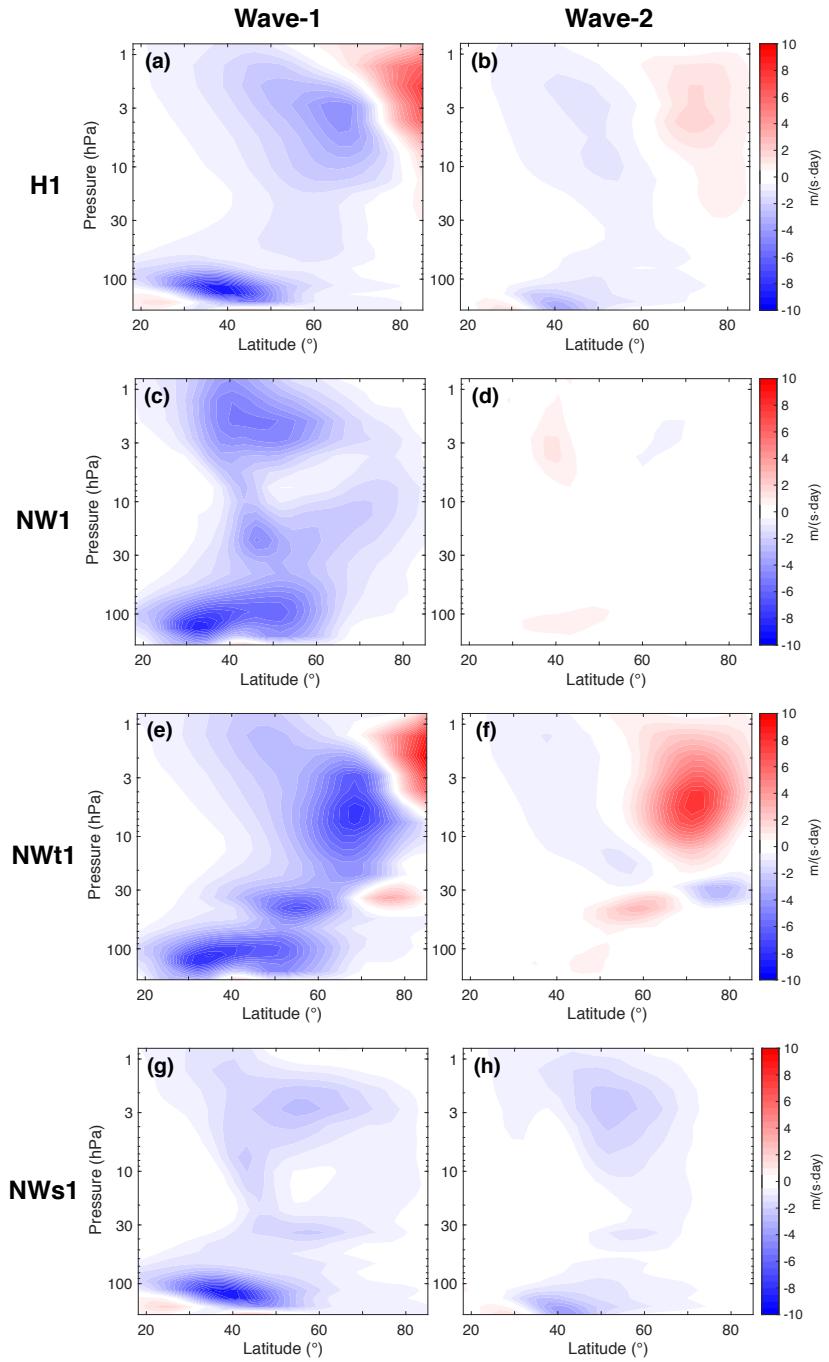


Figure 3. Divergence of EP flux for H1 (a and b), NE1–NW1 (c and d), NEt1–NWt1 (e and f) and NEs1–NWs1 (g and h). The left column shows wave 1 components, right column shows wave 2 components. The contour interval is $0.5 \text{ m}(\text{s}\cdot\text{day})^{-1}$.

