Influence of Eddy-Eddy Interactions and Tropical Wind Variability on Sudden Stratospheric Warming Formation

by

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B.S., Physics, University of New Hampshire (2013)

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Abstract

This thesis investigates the effects of eddy-eddy interactions (EEI) and tropical wind variability on sudden stratospheric warming (SSW) formation in an idealized atmospheric GCM. Chapter 2 introduces a method to produce split and displacement SSWs in comparable amounts using either wavenumber 1 or 2 tropospheric heating perturbations. The results are compared to those obtained with wavenumber 2 topographic forcing. It is shown that the fraction of SSWs forced by anomalously strong tropospheric wave flux in the model is similar to that of SSWs in the observed atmosphere, but that the fractions for splits and displacements are different. Furthermore, a large fraction of SSWs occur without significant anomalous tropospheric wave flux, indicating that stratospheric transmission of climatological tropospheric wave flux plays an important role in SSW formation.

Chapter 3 investigates the effects of EEI on SSW formation in the model by reproducing the model runs from Chapter 2 with EEI turned off in parts of the atmosphere. It is found that SSW frequencies can be strongly dependent on EEI throughout the atmosphere, but that EEI are required locally for splits and displacements to occur. Significant changes in SSW frequencies are obtained by turning off EEI locally, without changing the lower stratospheric wave forcing. Chapter 3 shows that while SSW formation can be considered a wave-mean flow interaction to first order, higher order processes are required to accurately reproduce both SSW frequencies and dynamics.

The wavenumber 2 heating run used in Chapters 2 and 3 produce spontaneous tropical wind oscillations in the stratosphere. Chapter 4 identifies the source of these oscillations, and investigates the effects of the oscillations on the stratospheric polar vortex. Model runs with suppressed tropical wind variability are compared to the control run of Chapter 2. A slight increase in SSW frequency can be found in the model runs with suppressed tropical variability. It is found that upper stratospheric equatorial wind anomalies are strongly correlated with polar vortex strength, and hypothesized
that westerly equatorial wind anomalies in the upper stratosphere can reinforce the conditions that lead to an anomalously strong polar vortex. A mechanism explaining this influence is presented.

Thesis Supervisor: R. Alan Plumb
Title: Professor Emeritus
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Chapter 1

Introduction

The atmosphere is generally divided into four distinct layers: the troposphere, stratosphere, mesosphere and thermosphere. The troposphere is the lowermost layer and contains about 85% of the atmospheric mass, and almost all atmospheric water vapor. The troposphere is well mixed due to convection and baroclinic eddies, and its temperature decreases with height. The highest levels that convective parcels generally reach mark the boundary between the troposphere and the stratosphere, and is called the tropopause. The height of the tropopause varies with latitude, from about 8 km in the cold polar regions to about 16 km in the tropics where convection is strong. The stratosphere contains roughly 15% of the atmospheric mass, and extends from the tropopause to the stratopause, located at a height of about 50 km. As the name suggests, the stratosphere is highly stratified and strongly inhibits convection. Unlike the troposphere, the temperature of the stratosphere increases with height due to absorption of ultraviolet solar radiation by stratospheric ozone. The mesosphere is located above the stratopause, and its temperature decreases with height. Like the stratosphere, the thermosphere is characterized by a temperature increasing with height, due to absorption of solar radiation connected to dissociation of nitrogen and oxygen molecules. Figure 1-1 shows the U.S. Standard Atmosphere (1976) temperature profile at midlatitudes for the troposphere, stratosphere and mesosphere. Data
This thesis is in the field of stratospheric dynamics, and the focus of the thesis is how eddy-eddy interactions and tropical wind oscillations influence the generation of sudden stratospheric warmings. In this chapter I describe the basic climatological structure of the winter stratosphere, what sudden stratospheric warmings are and how they form, and the mechanisms behind and effects of tropical wind oscillations.

1.1 Zonal mean climatology of the stratosphere

The structure of the stratosphere is largely determined by solar radiation. At solstice the summer hemisphere is strongly heated by solar radiation, while very little sunlight reaches the winter hemisphere polar region. This cools the winter polar region of the stratosphere, and creates a strong equator to pole temperature gradient. The temperature gradient induces a strong westerly circulation around the winter pole through
Figure 1-2: Zonal mean zonal winds (top) and temperatures (bottom) for July (left) and January (right). The means are calculated over the years 1979-1995 (zonal wind) and 1979-1998 (temperature). The black lines show the level of the tropopause. The panels come from NASA (2001).

thermal wind balance. This strong westerly circulation is known as the stratospheric polar vortex, or the polar night jet. The tropospheric latitudinal temperature structure and polar vortices are not as strongly influenced by the seasonal progression. This is due to heat transport from the equator to the poles by midlatitude atmospheric eddies and the ocean circulation, and strong reflection of solar radiation in the summer polar region due to the high albedo of ice and snow. The zonal mean zonal wind and temperature structure of the troposphere and stratosphere during Southern and Northern Hemisphere winter can be seen in Figure 1-2.

The most striking differences between the Southern and Northern Hemisphere winters are the differences in stratospheric polar temperature and polar vortex strength. The
Southern Hemisphere winter stratospheric polar region is on average about 20 K colder than its Northern Hemisphere counterpart. The mean January polar vortex strength barely exceeds 40 m/s, while the mean July polar vortex strength is about twice as strong. The reason for these discrepancies is the difference in planetary-scale wave activity between the two hemispheres. These waves are primarily formed in the troposphere from where they propagate upward, but Charney and Drazin (1961) showed that wave propagation is dependent on the mean flow as well as the phase speed and wavenumber of the wave. Their results show that stationary waves can only propagate into the stratosphere if the mean flow is westerly and if the zonal wavenumbers of the waves are sufficiently small. The consequence of this is that the majority of upward propagating wave activity in the stratosphere occurs during hemispheric winter (when the flow is westerly) and is of zonal wavenumbers 1 or 2. The large-scale topography (primarily the Himalayas and the Rocky Mountains) and land-sea heating contrasts (the continents and the Atlantic Ocean and Pacific Ocean) in the Northern Hemisphere produce frequent, high-amplitude Rossby waves of zonal wavenumbers 1 and 2, which propagate upward into the stratosphere during Northern Hemisphere winter. The waves grow in amplitude with decreasing pressure, until their amplitudes become too large and the waves break. The breaking waves deposit negative momentum and act to decelerate the mean flow. When this occurs in the stratosphere the result is a weakening of the stratospheric polar vortex and a temperature increase in the polar region. In comparison to the Northern Hemisphere, the Southern Hemisphere has little large-scale topography and a weak land-sea heating contrast due to the smaller sizes of the Southern Hemisphere continents, which results in less planetary-scale wave activity and a less disturbed stratospheric polar vortex.

1.2 Sudden stratospheric warmings

When the stratospheric wave breaking is frequent enough or strong enough the induced deceleration can reverse the direction of the polar vortex. Such extreme events
are known as sudden stratospheric warmings (SSWs), and they are usually defined as events where the winter zonal mean zonal wind at 10 hPa and 60° becomes easterly (Charlton and Polvani, 2007). The day when this first occurs is referred to as the central date. The observed frequency of SSWs is about 0.6 per year in the Northern Hemisphere. Only one SSW has been observed in the Southern Hemisphere.

The reason for the name is apparent when one looks at Figure 1-3, from NOAA CPC (2018). The figure shows the Northern Hemisphere polar cap temperature at 10 hPa from 1999 through 2000, together with the climatological temperature and the temperature variability. SSWs occurred on February 26 1999 and March 20 2000, and they are associated with temperature increases of 40 K and 20 K, respectively, over the course of one to two weeks. The weakening and distortion of the polar vortex that occurs during a SSW is accompanied by a subsidence of air in the polar region, and as the air sinks it heats up adiabatically. The polar cap temperature following
the first SSW reached a temperature that exceeds the maximum temperature during most summers. Notice that a sharp temperature increase also took place from the beginning of January 2000 until early February 2000. This was not classified as a SSW based on the definition above: while the polar cap region warmed significantly the zonal mean zonal wind at 10 hPa and 60°N did not become easterly. There is no universally agreed upon definition of a SSW, and the frequency of SSWs is somewhat sensitive to the definition used (Butler et al., 2015). However, the Charlton and Polvani (2007) definition is the one most commonly used and it will form the basis for SSW definitions throughout this thesis.

Figure 1-3 also shows that the Northern Hemisphere polar cap temperature variability has a strong seasonal variation: the measured winter temperatures for a given day vary by 40-50 K, compared to about 10 K for the summer temperatures. This is due to the fact that wave propagation into the stratosphere is prohibited during summer, due to the prevailing easterly winds. The stratospheric climate is heavily dependent on wave forcing, and this is clear to see in Figure 1-3.

SSWs are typically thought of as two types of events: vortex displacements (wavenumber 1 events) and vortex splits (wavenumber 2 events). Figure 1-4 shows the absolute vorticity over the pole at 10 hPa on the central dates of the two SSWs seen in Figure 1-3. The 1999 SSW was a split, while the 2000 SSW was a displacement. As the names suggest, the polar vortex tends to split into two more or less equally large vortices during splits, while the vortex is displaced away from the pole while shedding filaments of high vorticity during displacements. Like the case of SSWs themselves, no universally agreed upon separation between splits and displacements exists. SSWs in the observed atmosphere are roughly equally distributed between splits and displacements, and some authors have found the two types to be dynamically distinct (Charlton and Polvani, 2007). Some studies have found that splits and displacements have different impacts at the surface (e.g., Mitchell et al., 2013; Seviour et al., 2013), while other authors argue that differences in surface impacts following splits and displacements are not consistent for different classifications, and that a large number
Figure 1-4: Absolute vorticity at 10 hPa on the central dates of the February 26 1999 (left; split) and March 20 2000 (right; displacement) SSWs seen in Figure 1-3, calculated from ERA-Interim reanalysis data (Dee et al., 2011). The highest values of absolute vorticity exceeded the range of the colorbar, but the lower range was chosen to highlight the structure of the polar vortices.

of events is required to distinguish the different responses (Maycock and Hitchcock, 2015).

SSWs are the dominant sources of variability in the Northern Hemisphere winter stratosphere, but exactly how SSWs are generated is not fully understood. SSWs have traditionally been thought to be caused by the breaking of unusually large upward propagating planetary-scale waves originating in the troposphere (Matsuno, 1971), but more recent research has indicated that the state of the stratosphere plays a crucial role in SSW formation. Birner and Albers (2017) showed that SSWs and SSW-like events found in reanalysis data often occur without any preceding strongly anomalous tropospheric wave activity, and that most anomalous tropospheric wave events have very small impacts on stratospheric circulation. The authors claimed that the climatological wave activity propagating from the troposphere is enough to cause SSWs, and that whether SSWs happen or not depends on the state of the stratosphere. That SSWs can occur without anomalous tropospheric wave activity has also been found in idealized GCMs with suppressed tropospheric variability (Scott and Polvani, 2004, 2006). Furthermore, Hitchcock and Haynes (2016) found that the evolution of the stratospheric mean state during SSWs is crucial in determining the upward fluxes.
of wave activity during the warming, and that wave amplification during SSW onset was reduced by about 50% when variations in the stratospheric zonal mean state were suppressed in an idealized GCM.