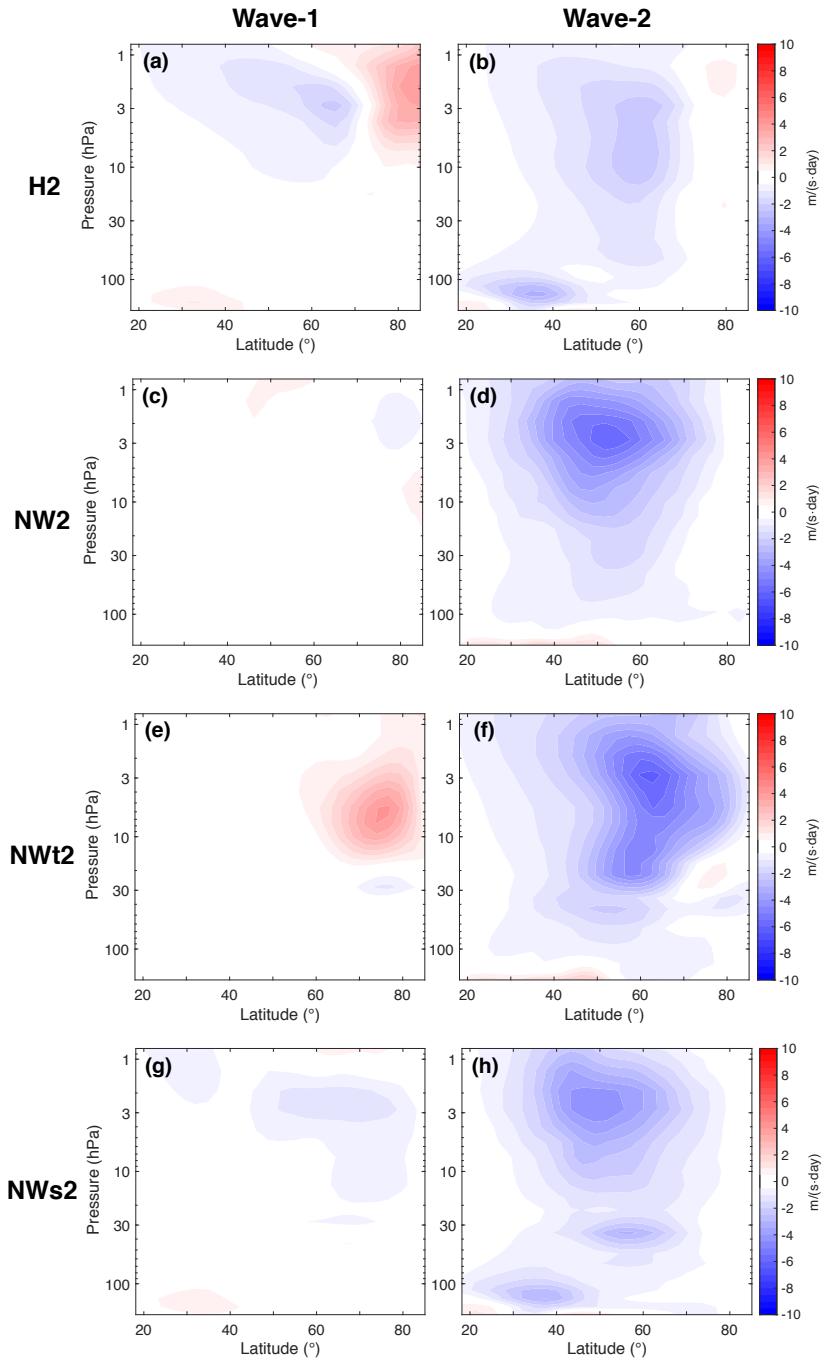


Figure 4. Divergence of EP flux for H2 (a and b), NE2-NW2 (c and d), NEt2-NWt2 (e and f) and NES2-NWs2 (g and h). The left column shows wave 1 components, right column shows wave 2 components. The contour interval is $0.5 \text{ m}(\text{s}\cdot\text{day})^{-1}$.

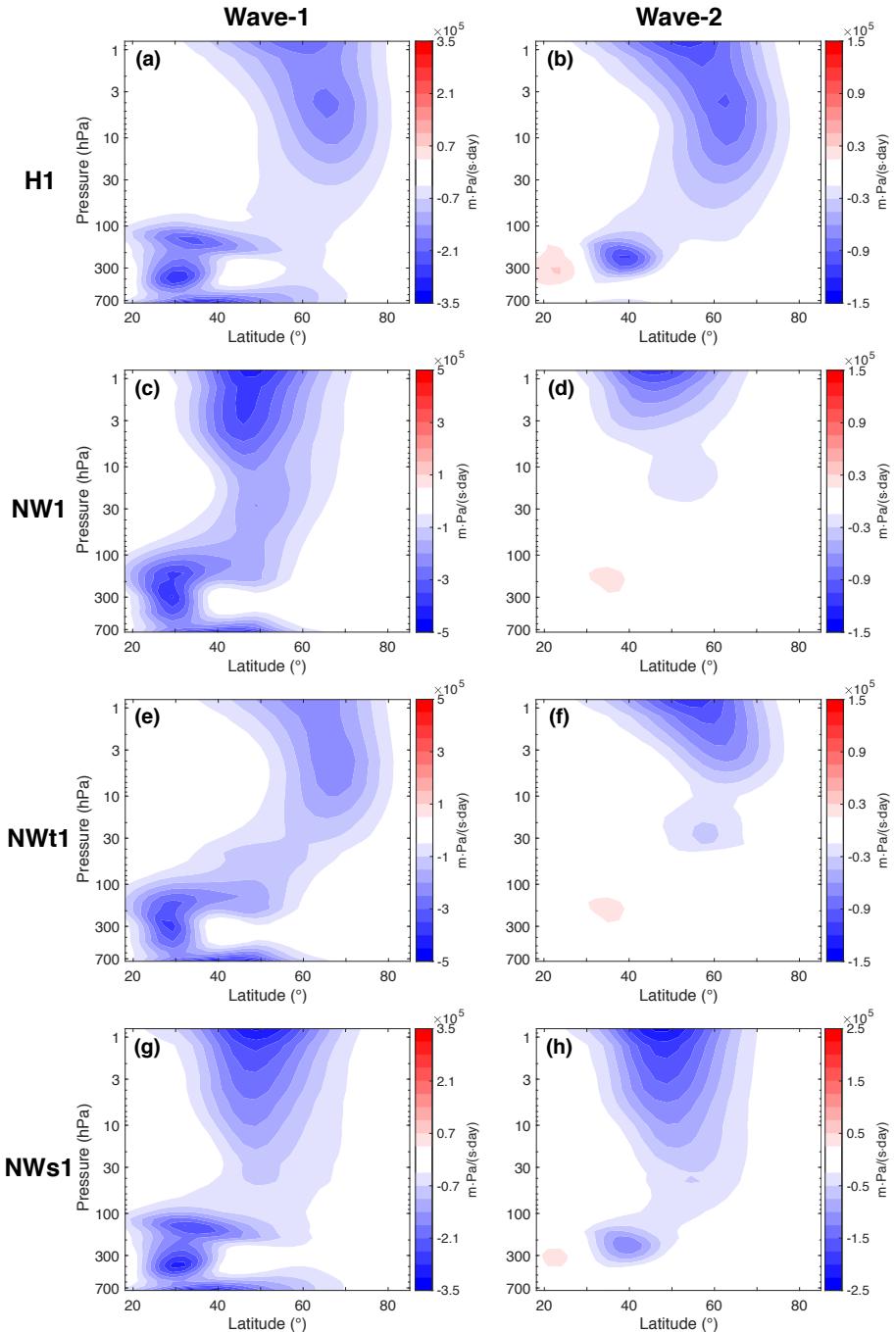


Figure 5. Vertical component of EP flux (F_p) for H1 (a and b), NEt1-NW1 (c and d), NEt1-NWt1 (e and f) and NEs1-NWs1 (g and h). The left column shows wave 1 components, right column shows wave 2 components. Values are scaled by p_0/p , where $p_0 = 1000$ hPa. Contour intervals are 35 (a and g), 15 (b, d and f), 50 (c and e), and 25 (h) $\text{km}\cdot\text{Pa}(\text{s}\cdot\text{day})^{-1}$.