Idealized atmospheric GCMs provide an inexpensive way to investigate the dynamics of the stratosphere, but SSWs produced in idealized GCMs are often heavily biased towards either splits or displacements. In Chapter 2 of this thesis I present a method to produce splits and displacements in comparable amounts using tropospheric forcing in an idealized atmospheric GCM. The results of two model setups that produce large amounts of both splits and displacements are compared to that of a setup that produces mostly splits. The results of the model runs are used to investigate the extent to which SSWs are forced by anomalous tropospheric forcing. The effects of splits and displacements on the polar region are also compared.

The tropospheric forcings used to produce splits and displacements in Chapter 2 are of zonal wavenumber 1 and 2, respectively. Since displacements and splits are wavenumber 1 and wavenumber 2 events and either of the forcings produce both types of events, some of the wave activity propagating from the troposphere to the stratosphere must be transferred from the wavenumber of the forcing (1 or 2) to the other main stratospheric wavenumber (2 or 1). The role of eddy-eddy (wave-wave) interactions (EEI) in SSW formation is investigated in detail in Chapter 3 by turning off EEI in the atmospheric GCM used in Chapter 2. Particular emphasis is placed on the importance of EEI for SSW frequencies and split and displacement ratios. Model runs where EEI are allowed at some pressure levels but removed at others are also analyzed, to determine if there are pressure levels where EEI are particularly important for SSW formation.
1.3 Tropical wind oscillations in the observed atmosphere

The quasi-biennial oscillation (QBO) is the major source of variability in the lower tropical stratosphere, while the semiannual oscillation (SAO) is behind most of the tropical variability above (Gray, 2013). The QBO consists of equatorial zonal mean zonal wind reversals around the equator from around 5 hPa to about 100 hPa, with amplitudes of about 15 m/s. The vertical extent of the SAO ranges from about 5 hPa to the lower mesosphere above 0.1 hPa, and its amplitude is about 10 m/s. The names indicate the periods of the oscillations, with a period of about 28 months for QBO and 6 months for the SAO. While the SAO is confined to given pressure levels, the QBO propagates downward with time. Figure 1-5 shows the zonal mean zonal wind at the equator from 1980 through 2017. The SAO can be glimpsed at the upper pressure levels, and the downward propagation of the QBO is clearly seen.

The mechanisms behind the QBO have been extensively studied (Baldwin et al., 2001, and references therein), and they can be summarized as two-way interactions between vertically propagating equatorial waves and the mean flow. The propagation of waves depends on the strength and direction of the mean flow and the phase speed of the waves, but the propagating waves converge in the stratosphere and deposit momentum, thereby altering (and eventually reversing) the mean flow. The convergence of the waves changes the flow such that the region of maximum wave convergence descends with time, which is where the downward-propagating pattern seen in Figure 1-5 comes from. When the mean flow has reversed, waves of the opposite phase speed will be permitted to propagate into the stratosphere and interact with the mean flow, thereby eventually reversing the mean flow again. The long period of the oscillation that is induced can only occur in the tropics, since wave forcing will be balanced by the Coriolis force at higher latitudes. This mechanism was described schematically in Plumb (1984).
Figure 1-5: Monthly mean zonal mean zonal wind at the equator from 1980 through 2017. The dashed black lines show the level of the tropopause. The figure comes from NASA (2018), and the values have been obtained from MERRA-2 reanalysis data.

There are a wide variety of waves responsible for the QBO, of three categories (Baldwin et al., 2001; Gray, 2013): Kelvin and Rossby-gravity waves that are confined to the equatorial region; inertia-gravity waves, which are not always trapped in the equatorial region; and gravity waves, which can occur at all latitudes. The SAO is forced by equatorial waves whose group velocities or phase speeds are fast enough to propagate through the QBO winds, but is also likely modulated by planetary waves and advection (Gray, 2013).

Many GCMs do not reproduce the equatorial wind oscillations observed in the atmosphere. Of the 13 stratosphere resolving models in the Coupled Model Intercomparison Project Phase 5 (CMIP-5) only 5 produce spontaneous QBOs (Mitchell et al., 2015). Dry dynamical core GCMs similar to the one used in this thesis sometimes produce QBO-like oscillations, although their periods of oscillation rarely match those of the observed QBO: Yao and Jablonowski (2015) produced QBO-like oscillations in three idealized GCMs with periods of 3.6, 13 and longer than 13 years.
Despite being a tropical phenomenon, the QBO has been found to affect extratropical and midlatitude circulation. Holton and Tan (1980) were the first to show that the QBO influenced Northern Hemisphere winter circulation, and hypothesized that the QBO changed the location of the zero mean zonal wind line, thereby altering the propagation of planetary-scale waves into the stratosphere. The effect of the QBO on high latitudes has been the subject of many studies since then, using both observations and modeling (Anstey and Shepherd, 2014, and references therein). While the mechanism suggested by Holton and Tan (1980) exerts its influence in the lower stratosphere, other authors have suggested that higher level tropical wind variability may also influence the extratropics and midlatitudes. Gray et al. (2001b) found a strong correlation between upper stratospheric (~ 1 hPa) equatorial winds and polar temperature in rocketsonde data during January and February, with anomalously strong upper stratospheric equatorial winds coinciding with a cold and strong Northern Hemisphere polar vortex. Gray et al. (2001a) reproduced this correlation in an idealized stratosphere-mesosphere model, where the winds were relaxed to the observed rocketsonde data. They found that the equatorial wind structure for the full stratosphere and lower mesosphere (16-58 km) was required to reproduce the influence of the QBO on polar vortex strength, and that using only lower or upper stratospheric wind anomalies was not enough to capture the Holton and Tan (1980) relationship seen in observations. Gray (2003) followed up on this result by imposing anomalous equatorial zonal winds in the upper and lower stratosphere of the same model, and found that North Pole temperatures were rather insensitive to lower stratospheric wind variations, while anomalously westerly upper stratospheric equatorial winds led to a much less disturbed vortex and fewer SSWs. If the upper stratospheric equatorial winds do exert an influence on stratospheric winter circulation, there are several phenomena that can influence the circulation through this mechanism. In addition to the QBO and SAO the solar cycle also exerts a strong influence on the upper stratospheric winds and temperature. Some authors have found evidence for a modulation of stratospheric circulation by the solar cycle, where high solar activity leads to an altered meridional temperature gradient in the upper stratosphere, thereby changing
wave convergence and meridional circulation in the winter hemisphere (Kodera and Kuroda, 2002; Kodera and Shibata, 2006).

One of the model runs used in Chapters 2 and 3 produces equatorial wind oscillations with middle stratospheric magnitudes similar to that of the QBO (10-15 m/s), and with even larger magnitudes in the upper stratosphere (up to 30 m/s). Unlike the QBO these oscillations propagate upward with time, and the timescale of the oscillations range from a few hundred to over a thousand days. Furthermore, periodic oscillatory behavior was observed in the autocorrelation function (ACF) of Northern Hemisphere winter polar vortex strength in the model run. Since the model was run under Northern Hemisphere winter conditions without any imposed time-dependency, these oscillations must be generated internally in the model. Chapter 4 of this thesis investigates the connection between these spontaneous equatorial wind oscillations and midlatitude winds in the model used in Chapters 2 and 3. An extended Northern Hemisphere winter run is analyzed, along with two Northern Hemisphere winter runs of equal length but with equatorial wind damping applied. Model runs with a seasonal cycle are also performed, both with and without equatorial wind damping. The chapter identifies the mechanisms that cause the tropical wind oscillations, and finds a weak influence on midlatitude winds by upper stratospheric equatorial wind oscillations in the Northern Hemisphere winter run. A mechanism through which this influence is exerted is proposed.
Chapter 2

Sudden stratospheric warming formation in an idealized GCM using three types of tropospheric forcing

2.1 Introduction

In order to obtain a Northern Hemisphere winter-like stratosphere in an atmospheric GCM one needs to create planetary-scale waves in the model. Stationary planetary waves can be forced by large-scale topography (Charney and Eliassen, 1949), land-sea heating contrasts (Smagorinsky, 1953), and the nonlinear interactions of synoptic scale eddies (Scinocca and Haynes, 1998). A way to produce Northern Hemisphere-like stratospheric wave activity commonly used in the literature is to use topography as a “wave maker” (e.g., Taguchi and Yoden, 2002; Gerber and Polvani, 2009; Sheshadri et al., 2015). Topography provides a simple way to tune the wavenumber and amplitude of tropospheric planetary-scale waves, and the regimes of stratospheric vari-

\[1\]This chapter is based on Lindgren, E. A., A. Sheshadri and R. A. Plumb: Sudden stratospheric warming formation in an idealized GCM using three types of tropospheric forcing, submitted for publication in J. Geophys. Res. Atmos.
ability that result from the chosen forcing can then be explored. Gerber and Polvani (2009) added topography to the idealized GCM used by Polvani and Kushner (2002), and found that SSW frequencies similar to those of the observed Northern Hemisphere could be produced when running the model under perpetual winter conditions and with wavenumber 2 (wave-2) topography. These SSWs were all vortex splits. Gerber and Polvani (2009) found that the most Northern Hemisphere winter-like behavior occurred with wave-2 topography and amplitudes ranging from 2500 m to 3500 m. They also found that with increasing amplitude of wavenumber 1 (wave-1) forcing the polar vortex went from being strong to destroyed, with no SSW regime in between. Other authors have produced SSWs in models with wave-1 topographic forcing, both under perpetual winter conditions (Taguchi et al., 2001) and with a seasonal cycle (Taguchi and Yoden, 2002). Martineau et al. (2018) varied the stratospheric temperature profile in the same dry dynamical core model used by Polvani and Kushner (2002) and Gerber and Polvani (2009) and found that SSWs were produced with both wave-1 and wave-2 topographic forcing.

Sheshadri et al. (2015) used a model based on that of Gerber and Polvani (2009), and investigated its behavior when run with a seasonal cycle and different types of topographic forcings. An important difference between the Gerber and Polvani (2009) and Sheshadri et al. (2015) models was the transition between tropospheric and stratospheric equilibrium temperatures: the transition occurred at 200 hPa for Sheshadri et al. (2015) compared to 100 hPa for Gerber and Polvani (2009). Sheshadri et al. (2015) found that 4000 m wave-2 topography yielded a SSW frequency of 0.62 SSWs/year: a very close match to the 0.61 SSWs/year they found in reanalysis data for the years 1958 through 2013. Like Gerber and Polvani (2009), they also found that the wave-2 topographic forcing only produced split SSWs. Unlike Gerber and Polvani (2009), Sheshadri et al. (2015) found that wave-1 forcing produced both splits and displacements, although they were unable to find an amplitude of wave-1 forcing that produced a realistic SSW frequency. The highest SSW frequency they obtained with wave-1 forcing was 0.1 SSWs/year. Martineau et al. (2018) also found that SSWs
were less common with wave-1 topographic forcing compared to wave-2 forcing. They obtained splits and displacements with both wave-1 and wave-2 topographic forcing, although Martineau et al. (2018) also found that wave-1 forcing favored displacements while almost all SSWs with wave-2 forcing were splits.

Scott and Polvani used idealized models to argue that SSWs and similar extreme stratospheric events can be caused by the state of the stratosphere, without anomalous tropospheric forcing. They used time independent wave-1 tropospheric heating sources (Scott and Polvani, 2004) as well as wave-1 and wave-2 geopotential perturbations (Scott and Polvani, 2006) to create SSW-like events in an idealized atmospheric GCM where the tropospheric variability was suppressed. While their results show that SSWs can occur as long as climatological tropospheric forcing of wavenumber 1 or 2 exists, the results of Gerber and Polvani (2009) and Sheshadri et al. (2015) indicate that both the frequency and type of SSWs can be heavily dependent on the wavenumber of the tropospheric forcing.

In this chapter I investigate the extent to which SSWs in this model are forced by anomalous tropospheric wave fluxes, and if the results are sensitive to the forcing used or the type of SSW produced. The results are compared to the observational results of Birner and Albers (2017). I begin by presenting a method to simulate Northern Hemisphere winter-like stratospheric variability with diabatic heating instead of topography. I run an adapted version of the model used by Gerber and Polvani (2009) and Sheshadri et al. (2015) with tropospheric heating perturbations as a substitute for topography. The model is run under perpetual Northern Hemisphere winter conditions with both wave-1 and wave-2 heating perturbations, and the results are compared to those of a perpetual winter version of the setup used by Sheshadri et al. (2015). I use two different split and displacement classifications to show that, unlike in the topographically forced run, both wave-1 and wave-2 tropospheric heating produce large numbers of both splits and displacements. The SSW classifications are also tested on ERA-Interim reanalysis data for validation purposes. I analyze the anomalous wave fluxes associated with SSWs by separating the vertical
Eliassen-Palm (EP) flux around splits and displacements into wavenumber components. I find that SSWs in this model form both as a direct result of anomalous tropospheric wave activity and due to internal stratospheric processes which alter the propagation of tropospheric wave flux into the stratosphere, and that the processes by which splits and displacements are formed are different when wave-1 heating is used. I also find that the fraction of SSWs forced by tropospheric wave flux anomalies exceeding two standard deviations is comparable to the fraction found in observations. Furthermore, I analyze the polar cap geopotential height anomalies around splits and displacements, and find that splits and displacements have different surface signatures when the model is forced by wave-1 tropospheric heating.

The model setup as well as SSW definition and classifications are described in section 2.2. In section 2.3 I compare the climatologies of the model runs with wave-1 and wave-2 heating perturbations to that of the perpetual winter version of the Sheshadri et al. (2015) setup. Section 2.4 details the SSW statistics of the runs and the ERA-Interim data, while in section 2.5 I look at the vertical wave-1 and wave-2 EP flux anomalies around SSWs. The analysis of geopotential height anomalies associated with splits and displacements can be found in section 2.6. I present my conclusions in section 2.7.

### 2.2 Methods

Both model results and reanalysis data are analyzed in this chapter. The reanalysis data comes from the ERA-Interim Project (Dee et al., 2011) and contains four times daily observations from 1 January 1979 through 31 October 2016. Daily mean values of relative vorticity, zonal mean wind and geopotential height were used. The Model setup subsection below describes the simplified atmospheric GCM that was used to model the Northern Hemisphere winter stratosphere. The methods used to identify and classify SSWs are outlined in the second part of this section.
2.2.1 Model setup

The model is an adapted version of the one used by Polvani and Kushner (2002). It is dry and hydrostatic, with T42 resolution in the horizontal and 40 vertical \( \sigma \) levels, where \( \sigma = p/p_s \) and \( p_s \) is surface pressure. Newtonian relaxation to a zonally symmetric equilibrium temperature replaces radiation and convection schemes. Following Sheshadri et al. (2015), the transition between stratospheric and tropospheric conditions occurs at 200 hPa. The equations for the equilibrium temperature can be found in Polvani and Kushner (2002). \( c \) is an asymmetry parameter in the equation for tropospheric equilibrium temperature, and unlike Polvani and Kushner (2002) and Sheshadri et al. (2015) I chose a value of \( c = 0 \) K to keep the tropospheric equilibrium temperature symmetric about the equator. Polvani and Kushner (2002) and Sheshadri et al. (2015) used \( c = 10 \) K (which makes the Northern Hemisphere troposphere colder), but that resulted in unrealistically long annular mode timescales in the troposphere, especially in the absence of topography (Sheshadri et al., 2015). With \( c = 0 \) K the annular mode timescales are reasonable in the absence of topography or heating perturbations (Sheshadri and Plumb, 2017). Unlike Sheshadri et al. (2015), the simulations in this chapter did not include a seasonal cycle, but were run under perpetual Northern Hemisphere winter conditions.

A tropospheric diabatic heating perturbation was added to the winter hemisphere to induce Northern Hemisphere winter-like wave activity. The format of the heating perturbation was

\[
Q_0 (\lambda, \phi, p) = \begin{cases} 
q_0 \sin(m \lambda) \exp \left[ -\frac{1}{2} \left( \frac{\phi - \phi_0}{\sigma_\phi} \right)^2 \right] \sin \left( \frac{\pi \log (p/p_0)}{\log (p_t/p_0)} \right), & p_t \leq p \leq p_0 \\
0, & \text{otherwise},
\end{cases}
\]

(2.1)

where \( \lambda \) is longitude, \( \phi \) is latitude, \( p \) is pressure, \( q_0 = 6 \) K/day, \( m \) is the longitudinal
Figure 2-1: The vertical and latitudinal extent of the heating perturbation at the longitude of maximum heating. The contour interval is 0.5 K/day.

wavenumber (1 or 2), \( \phi_0 = 45^\circ \cdot 2\pi / 360^\circ \), \( \sigma_\phi = 0.175 \), \( p_0 = 8.0 \cdot 10^4 \) Pa, and \( p_t = 2.0 \cdot 10^4 \) Pa. Figure 2-1 shows the format of the heating perturbation at the longitude of maximum heating. The model was run with either wavenumber \( m = 1 \) or \( m = 2 \). The value of \( q_0 = 6 \) K/day was chosen because with \( m = 2 \) this heating amplitude produced a SSW frequency similar to a perpetual winter version of the 4000 m wave-2 topographic forcing model used by Sheshadri et al. (2015). The latter was found to produce a Northern Hemisphere winter-like frequency of SSWs when run with a seasonal cycle (0.62 SSWs per year during a 50 year run). Unlike the runs with tropospheric heating, there is a tropospheric asymmetry parameter \( \epsilon = 10 \) K in the model run with topographic forcing. Hereafter the tropospheric heating wave-1 run will be referred to as H1, the tropospheric heating wave-2 run as H2, and the 4000 m wave-2 topographic forcing run as T2. The model runs were 31,000 days long, and the first 1100 days were discarded.
2.2.2 SSW definition and classification

I used the Charlton and Polvani (2007) definition of a SSW: a change from westerly to easterly zonal mean zonal wind at 60°N and 10 hPa. The first day on which this condition is fulfilled is considered the central date. No SSW can be identified without at least 20 consecutive days of westerly zonal mean zonal wind before the central date. This condition prevents the same SSW from counting as two different events because of short term wind fluctuations. To make sure that final warmings are not counted as SSWs in the analysis of the ERA-Interim data I used the same condition that Charlton and Polvani (2007) used: at least 10 consecutive days of westerly zonal mean zonal wind are required after the central date and before 30 April if the wind reversal is to be counted as a SSW, otherwise it is assumed to be the final warming.

Just like in Charlton and Polvani (2007) I looked at dates between 5 days before the central date and 10 days after to classify SSWs as splits or displacements. Unlike Charlton and Polvani (2007) I adopted a classification that is based on geopotential height and not absolute vorticity, and that requires considerably fewer calculations. I considered daily wave-1 \( (A_1) \) and wave-2 \( (A_2) \) amplitudes of geopotential height at 10 hPa and 60°N, and if for any day in the specified range

\[ A_2 - k \cdot A_1 \geq 0, \]

(2.2)

the SSW was considered a split. If those conditions were never fulfilled the SSW was considered a displacement. This method is similar to the one used by Yoden et al. (1999), who also used wave amplitudes of geopotential height to classify SSWs. The parameter \( k \) allows skewing the results towards more or fewer splits. I tried values of \( k \) ranging from 0.8 to 1.1, and settled on \( k = 1.0 \). This value of \( k \) resulted in a split and displacement classification that produced ratios very similar to those obtained by manual inspection of 10 hPa absolute vorticity surfaces. This type of classification will be referred to as “subjective analysis,” after the expression used by Charlton
and Polvani (2007). The subjective analysis consisted of looking at daily values of absolute vorticity at 10 hPa between 5 days before the central date and 10 days after, and classifying the SSW as a split if during any day there were two distinct vortices of comparable magnitudes and horizontal extents.

2.3 Climatology of model runs

The zonal mean zonal winds and zonal mean temperatures for the thermally and topographically forced runs are shown in Figure 2-2. One prominent difference between the different types of forcing is the location of the tropospheric jet maximum. In the case of T2 (Figure 2-2c) the Northern Hemisphere tropospheric jet reaches a maximum around 29°N, while the runs with thermal forcing (Figures 2-2a and b) have the tropospheric jets maximized further poleward: around 43°N for H2 and 35°N for H1. This makes the separation between the tropospheric jet and polar night jet smaller for H1 and H2 compared to T2. The mean location of the tropospheric jet cannot be easily changed when using topographic forcing: Gerber and Polvani (2009) noted that the location of the jet maximum changed by less than 2° in latitude when varying the polar vortex strength under topographic settings that produced SSWs (see Figure 2b in their paper). Another model parameter that affects the mean location of the tropospheric jet in control runs without topography or anomalous heating is the hemispheric asymmetry parameter, but when changing the asymmetry parameter to values ranging from \( \alpha = -30 \) K to \( \alpha = 30 \) K in the presence of topography the mean position of the Northern Hemisphere jet does not change by more than a few degrees (results not shown). Instead it seems the large topography itself is impeding any significant latitudinal shift of the jet. This is not the case in the thermally forced runs, where the jet position is not as strongly constrained.