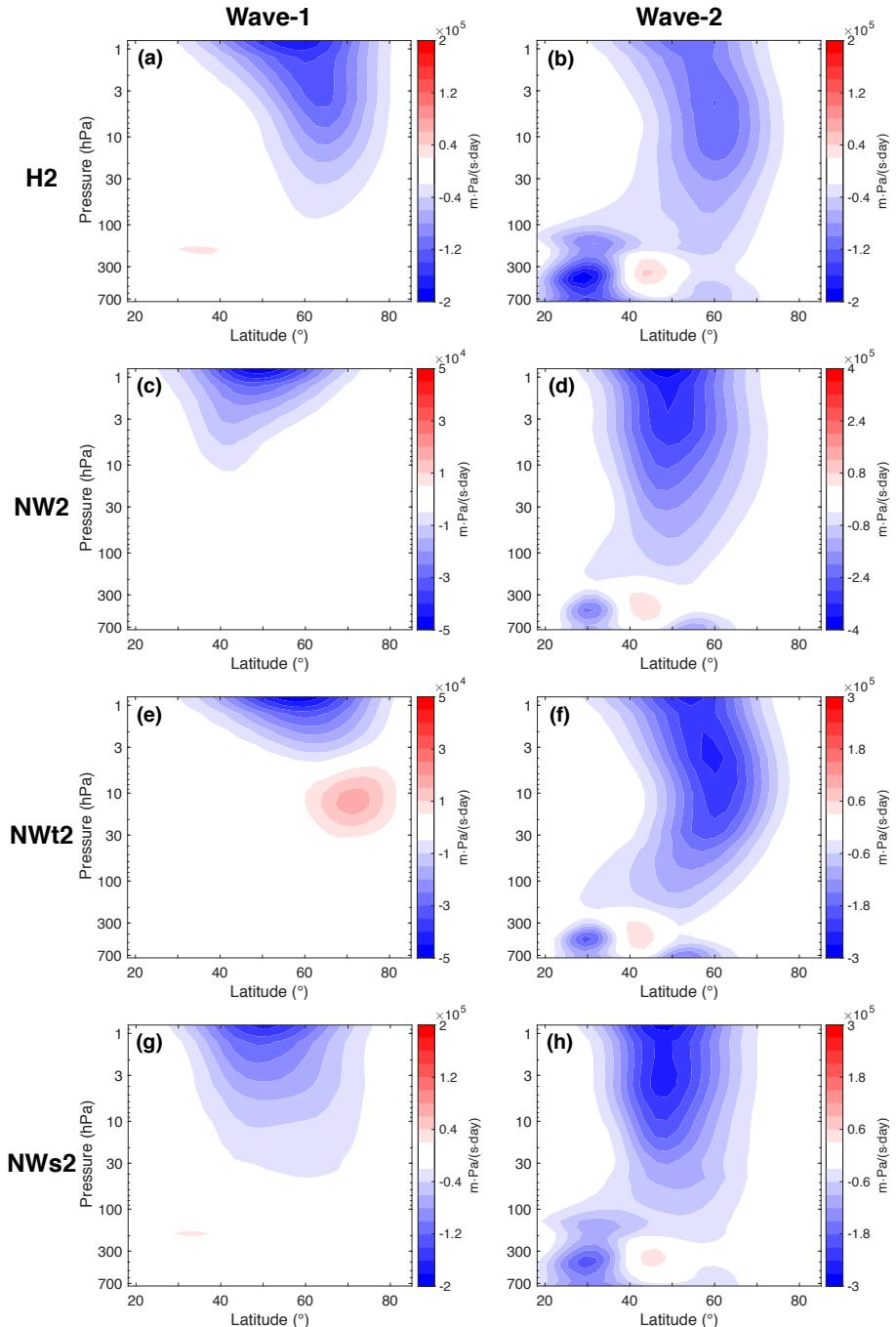


Figure 6. Vertical component of EP flux (F_p) for H2 (a and b), NE2NW2 (c and d), NEt2NWt2 (e and f) and NEs2NWs2 (g and h). The left column shows wave 1 components, right column shows wave 2 components. Values are scaled by p_0/p , where $p_0 = 1000$ hPa. Contour intervals are 20 (a, b and g), 5 (c and e), 40 (d), and 30 (f and h) $\text{km}\cdot\text{Pa}(\text{s}\cdot\text{day})^{-1}$.