Another major difference between the topographically and thermally forced runs is the zonal mean zonal winds in the tropical tropopause and lower stratosphere. In the thermally forced runs the zonal mean zonal wind around 50 hPa and the equator is
close to 0 m/s, and there are easterlies of magnitude around 20 m/s centered around 100 to 200 hPa and extending from 20°S to the northern midlatitudes. In T2 the easterlies are more symmetric about the equator and much weaker. There is also no stratospheric equatorial region in the T2 run where the zonal mean zonal wind approaches zero.

The polar night jet has a higher mean strength in T2 compared to the thermally forced runs: just below 60 m/s compared to about 55 m/s for H2 and just over 40 m/s for H1. The jet maximum occurs around 70°N for all runs.

The most prominent difference in the zonal mean temperatures between the different forcings can be found in the upper troposphere and lower stratosphere in the northern midlatitudes. While T2 has a slightly higher zonal mean temperature in that region compared to other latitudes at the same pressure levels (Figure 2-2f), the temperature
is much higher in the thermally forced runs: about 20 K warmer for H2 (Figure 2-2e) and 40 K for H1 (Figure 2-2d) compared to T2.

Figure 2-3 shows the wave-1 and wave-2 components of the divergence of EP flux in the Northern Hemisphere for the three runs. Higher wavenumbers do not contribute significantly to the overall divergence of EP flux in the stratosphere. The divergence of EP flux was calculated based on Edmon Jr. et al. (1980), and the equations were:

\[ F(\phi) = - \cos \phi v'u', \]  
\[ F(p) = f \cos \phi \frac{\overline{v'\theta'}}{\overline{p}}, \]  
\[ \nabla \cdot \vec{F} = \frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} (F(\phi) \cos \phi) + \frac{\partial}{\partial p} F(p). \]  

In the above equations \( \phi \) is latitude, \( p \) is pressure, \( r \) is Earth’s radius, \( f \) is the Coriolis parameter, \( u \) and \( v \) are zonal and meridional winds respectively, and \( \theta \) is potential temperature. Overbars denote zonal means, and primes deviations from the zonal mean. \( F(\phi) \) and \( F(p) \) are the meridional and vertical components of EP flux, respectively. \( \nabla \cdot \vec{F} \) is the divergence of EP flux, and Figure 2-3 shows \( \frac{1}{\cos \phi} \nabla \cdot \vec{F} \). This scaling makes the EP flux proportional to zonal acceleration.

The divergence of EP flux is similar between H1 and H2. There is strong wave-1 convergence in the lower stratospheric extratropics (-9 m/(s·day)) and upper stratospheric high latitudes (-4 m/(s·day)) in H1, and strong divergence (7 m/(s·day)) poleward of about 75°N and above 10 hPa (Figure 2-3a). There is weak wave-2 convergence (around -1 m/(s·day)) throughout most of the stratosphere, except poleward of about 65°N where there is convergence approaching 2 m/(s·day) (Figure 2-3b). The wave-1 convergence and divergence in H2 (Figure 2-3c) is structurally similar but weaker by roughly a factor of two compared to that of H1 except in the lower stratosphere, where H2 does not have wave-1 convergence. The structure in wave-2 for H2 (Figure 2-3d) is also similar to that of H1, but about twice as large in the
Figure 2-3: Divergence of EP flux for H1 (top), H2 (middle) and T2 (bottom). The left column shows the wave-1 components, while the right column shows the wave-2 components. The contour interval is $0.5 \text{ m}/(s \cdot \text{day})$. 
magnitude of convergence but with less divergence. These results make sense in an intuitive way: when the model is forced with wave-1 heating there is more wave-1 activity compared to wave-2, and vice versa when the model is forced with wave-2.

The EP flux divergence for T2 is very different compared to that of H1 and H2. There is strong wave-1 EP flux divergence (6 m/(s \cdot day)) in the polar upper stratosphere (Figure 2-3e), and strong wave-2 convergence (−7 m/(s \cdot day)) poleward of 40°N (Figure 2-3f). This suggests that practically all wave breaking in the stratosphere of T2 is wave-2. Figure 2-3 seems to indicate that the wavenumber of the forcing in the T2 run has a strong impact on the wave activity in the stratosphere, while the wavenumber of the heating in the thermally forced runs has a comparatively small impact on the climatological stratospheric wave activity.

2.4 SSW statistics

The SSW frequencies as well as split and displacement ratios are summarized in Table 2.1. The SSW frequencies of H2 and T2 were 0.48 and 0.42 SSWs per 100 days, respectively. These frequencies are very similar to the SSW frequency found in ERA-Interim reanalysis data ranging from January 1st 1979 through October 31st 2016: 0.42 SSWs per 100 extended winter days (November through March). In the case of H1 the heating amplitude was kept equal to that of H2 to compare the differences between wave-1 and wave-2 forcing, without any thought to matching the SSW frequency to T2 or reanalysis data. This resulted in a SSW frequency of 0.66 per 100 days for H1. There were no SSWs in a 10,900 day long control run without tropospheric wave forcing.

While T2 and H2 have similar SSW frequencies, T2 produces almost only splits while more than a quarter of the SSWs in H2 are displacements. H2 produced 145 SSWs in 30,000 days, and with the wave amplitude classification (WAC) described in equation 2, 108 of those were splits (74%). With subjective analysis (SA) the number of
Table 2.1: SSW frequency and classification distribution. For ERA-Interim the SSWs are counted per 100 extended winter days (November through March).

<table>
<thead>
<tr>
<th></th>
<th>H1</th>
<th>H2</th>
<th>T2</th>
<th>ERA-Interim</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSWs (SSWs per 100 days)</td>
<td>199 (0.66)</td>
<td>145 (0.48)</td>
<td>127 (0.42)</td>
<td>24 (0.42)</td>
</tr>
<tr>
<td>Splits (fraction), WAC</td>
<td>118 (0.59)</td>
<td>108 (0.74)</td>
<td>124 (0.98)</td>
<td>12 (0.5)</td>
</tr>
<tr>
<td>Splits (fraction), SA</td>
<td>108 (0.54)</td>
<td>104 (0.72)</td>
<td>123 (0.97)</td>
<td>13 (0.54)</td>
</tr>
<tr>
<td>Agreement WAC and SA</td>
<td>0.76</td>
<td>0.83</td>
<td>0.98</td>
<td>0.88</td>
</tr>
</tbody>
</table>

splits was 104 (72%). The two different classifications agreed with each other 83% of the time. Figure 2-4 shows the absolute vorticity at 10 hPa on the central date of two events from H2 that both classifications agreed were a split and a displacement, respectively. The split (Figure 2-4a) is typical of splits in H2 in that the two vortices formed immediately after the split are usually located around longitudes 90° and 270° on the 10 hPa surface, or about 45° east of the location of maximum heating. This is similar to the results of Matthewman et al. (2009), who in their analysis of reanalysis data found that there was little variation of longitudinal vortex position between individual SSWs, although they found this to be true for both splits and displacements.

The difference in fraction of splits between WAC and SA was somewhat larger for H1: 118 splits (59%) with WAC and 108 splits (54%) with SA. The agreement between classifications, 76%, was lower than for H2.

While Sheshadri et al. (2015) and Gerber and Polvani (2009) found that wave-2 topographic forcing produced only splits, I found 3 displacements (2%) with WAC and 4 (3%) with SA. The split and displacement distribution that I found in T2 is therefore more similar to the one obtained by Martineau et al. (2018), who also found that while wave-2 topographic forcing produced some displacements almost all SSWs were splits.

To make sure that WAC was not only applicable to this model I compared split and displacement ratios from the ERA-Interim data obtained using both WAC and SA. Out of the 24 SSWs in the data set, WAC classified 12 as splits compared to 13 using
SA. The classifications agreed for 21 out of 24 events (88%), and I concluded that WAC also produces reasonable split and displacement ratios with reanalysis data.

Charlton and Polvani (2007) used the ERA-40 dataset from 1 September 1957 to 31 August 2002, and in the period that overlapped with the ERA-Interim dataset used here there were 15 SSWs. WAC (SA) agreed with the Charlton and Polvani classification (hereafter CPC) for 11 (12) of those 15 SSWs, or 73% (80%) of the time. CPC identified 6 as splits and 9 as displacements, while WAC (SA) identified 8 (9) as splits and 7 (6) as displacements. The agreement between CPC and WAC is lower than the one between WAC and SA for both model runs and reanalysis data, and my methods (both WAC and SA) overestimate the amount of splits compared to CPC.

2.5 Wave fluxes around splits and displacements

The climatological EP flux divergences of the three runs (Figure 2-3) indicate that wave-2 dominates the stratospheric wave activity in T2, while the stratospheric wave activity is less dependent on the wavenumber of the forcing in H1 and H2. This is also clear in the split and displacement ratios of the two runs, where both H1
and H2 produce large numbers of both splits and displacements while T2 produces almost only splits. To understand how the wavenumber of the tropospheric forcing relates to the split and displacement ratios I analyzed vertical EP flux as a function of pressure and latitude around SSWs in the three runs. The wave-1 and wave-2 components of Equation 2.3b were calculated, and then averaged between 40°N and 80°N and composited on the central dates of SSWs. The SSWs were divided into splits and displacements using WAC, and the fluxes around splits and displacements were composited separately. The time mean was subtracted and the results are displayed in Figures 2-5 through 2-7 in terms of standard deviations. Versions of Figures 2-5 through 2-7 where the flux anomalies are expressed as numerical values instead of standard deviations can be found in Appendix A. The statistical significance was assessed with a t-test.

The flux anomalies for H1 are shown in Figure 2-5. About 10 days before the central date of displacements there is a statistically significant wave-1 flux anomaly (Figure 2-5a) that reaches almost -2 standard deviations (hereafter SD). The anomaly is statistically significant far down in the troposphere, and the magnitude of the flux anomaly is largest near the surface (see Figure A-1a in Appendix A) even though the standard deviation is larger in the stratosphere. It therefore seems that, on average, displacements in H1 are associated with a tropospheric wave-1 flux anomaly. There is a negative wave-2 anomaly confined to the stratosphere associated with displacements (Figure 2-5b), of a lesser magnitude and occurring on average about 5 days before the central date. The anomalies associated with splits are interesting: there is a negative wave-1 anomaly (Figure 2-5c) of slightly less than 1.5 SD in magnitude in the stratosphere, which forms about 10 days before the central date. On average, the tropospheric wave-1 flux associated with splits is smaller in both magnitude and vertical extent and shorter in duration compared to the tropospheric wave-1 anomaly during displacements. Once again the negative wave-2 anomaly (Figure 2-5d) forms about 5 days before the central date, with a magnitude of over 1 SD. Somewhat surprisingly, the wave-2 anomaly is smaller for splits compared to displacements.
Figure 2-5: Vertical wave-1 (left) and wave-2 (right) EP flux anomalies around displacements (top) and splits (bottom) in the H1 run, expressed in terms of standard deviations. The EP flux anomalies were averaged between 40°N and 80°N. Areas within green lines are statistically significant at a 95% confidence level. The contour interval is 0.25 standard deviations.

While the vertical EP flux anomalies are almost entirely confined to the stratosphere when expressed in terms of standard deviations, the magnitude of climatological EP flux is much larger in the troposphere than in the stratosphere. Therefore splits in H1 are associated with a reasonably large tropospheric wave-1 flux anomaly, even though Figure 2-5c makes the anomaly look small (for comparison, see Figure A-1c in Appendix A). Nevertheless, splits in H1 are on average associated with less anomalous tropospheric wave activity compared to displacements.

Both splits and displacements in H2 are preceded by EP flux anomalies originating in the troposphere. During displacements there is a small, negative (less than 0.5 SD in magnitude) tropospheric wave-2 anomaly that starts about 15 days before
Figure 2-6: Same as Figure 2-5 but for H2. Notice the increased range of the colorbar for d).

the central date but disappears after around 5 days (Figure 2-6b). When the wave-2 anomaly disappears a negative tropospheric wave-1 anomaly is instead formed, reaching a magnitude exceeding 0.5 SD (Figure 2-6a). This wave-1 anomaly extends into the stratosphere, where the anomaly approaches −1.5 SD just before the central date. Centered around the central date of displacements is a stratospheric wave-2 anomaly of around −1.5 SD. There is a relatively large (up to 1 SD) positive wave-1 flux anomaly following the displacement that extends all the way to the upper troposphere.

The EP flux anomalies around splits in H2 show that the anomalous wave flux around splits is dominated by wave-2. A large negative EP flux anomaly starts to form about 12 days before the central date of splits (Figure 2-6d). The anomaly extends from the troposphere to the stratosphere, and while the magnitude of the tropospheric anomaly
is small in terms of standard deviations the absolute vertical EP flux anomaly is large (see Figure A-2d in Appendix A). The magnitude of the negative anomaly exceeds 3 SD in the middle stratosphere around the central date. Positive anomalies of less than 0.5 SD can be found in the troposphere up to 20 days after the central date. In comparison, there is a small (less than 0.5 SD in magnitude) negative wave-1 anomaly preceding the split, with a modest vertical extent (Figure 2-6c). A weak (less than 0.5 SD) but statistically significant positive wave-1 flux anomaly follows the splits.

The EP flux anomalies around splits in T2 are found in Figure 2-7. Since almost all SSWs in T2 were splits no meaningful statistics could be obtained for displacements, so the anomalies around displacements are not shown in Figure 7. Pressure levels below 450 hPa are unavailable for analysis due to the topography. Structurally the flux anomalies are very similar to those found in splits in H2. There are practically no wave-1 anomalies preceding the SSWs, but positive anomalies following the events of up to 1 SD (Figure 2-7a). The wave-2 flux anomalies become statistically significant around 15 days before the central date, and they are larger in the troposphere than the stratosphere before the central date (Figure 2-7b). Just before the central date the negative wave-2 anomalies reach their largest magnitude of over 3 SD.
Birner and Albers (2017) found that 7 out of 28 SSWs (25%) in their reanalysis dataset were preceded by lower tropospheric wave events (LTWEs). They defined an LTWE as a 10-day average upward EP flux at 700 hPa averaged between 45°N and 75°N that exceeded two standard deviations. They looked at both wave-1 and wave-2 fluxes. To complement Figures 2-5 through 2-7 I performed a similar analysis, where I searched for the strongest 10-day average upward EP flux at 450 hPa averaged between 40°N and 80°N before each SSW in the datasets. The 10-day averages were calculated with center dates ranging from 20 days before the SSW to the central date. The pressure level of 450 hPa was chosen so that T2 could be included. Unlike Birner and Albers (2017) I did not separate the vertical EP flux into wavenumbers for this analysis, since the average structure of the wave flux anomalies were already shown in Figures 2-5 through 2-7 and I am interested in the full flux around the events. I found that the number of SSWs preceded by vertical EP flux anomalies exceeding two standard deviations (2 SD) was 46 (23%) for H1 and 32 (22%) for H2. To assess the statistical significance of differences between splits and displacements I produced 100,000 sets of random data for both H1 and H2, where each set contained a number of entries equal to the number of SSWs in the runs (199 and 145, respectively). The entries consisted of ones and zeros, where the number one represented a SSW forced by anomalous flux exceeding the threshold of 2 SD. Therefore, 46 out of 199 and 32 out of 145 entries were designated as ones. According to WAC there were 118 splits and 81 displacements in H1, and 108 splits and 37 displacements in H2. I randomly picked the corresponding numbers of splits and displacements from each dataset. From the 100,000 datasets I could then find the 5th and 95th percentiles of splits and displacements that would be forced by anomalous vertical EP flux exceeding 2 SD if there was no difference between the forcing of splits and displacements. The results can be found in Table 2.2.

SSWs in both H1 and H2 are preceded by anomalous vertical EP flux exceeding 2 SD at about the same frequency as the number found by Birner and Albers (2017): 23% and 22% for H1 and H2, compared to 25%. Interestingly, there is no statistically
significant difference between splits and displacements in H2, even though Figure 6 indicates that splits are on average preceded by strong wave-2 flux. This is likely due to the fact that displacements in H2 are on average preceded by both anomalous wave-1 and wave-2 flux, and even though the fluxes from the individual wavenumbers are small on average their sums add up to a relatively strong total flux. Splits in H2, on the other hand, are on average not preceded by any significant tropospheric wave-1 flux. The fractions of splits and displacements in H1 that are preceded by anomalous flux exceeding 2 SD are different from each other at a 95% confidence level, with 32% of displacements and 17% of splits forced by such tropospheric wave flux anomalies. Like Figure 5 indicated, displacements in H1 are more likely to be forced by anomalous tropospheric wave-1 flux compared to splits. Given the small amounts of displacements in T2 I did not perform any statistical analysis for that model run, but a large amount (35%) of SSWs in T2 are forced by flux anomalies exceeding 2 SD. The fraction is the same for total SSWs and splits only.

It seems that when SSWs are associated with tropospheric flux anomalies exceeding 2 SD the SSW is more likely to depend on the wavenumber of the forcing, with wave-1 forcing favoring displacements and wave-2 favoring splits. The results from H1 show that almost twice the fraction of displacements are associated with such anomalies compared to splits, and although the differences are not statistically significant in H2 there is an indication that this may be the case in that model setup as well, with the fraction associated with splits being 50% higher than that of displacements.

The fact that the fractions of SSWs forced by anomalous tropospheric wave flux exceeding 2 SD in the three model runs is similar to what Birner and Albers (2017) found in reanalysis data indicates that SSWs produced in this model may be formed by processes similar to those in the observed atmosphere. T2 has much too many splits compared to observations and can reasonably be expected to be least similar to observations in terms of SSW generation, but even in that model run about two thirds of SSWs occur without anomalous tropospheric wave flux exceeding 2 SD. This suggests that previous authors who have used this idealized GCM with topographic
Table 2.2: Number of splits and displacements in the three runs that were forced by anomalous vertical EP flux, along with the numbers expected if there was no difference between the forcing of splits and displacements. See text for details.

<table>
<thead>
<tr>
<th></th>
<th>H1</th>
<th>H2</th>
<th>T2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Displacements (fraction)</td>
<td>26 (0.32)</td>
<td>6 (0.16)</td>
<td>2 (0.67)</td>
</tr>
<tr>
<td>Expected disp. (5th/95th perc.)</td>
<td>19 (14/24)</td>
<td>8 (5/12)</td>
<td>-</td>
</tr>
<tr>
<td>Splits (fraction)</td>
<td>20 (0.17)</td>
<td>26 (0.24)</td>
<td>43 (0.35)</td>
</tr>
<tr>
<td>Expected splits (5th/95th perc.)</td>
<td>27 (22/32)</td>
<td>24 (20/27)</td>
<td>-</td>
</tr>
</tbody>
</table>

wave-2 forcing (e.g., Gerber and Polvani, 2009; Sheshadri et al., 2015; Martineau et al., 2018) may have produced SSWs for reasons similar to those in the observed atmosphere.

The choice of two standard deviations as a threshold for considering an SSW “forced” by tropospheric wave flux anomalies is high and somewhat arbitrary. However, even if the threshold is decreased to one standard deviation only 73%, 65% and 76% of SSWs in H1, H2 and T2 reach this threshold. That a quarter to a third of SSWs in these runs occur without the tropospheric flux anomalies in a 21 day window reaching even one standard deviation indicates that a significant amount of SSWs occur without anomalous tropospheric forcing. These SSWs are therefore most likely caused by internal stratospheric processes, where the stratosphere allows a larger than usual amount of tropospheric wave flux to propagate into the upper stratosphere. Hitchcock and Haynes (2016) found that the evolution of the state of the stratosphere during SSW onset is crucial in determining the wave flux during the event, and the results above further confirm the importance of the state of the stratosphere in SSW formation.
2.6 Differences in polar cap height following splits and displacements

In order to investigate the impacts of splits and displacements on tropospheric circulation in the model I composited zonal mean geopotential height anomalies from 65°N to 90°N on the central dates of splits and displacements in H1 and H2 (using WAC), and on splits in T2. The anomalies were normalized by the standard deviations of geopotential height in the respective runs. Once again a t-test was used to assess statistical significance at the 95% level. Figures 2-8c and d show the anomalies in H2 following displacements and splits, respectively. In both cases the anomalies are positive before and after the central date, and the anomalies last longer in the lower stratosphere compared to the upper stratosphere. There are tropospheric anomalies associated with both types of events, but in the case of splits the tropospheric anomalies seem to precede stratospheric anomalies of similar magnitude by about 20 days. The lower tropospheric anomalies associated with splits remain statistically significant for over 80 days following the central date, although the anomalies following displacements are slightly weaker and not statistically significant to the same extent. The largest tropospheric anomalies associated with splits approach 1 SD, but tropospheric anomalies barely exceed 0.5 SD for displacements. However, the difference between tropospheric anomalies associated with displacements and tropospheric anomalies associated with splits is only statistically significant at the 95% level between about 12 and 4 days before the central date, where the anomalies before splits are slightly stronger (assessed using a t-test; not shown), so while both splits and displacements have statistically significant zonal mean surface impacts in H2, there is little evidence for splits and displacements having different zonal mean surface impacts with this forcing.