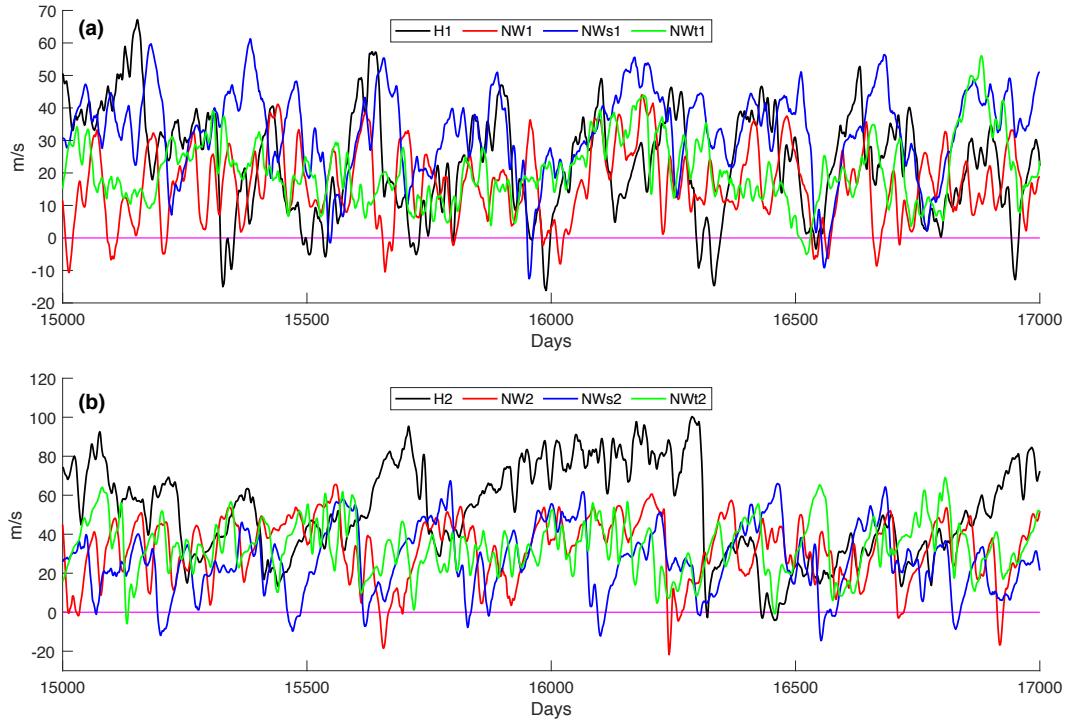


Figure 7. Zonal mean zonal wind at 10 hPa and 60° N for H1/H2 (black), NE1NW1/NE2-NW2 (red), NEs1NWs1/NEs2-NWs2 (blue) and NEt1NWt1/NEt2-NWt2 (green) for model runs with wave 1 (a) and wave 2 (b) heating. The data has been smoothed with a 10 day filter. The magenta line marks 0 ms^{-1} .