The difference between splits and displacements is larger for H1. Displacements (Figure 2-8a) are preceded by statistically significant negative anomalies of a magnitude over 0.5 SD that extend from the surface to the middle stratosphere, and following
Figure 2-8: Geopotential height anomalies averaged from 65°N to 90°N and normalized by their standard deviations. The anomalies are composited on the central dates of displacement events (left) and split events (right) in H1 (top) and H2 (middle), using WAC. The bottom figure shows the anomalies around splits in T2. Areas within green lines are statistically significant at a 95% confidence level. The contour interval is 0.25 standard deviations.
the central date there are no tropospheric anomalies. There are barely any negative anomalies preceding splits (Figure 2-8b), and there are positive anomalies approaching 0.5 SD following the central date. These tropospheric anomalies last from almost 20 days after the central date to about 80 days after the central date. Unlike H2, a t-test reveals that the difference between splits and displacements is statistically significant at the 95% level throughout most of the troposphere (not shown). Although the zonal mean geopotential height anomalies around splits and displacements in H1 look very different, their tropospheric impacts are quite similar if the anomalies before the central date are compared with the anomalies following: in both cases the SSW changes the anomaly by about 0.5 SD. With this in mind it seems the zonal mean geopotential height anomalies do not tell us much about the difference in tropospheric impact of splits and displacement, but rather the state of the polar region before the events. It is possible that the state of the polar region may affect whether splits or displacements are produced in this model configuration, although it should be noted that Charlton and Polvani (2007) did not find significant polar cap temperature differences between splits and displacements in their reanalysis data set.

Splits in T2 (Figure 2-8e) are not preceded by any tropospheric anomalies, and among the model setups it seems splits in T2 has the smallest impact on polar geopotential height in the lower troposphere. While there are positive anomalies in the upper troposphere following the central date, the only surface impacts can be found on the central date and between about 20 and 30 days following the splits, and they do not reach 0.5 SD.

The differences between tropospheric impacts of splits and displacements in H1 are similar to those found by Mitchell et al. (2013). They used ERA-40 reanalysis data and calculated the NAM as the leading EOF of daily geopotential anomalies poleward of 20°N, and looked at NAM anomalies associated with splits and displacements after subtracting the seasonal cycle. They found that negative NAM anomalies following displacements stopped at the tropopause, while negative NAM anomalies following splits could reach the surface and persist for around 60 days (Figure 4 in their paper).
The structure of zonal mean geopotential height anomalies associated with splits and displacements in H1 are very similar to the results obtained by Mitchell et al. (2013). It seems the wavenumber of the forcing is important for the tropospheric response to SSWs in this model.

2.7 Conclusions

In this chapter I have demonstrated that Northern Hemisphere winter-like stratospheric variability can be produced in an atmospheric GCM with tropospheric heating perturbations. When run with topographic wave-2 forcing (T2) this model produces almost only split SSWs, but I found that with tropospheric heating the model produces large numbers of both splits and displacements with both wave-1 (H1) and wave-2 forcing (H2).

I created a wave amplitude classification (WAC) of splits and displacements, and showed that it compared favorably with subjective analysis (SA) of SSWs. The classification compared less favorably with the one used by Charlton and Polvani (2007) when analyzing reanalysis data, and I overestimate the amount of splits with both SA and WAC compared to Charlton and Polvani (2007). I further showed that while H1 and H2 both cause splits and displacements, the ratio of splits and displacements for H1 is slightly more sensitive to the SSW classification while the agreement between classifications for H1 is lower compared to H2.

I used the results of the three model runs (T2, H1, H2) to investigate the extent to which SSWs are caused by anomalous tropospheric wave activity. I analyzed anomalous wave-1 and wave-2 vertical EP flux composited on splits and displacements, as well as the 10-day anomalous tropospheric vertical EP flux around each SSW. I found that in the case of splits in T2 and H2, the anomalous wave flux around the events is dominated by its wave-2 component. Displacements in H2 and H1 were associated with anomalous tropospheric wave-1 EP flux, but in the case of displacements in H2
the events were also preceded by anomalous tropospheric wave-2 flux. Splits in H1, however, were on average preceded by very little anomalous tropospheric EP flux, and the anomalous EP flux associated with the events seemed to be found mostly in the stratosphere. Birner and Albers (2017) found that only 7 out of 28 (25%) SSWs in reanalysis data were preceded by anomalous tropospheric EP fluxes exceeding two standard deviations. My analysis of anomalous EP flux around individual events gave similar fractions of SSWs above this threshold: 23%, 22% and 35% for H1, H2, and T2 respectively. I also found that the fraction of displacements forced by strong, anomalous tropospheric wave fluxes was higher than splits in H1, at a 95% confidence level. I found no statistically significant difference in this fraction between splits and displacements in H2. When the threshold for anomalous vertical EP flux is lowered to one standard deviation a quarter to a third of SSWs do not reach this threshold, which strongly suggests that some of the observed SSWs are caused by internal stratospheric processes which alter the amount of tropospheric wave flux permitted to propagate into the stratosphere. It therefore seems that SSWs in this model can form both as a result of anomalous tropospheric wave activity and changes in the state of the stratosphere. The fact that the fraction of SSWs associated with strong tropospheric EP flux anomalies is similar to the fraction found in reanalysis data indicates that this model may be producing SSWs for reasons similar to those in the observed atmosphere. Apart from validating the mechanisms behind SSW formation in these model runs, this might also apply to other authors who have used versions of this model with topographic wave-2 forcing to produce SSWs (e.g., Gerber and Polvani, 2009; Sheshadri et al., 2015; Martineau et al., 2018).

Tropospheric polar geopotential height anomalies associated with splits were found to be different from those of displacements in H1, and similar in structure to anomalies found by Mitchell et al. (2013). Interestingly, the tropospheric effects of splits and displacements in H1, as measured by polar geopotential height anomalies preceding and following SSWs, were not very different. Instead it seems the zonal mean geopotential height before displacements is lower compared to before splits, at a 95%
confidence level. This indicates that the state of the polar region may have an influence on whether a split or displacement is formed in this model configuration. There were barely any statistically significant differences between splits and displacements in H2. It seems the wavenumber of the forcing is important for tropospheric anomalies associated with SSWs in this model, and especially displacements.

In this chapter I investigated the formation of splits and displacements in a frequently used idealized atmospheric GCM by analyzing SSWs formed when using three different types of forcing. I found that while the ways SSWs form vary with the type of forcing used and type of SSW produced the fraction of SSWs forced by anomalous tropospheric wave flux is relatively similar to the fraction previously found in reanalysis data, regardless of the type of forcing. I also found that the zonal mean surface signatures of splits and displacements were only different with one type of forcing, but that the difference between the impacts of splits and displacements is very small. Rather than splits and displacements having different impacts with this forcing, it seems more likely that the state of the polar region affects the type of SSW produced.
Chapter 3

The role of eddy-eddy interactions in sudden stratospheric warming formation

3.1 Introduction

SSWs occur when large upward propagating planetary scale waves break in the stratosphere and deposit momentum, thereby disturbing the winter polar vortex enough to exceed a given threshold (Butler et al., 2015), the most common of which is a reversal of the zonal mean westerlies at 60° and 10 hPa (Charlton and Polvani, 2007). SSWs can therefore be considered wave-mean flow interactions, and it has long been known that SSW-like zonal mean wind variations can occur in model setups as simple as one-dimensional β-plane models (e.g., Holton and Mass, 1976; Yoden, 1990). However, displacements and splits are wave-1 and wave-2 disturbances, respectively, and in Chapter 2 it was shown that large amounts of both split and displacement SSWs occur with both pure wave-1 and pure wave-2 heating perturbations. Furthermore, both wave-1 and wave-2 flux anomalies were associated with both splits and displace-
ments, with both heating wave-1 (H1) and wave-2 (H2) forcing. This indicates that some wave-wave (eddy-eddy) interaction takes place somewhere between the wave formation in the troposphere and the wave breaking in the stratosphere.

There is reason to believe that splits and displacements are not equally dependent on eddy-eddy interactions (hereafter shortened EEI). Previously published work has found that the vertical structures of splits and displacements in reanalysis data are different, with splits having deep, barotropic structures while displacements are clearly baroclinic (Matthewman et al., 2009). Furthermore, there is some evidence for splits being caused by resonance rather than anomalous wave forcing, both in reanalysis data (Albers and Birner, 2014) and simple models (Matthewman and Esler, 2011). In contrast, Esler and Matthewman (2011) found that displacements in the same model exhibited a much more complicated non-linear behavior compared to splits, and that strong non-linear processes were required to explain the amplitude of the model response to a given forcing. Albers and Birner (2014) found that displacements in reanalysis data result from waves breaking in the stratosphere and weakening the vortex, which is more in line with the traditional view of SSWs (e.g., Matsuno, 1971).

The importance of anomalous tropospheric wave fluxes for SSWs was investigated in detail in Chapter 2, where it was shown that the fraction of SSWs associated with strong tropospheric wave flux anomalies in the GCM was similar to the one found in reanalysis data by Birner and Albers (2017). Birner and Albers (2017) emphasized the importance of what they referred to as the “communication layer:” the 300-200 hPa region around or just above the tropopause. Polvani and Waugh (2004) also found that anomalous wave fluxes at 100 hPa and further down in the troposphere precede SSWs, and that the origin of SSWs can be found in the troposphere. In Chapter 2 it was found that displacements in H1 were more likely to be associated with strong tropospheric wave flux anomalies compared to splits in H1. This indicates that splits and displacements may not be equally dependent on dynamical changes, including changes in EEI, in given parts of the atmosphere.
The effects of EEI can be removed from model simulations by calculating the advection of eddy fluctuations by eddy winds, and replacing them with their zonal mean values. O’Gorman and Schneider (2007) used this approach with an idealized GCM and found that many aspects of the general circulation remain the same without EEI. Many authors have used similar approaches to investigate the effects of EEI on zonal jet formation and equilibration, both with idealized GCMs (Chemke and Kaspi, 2016) and β-plane models (Srinivasan and Young, 2012; Tobias and Marston, 2013; Constantinou et al., 2014).

In this chapter I investigate the role of EEI in SSW formation by removing them using the method of O’Gorman and Schneider (2007). I will answer three specific questions related to EEI and SSW formation:

1. To what extent do EEI affect SSW frequencies?

2. To what extent are splits and displacements, and their ratios, affected by EEI?

3. At what pressure levels are EEI important for SSWs, splits and displacements?

In order to answer these questions I ran the H1 and H2 model setups from Chapter 2 under three additional settings for each forcing: one without EEI anywhere, one without EEI in the troposphere and lower stratosphere, and one without EEI in the upper stratosphere. The latter two are hereafter referred to as the mixed runs. Section 3.2 describes the method used to remove EEI, and the way it was implemented in the model runs. Section 3.3 compares the zonal wind and wave flux climatologies of the eight model runs. Section 3.4 compares the SSW frequencies and split and displacement ratios of the different model runs. Finally, a discussion of the results and conclusions can be found in Section 3.5.
3.2 Removal of eddy-eddy interactions

Following O’Gorman and Schneider (2007), we calculated the tendency due to EEI (the eddy-eddy tendency), subtracted it from the total tendency of horizontal wind and temperature, and added the zonal mean value of the eddy-eddy tendency (the mean tendency) to the total tendency equation. This substitution was described by O’Gorman and Schneider (2007) by using the equation for temperature tendency as an example. In the control run the evolution is

\[
\frac{\partial T}{\partial t} = -v \frac{\partial T}{\partial y} + ..., \\
= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v} \frac{\partial T'}{\partial y} - v' \frac{\partial \bar{T}}{\partial y} - v' \frac{\partial T'}{\partial y} + ..., \tag{3.1}
\]

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 Equation 3.1 describes the advection of temperature eddies due to meridional wind eddies. In the model runs where EEI are not allowed Equation 3.1 becomes

\[
\frac{\partial T}{\partial t} = -v \frac{\partial T}{\partial y} + ..., \\
= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v} \frac{\partial T'}{\partial y} - v' \frac{\partial \bar{T}}{\partial y} + ..., \tag{3.2}
\]

where the contribution due to EEI has been replaced by its zonal mean value.

The wave-1 and wave-2 heating perturbations from Chapter 2 were used in this chapter as well. In addition to the H1 and H2 runs, three additional runs with varying levels of EEI allowed were performed for each wavenumber. All model runs were 31,100 days long, and the last 30,000 days were used for the analysis. The vertical
structure of the four model runs used for each wavenumber can be found in Figure 3-1.

In the control runs (model runs H1 and H2 from Chapter 2; black line in Figure 3-1) EEI are allowed everywhere. In the no EEI-anywhere run (hereafter shortened NE1 or NE2 depending on the wavenumber of the forcing; red line in Figure 3-1) the substitution from Equation 3.1 to Equation 3.2 is performed at every pressure level.

In the mixed runs the model switches between allowing and not allowing EEI linearly with pressure. In the case of the runs with no EEI allowed in the troposphere and lower stratosphere (hereafter shortened NET1 or NET2; green line in Figure 3-1) the following substitution is made:

\[
\frac{\partial T}{\partial t} = \begin{cases} 
\frac{\partial T}{\partial t} - u \frac{\partial T'}{\partial y} + v \frac{\partial T'}{\partial y}, & p > p_1, \\
\frac{\partial T}{\partial t} + \left(1 - \frac{p_1 - p}{p_1 - p_2}\right) \left(-u \frac{\partial T'}{\partial y} + v \frac{\partial T'}{\partial y}\right), & p_2 \leq p \leq p_1, \\
\frac{\partial T}{\partial t}, & p < p_2.
\end{cases}
\]

Figure 3-1: Vertical structure of the four runs used for each wavenumber. See text for details.
In the above equation the temperature tendency from Equations 3.1 and 3.2 has been used as an example. \( p_1 = 50 \text{ hPa} \) and \( p_2 = 30 \text{ hPa} \). Similarly, when EEI are not allowed in the upper stratosphere (hereafter shortened NEs1 or NEs2; blue line in Figure 3-1) the equation describing the substitution is

\[
\frac{\partial T}{\partial t} = \begin{cases} 
\frac{\partial T}{\partial t} + \frac{p_1 - p}{p_1 - p_2} \left( -v \frac{\partial T'}{\partial y} + v' \frac{\partial T'}{\partial y} \right), & p_2 \leq p \leq p_1, \\
\frac{\partial T}{\partial t} - \frac{v}{\partial y} \frac{\partial T'}{\partial y} + v' \frac{\partial T'}{\partial y}, & p < p_2.
\end{cases}
\tag{3.4}
\]

The mixed runs enable an investigation of the effects of EEI 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, which is probably due to the strong climatological wave convergence in that region (see Figure 2-3a through d). Instead the 50 hPa and 30 hPa levels were chosen as start and end points of the transition region, since that is an unusually calm region of the atmosphere in terms of wave activity.

### 3.3 Climatology

Figures 3-2 and 3-3 show the climatological zonal mean zonal wind for the eight model runs. Comparisons of the control runs (Figures 3-2a and 3-3a) to the no EEI-anywhere runs (Figures 3-2b and 3-3b) show that the model keeps much of the climatological zonal mean zonal wind structure even in the absence of EEI, although with some notable exceptions. For one, the polar night jet is much more separated from the tropospheric jet when EEI are not allowed in the troposphere and lower stratosphere (panels b and d in Figures 3-2 and 3-3). The jet strength, however, is largely unaffected in the no EEI-anywhere runs. This is not the case in all model runs, especially NEs2 (Figure 3-3c) compared to the other model runs with wave-
forcing. The area of zonal mean zonal wind approaching zero m/s found in the equatorial lower stratosphere in the control runs is not reproduced when EEI are not allowed in the troposphere and lower stratosphere (panels b and d), indicating that EEI play an important role in this area. Furthermore, there are two tropospheric jets in the 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 EEI, and Chemke and Kaspi (2016) showed that EEI actually decrease the number of eddy-driven jets in the atmosphere by narrowing the latitudinal region where zonal jets can appear. They found that eddy-mean flow interactions can maintain several zonal jets when EEI are removed.

Comparisons between the mixed runs and the control or no EEI-anywhere runs show that changes in the upper stratosphere have very little influence on the climatology of the troposphere and lower stratosphere. When EEI are not allowed in the upper stratosphere only (Figures 3-2c and 3-3c) the zonal mean zonal winds at pressure levels below 50 hPa are very similar to those of the control runs (Figures 3-2a and 3-3a). Similarly, when EEI are not allowed in the troposphere and lower stratosphere (Figures 3-2d and 3-3d) the climatology at pressure levels below 50 hPa are very similar to those of the no EEI-anywhere runs (Figures 3-2b and 3-3b).

Figures 3-4 and 3-5 show the climatological wave-1 and wave-2 components of divergence of EP-flux for the model runs. Panels a and b are reproduced from panels a through d in Figure 2-3, and show the values for H1 and H2. The fact that both H1 and H2 have significant wave-1 and wave-2 EP flux divergence components in the stratosphere shows that there is a large amount of both wave-1 and wave-2 activity in both control runs. As was discussed in Chapter 2, this results in large amounts of both splits and displacements in both runs. In contrast, removal of EEI everywhere (panels c and d) results in an EP flux divergence completely dominated by the wavenumber of the forcing, with NE1 having practically only wave-1 and no wave-2 EP flux convergence (Figure 3-4c and d) while the opposite is true for NE2 (Figure 3-5c and d). This result is not surprising: removal of EEI means that waves can only
Figure 3-2: Zonal mean zonal winds for H1 (a), NE1 (b), NEs1 (c) and NEt1 (d). The contour interval is 5 m/s.
Figure 3-3: Zonal mean zonal winds for H2 (a), NE2 (b), NEs2 (c) and NEt2 (d). The contour interval is 5 m/s.
interact with the mean flow, which limits the possibilities of energy transfer between wavenumbers.

Just like in Figures 3-2 and 3-3, the mixed runs (panels e through h) show that pressure levels below 50 hPa are remarkably similar to those in the control runs (for NEs1 and NEs2) or no EEI-anywhere runs (for NET1 and NET2). The mixed runs exhibit some other interesting aspects as well: when EEI are turned off in the troposphere and lower stratosphere there is a region of EP flux divergence in the upper stratosphere, where EEI are allowed (Figures 3-4f and 3-5e). This indicates a source of wave activity at this level. The wavenumber of this wave source is opposite to that of the tropospheric forcing and coinciding with areas of large EP flux convergence in the wave number of the forcing, which suggests that once EEI are allowed some of the wave activity is transferred from the wavenumber of the forcing to the other of the two major stratospheric wavenumbers. Another aspect worth noting is that the stratospheric EP flux convergence of the wavenumber opposite to that of the forcing is comparable to that of the control runs when EEI are turned off in the upper stratosphere only (panels g and h). This is especially true for NEs1, which has a substantial amount of stratospheric wave-2 EP flux convergence (Figure 3-4h). There is stratospheric wave-1 EP flux convergence in NEs2, although it is more modest compared to the wave-2 component (Figure 3-5g compared to 3-5h). This indicates that much of the energy transfer between wavenumbers that occurs in the control runs happens below 50 hPa. Finally, it should also be noted that the areas of strongest EP flux convergence occur further poleward when EEI 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 waveguide, since the latitudinal polar night jet shifts between the model configurations are modest (Figures 3-2 and 3-3).

Figures 3-6 and 3-7 show the vertical component of EP flux \( (F_p) \) scaled by \( p_0/p \), where \( p_0 = 1000 \) hPa. The scaling enables plotting of vertical EP flux for the full depth of the atmosphere in the same panel. \( F_p \) can be thought of as a measurement
Figure 3-4: Divergence of EP flux for H1 (a and b), NE1 (c and d), NEt1 (e and f) and NEs1 (g and h). The left column shows the wave-1 components, while the right column shows the wave-2 components. The contour interval is 0.5 m/(s · day).
Figure 3-5: Divergence of EP flux for H2 (a and b), NE2 (c and d), NEt2 (e and f) and NEs2 (g and h). The left column shows the wave-1 components, while the right column shows the wave-2 components. The contour interval is 0.5 m/(s · day).
Figure 3-6: Scaled vertical component of EP flux for H1 (a and b), NE1 (c and d), NEt1 (e and f) and NEs1 (g and h). See text for details. The left column shows the wave-1 components, while the right column shows the wave-2 components. Notice the differences in colorbar ranges and contour intervals.
Figure 3-7: Scaled vertical component of EP flux for H2 (a and b), NE2 (c and d), NEt2 (e and f) and NEs2 (g and h). See text for details. The left column shows the wave-1 components, while the right column shows the wave-2 components. Notice the differences in colorbar ranges and contour intervals.
of the upward propagating wave activity in the model. Panels a and b in Figure 3-6 show that the climatological vertical EP flux in H1 have strong wave-1 and wave-2 components, with wave-2 being present even in the troposphere. The tropospheric wave-2 component disappears when EEI are removed everywhere, while the wave-1 component is stronger than in the control run (panels c and d). This suggests that the tropospheric wave-2 flux in the control run is formed through EEI, and that when EEI are removed the wave flux remains in wave-1. The climatological $F_p$ is very different when the model is forced by wave-2 heating: panels a and b in Figure 3-7 show that there is practically no tropospheric wave-1 flux in H2. Instead the climatological vertical EP flux seems to keep the wavenumber of the forcing. The difference between the control run and no EEI-anywhere is therefore smaller for wave-2 forcing compared to wave-1: in both H2 and NE2 (panels a and b versus c and d in Figure 3-7) the vertical EP flux is entirely dominated by the wave-2 component.