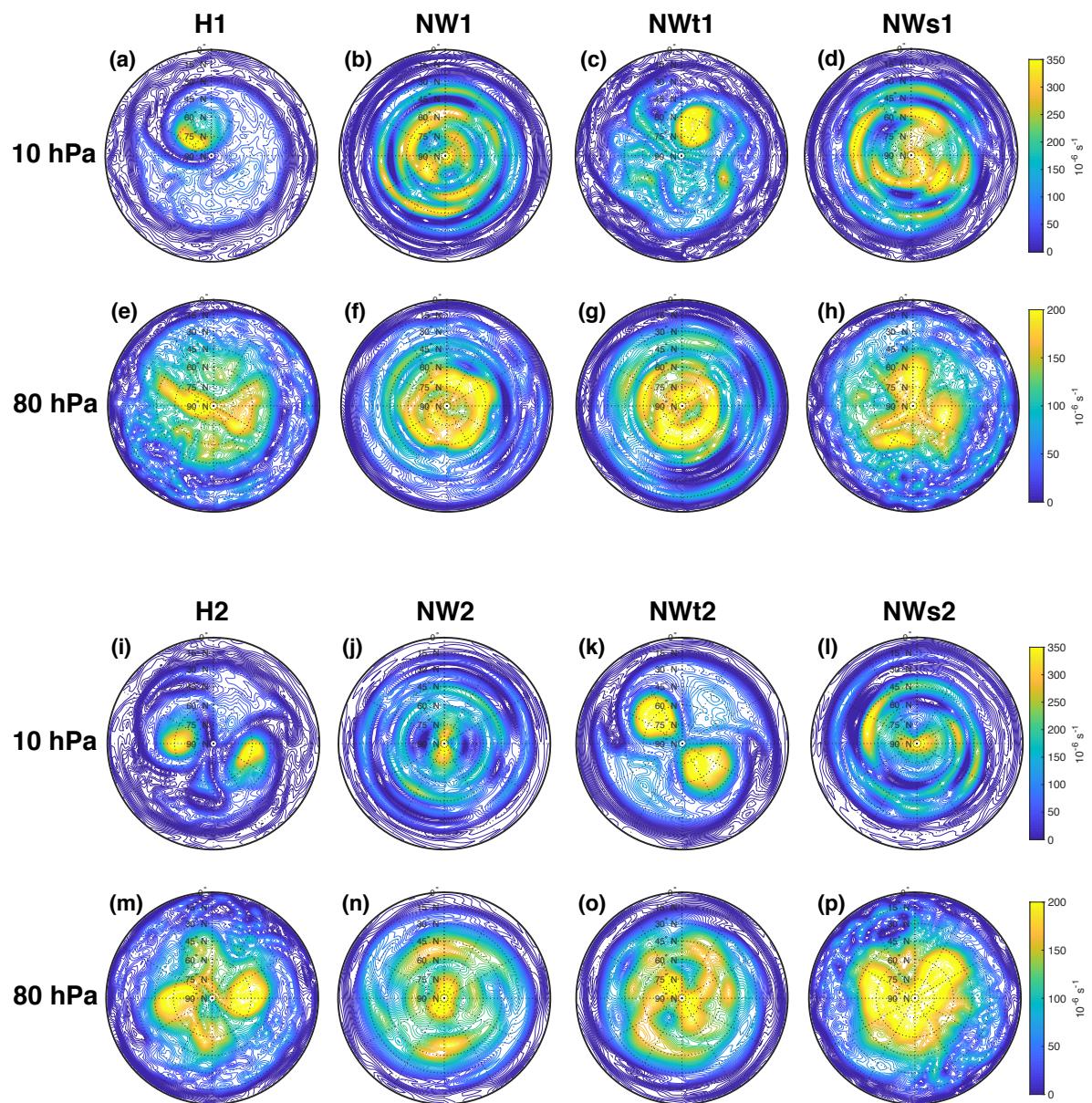


Figure 8. Absolute vorticity at 10 hPa and 80 hPa on the central dates of an SSW with wave 1 heating (top two rows) and wave 2 heating (bottom two rows). Displacements can be seen in (a) and (c), while (i) and (k) show splits.

Table 1. SSW frequencies, time mean and variability of \bar{u} at 10 hPa and 60° N, and classifications for the model runs. Split and displacement numbers are obtained with a wave amplitude classification; numbers in brackets indicate that the SSWs do not look like typical splits or displacements.

Model run	H1	NE1 NW1	NEt1 NWt1	NEs1 NWs1
Total SSWs (SSWs per 100 days)	199 (0.66)	247 (0.82)	132 (0.44)	93 (0.31)
\bar{u}_{1060} , mean (standard deviation) [ms $^{-1}$]	23 (16)	17 (12)	21 (10)	30 (14)
Total splits (fraction)	118 (0.59)	[143 (0.57)]	105 (0.80)	[74 (0.80)]
Model run	H2	NE2 NW2	NEt2 NWt2	NEs2 NWs2
Total SSWs (SSWs per 100 days)	145 (0.48)	153 (0.51)	134 (0.45)	199 (0.66)
\bar{u}_{1060} , mean (standard deviation) [ms $^{-1}$]	39 (27)	33 (19)	32 (16)	26 (19)
Total splits (fraction)	108 (0.74)	[153 (1)]	134 (1)	[179 (0.90)]