Just like in Figures 3-2 through 3-5, $F_p$ at pressure levels greater than 50 hPa depends on whether or not EEI are allowed at these levels: NE1 and NE2 (panels c and d) look very similar to NEt1 and NEt2, while H1 and H2 look like NEs1 and NEs2 in these regions. This indicates that changing the conditions for EEI above 50 hPa does not affect the climatological wave forcing from lower levels, and it will enable us to answer how important the upper stratosphere is for SSW generation compared to the lower levels that have been highlighted by other authors (Polvani and Waugh, 2004; Birner and Albers, 2017).

The wave sources that could be seen in Figures 3-4f and 3-5e can also be seen in the panels showing $F_p$: in Figure 3-6f the vertical wave-2 flux starts in this region, while Figure 3-7e shows both upward and downward vertical EP flux in the region of the wave source.

Figures 3-4 and 3-5 showed that the wave convergence occurred further poleward when EEI were allowed in the stratosphere. This is also true for the vertical EP flux, and there is an often substantial shift in upper stratospheric $F_p$ when EEI are turned
3.4 Impact on sudden stratospheric warmings

Tables 3.1 and 3.2 show the SSW frequencies and split and displacement ratios for the eight runs. A SSW was defined as reversal of the zonal mean westerlies at 60°N and 10 hPa (Charlton and Polvani, 2007), and the wave amplitude classification (WAC) described in Chapter 2 was used to classify the SSWs as splits or displacements.

Table 3.1 shows that removal of EEI affects SSW frequencies quite significantly when the model is forced by wave-1 heating. The SSW frequency in H1 is 0.66 SSWs per 100 days, but without EEI anywhere the frequency is increased to 0.82 (a 24% increase). The frequencies are lower in the mixed runs: 0.44 in NEt1 and 0.31 in NEs1 (decreases of 34% and 53% compared to the control run, respectively). It is reasonable that the SSW frequency in the wave-1 runs should be affected by EEI removal given how strongly EEI affects the climatological wave forcing. As was seen in Figure 3-6 H1 has strong wave-1 and wave-2 components of $F_p$ in the troposphere, and the wave-2 component disappears when EEI are not allowed. In NE1 the stronger wave-1 forcing and weaker wave-2 forcing results in a higher SSW frequency. However, the results from the mixed runs are more surprising: removal of EEI in the upper stratosphere only (NEs1) decreases the SSW frequency by over 50% compared to the control run. Similarly, allowing EEI in the upper stratosphere only (NEt1) decreases the SSW frequency by 45% compared to NE1. As was mentioned in the previous section, the tropospheric and lower stratospheric wave forcing depends on whether or not EEI are allowed at these levels, and not on the conditions in the upper stratosphere. The fact that H1 and NEs1 as well as NE1 and NEt1 have practically identical wave forcings and climatologies below 50 hPa but very different SSW frequencies shows that EEI in the stratosphere play a major role in SSW generation. It also highlights the fact that the upper stratosphere is not a passive recipient of wave forcing from below, even though the importance of tropospheric and lower stratospheric wave
Table 3.1: SSW frequencies and classifications for the four runs with wave-1 heating perturbation.

<table>
<thead>
<tr>
<th></th>
<th>H1</th>
<th>NE1</th>
<th>NEt1</th>
<th>NEs1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total SSWs (SSWs per 100 days)</td>
<td>199 (0.66)</td>
<td>247 (0.82)</td>
<td>132 (0.44)</td>
<td>93 (0.31)</td>
</tr>
<tr>
<td>Total splits (fraction), WAC</td>
<td>118 (0.59)</td>
<td>[143 (0.57)]</td>
<td>105 (0.80)</td>
<td>[74 (0.80)]</td>
</tr>
</tbody>
</table>

Forcing for SSW generation has often been emphasized (Polvani and Waugh, 2004; Birner and Albers, 2017). However, there is no clear answer to how EEI in the upper stratosphere influence SSW frequencies with wave-1 forcing: a comparison between H1 and NEs1 suggests that EEI are necessary in the upper stratosphere to get high SSW frequencies, while the results for NE1 and NEt1 indicate that allowing EEI in the upper stratosphere decreases the SSW frequency.

The differences in SSW frequencies between the four runs with wave-2 forcing can be seen in Table 3.2. Unlike the runs with wave-1 forcing, removal of EEI in the troposphere and lower stratosphere does not dramatically alter the SSW frequency: NE2 and NEt2 have SSW frequencies of 0.51 and 0.45 SSWs/100 days, compared to 0.48 in the control run. This makes sense when one considers the vertical EP fluxes seen in Figure 3-7: almost all tropospheric and lower stratospheric wave forcing in H2 is in the wavenumber of the forcing, so removal of EEI does not affect the climatological forcing as strongly as is the case when wave-1 forcing is used. However, when EEI are removed in the upper stratosphere only (NEs2) the SSW frequency increases by 37% compared to H2. This increase in SSW frequency could be the reason for the weakened climatological polar night jet seen in Figure 3-3c, although another explanation is that removal of EEI 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 upper stratospheric EEI play a major role in SSW generation. Interestingly, this difference in SSW frequency is completely different from the one between NE1 and NEs1: with wave-2 forcing the SSW frequency increases without EEI in the upper stratosphere, while the opposite is true with wave-1 forcing.

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Table 3.2: Same as Table 3.1 but for the runs with wave-2 heating perturbation.

<table>
<thead>
<tr>
<th></th>
<th>H2</th>
<th>NE2</th>
<th>NEt2</th>
<th>NEs2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total SSWs (SSWs per 100 days)</td>
<td>145 (0.48)</td>
<td>153 (0.51)</td>
<td>134 (0.45)</td>
<td>199 (0.66)</td>
</tr>
<tr>
<td>Total splits (fraction), WAC</td>
<td>108 (0.74)</td>
<td>[153 (1)]</td>
<td>134 (1)</td>
<td>[179 (0.90)]</td>
</tr>
</tbody>
</table>

Tables 3.1 and 3.2 also contain the numbers and fractions of splits in the model runs. H1 has 59% splits, and this number is increased to 80% in NEt1. This result seems counterintuitive: as was mentioned above H1 has strong climatological tropospheric wave-1 and wave-2 flux, while the tropospheric forcing is almost pure wave-1 when EEI are not allowed. A look at Figure 3-6 shows that NEt1 has stronger upper stratospheric wave-1 flux and weaker wave-2 flux compared to H1. A possible explanation for the split and displacement ratio in NEt1 can be found in panels e and f in Figure 3-4. 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 likely show regions of wave-1 to wave-2 energy transfer, rather than wave-mean flow interactions. With that in mind, the wave-2 convergence seen in panel f may be sufficient to cause a large number of splits if this convergence is due to wave-mean flow interactions. A possible explanation for the high split ratio in NEt1 is therefore this: strong tropospheric wave-1 flux reaches the 50 hPa level, where EEI are allowed. Here much of the wave-1 flux is transferred to wave-2 (the areas of convergence and divergence between around 50 and 30 hPa). The wave-2 source centered around 60°N produces vertical wave-2 flux (seen in Figure 3-6f) which interacts with the mean flow at pressure levels above 30 hPa. While the wave-2 EP flux convergence is large at these levels, much of the convergence is due to EEI and not wave-mean flow interactions, so the wave-2 mean flow interaction is of a magnitude comparable to that of the wave-1 mean flow interaction. In fact, addition of the wave-1 and wave-2 EP fluxes show that the wave-2 component is responsible for about 20-30% of the total climatological EP flux convergence between 50 and 60°N and 10 and 20 hPa (not shown). While the wave-1 component is still larger than the wave-2 component, this number is reasonable given that the corresponding ratio for H1 is 30-40% (not shown). Why NEt1 has a higher ratio of splits than H1 when
the climatological EP flux convergence of H1 has a larger fraction of wave-2 compared to NEt1 is not clear, but it is likely that the climatological EP flux structure is insufficient to fully describe SSW generation in this model.

The fractions of splits for NE1 (57%) and NEs1 (80%) are in brackets, and the reason for this can be found in Figure 3-8. The figure shows the absolute vorticity at 10 hPa and 80 hPa on the central dates of typical SSWs in H1, NE1, NET1 and NEs1. 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 for EEI and no EEI. 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. In contrast, the 10 hPa levels in NE1 and NEs1 (panels b and d) do not show either splits or displacements. While the WAC classifies a SSW as either a split or a displacement based on the relative amplitudes of wave-1 and wave-2 geopotential height, it is clear that splits or displacements do not occur in NE1 or NEs1. Instead it seems SSWs create meridional waves when zonal EEI are turned off. Figure 3-8 also shows that splits and displacements are local reactions to wave forcings. First, splits and displacements do not occur in NEs1 even though the 80 hPa structure looks similar to that of the control run (panel e compared to h). As has already been established, the climatological tropospheric and lower stratospheric wave forcing is very similar between these two runs. Second, NET1 shows that splits and displacements do occur even when EEI are turned off in the troposphere and lower stratosphere, just as long as EEI are allowed above. The structures of the NE1 and NET1 80 hPa levels (panels f and g) are very similar while the 10 hPa levels (panels b and c) are completely different, with NET1 producing a real displacement.

The fraction of splits in H2 was 74%. When EEI are turned off in the troposphere and lower stratosphere (NEt2) the model only produces splits. While almost all climatological tropospheric wave forcing in H2 is in wave-2, the small amount of wave-1 that does exist (Figure 3-7a) is apparently enough to make about every fourth SSW a displacement. Without EEI in the troposphere and lower stratosphere this wave-1
Figure 3-8: Absolute vorticity at 10 hPa (top row) and 80 hPa (bottom row) on the central dates of an SSW in H1 (displacement; a and e), NE1 (b and f), NEt1 (displacement; c and g) and NEs1 (d and h).

Figure 3-9: Absolute vorticity at 10 hPa (top row) and 80 hPa (bottom row) on the central dates of an SSW in H2 (split; a and e), NE2 (b and f), NEt2 (split; c and g) and NEs2 (d and h).
forcing disappears (Figure 3-7e). There is almost no climatological wave-1 EP flux convergence in NEt2 (Figure 3-5e), suggesting that there is very little wave-1 mean flow interaction.

WAC gives 100% and 90% splits in NE2 and NEs2, respectively, but once again the numbers are in brackets for the same reasons as NE1 and NEs1: splits and displacements cannot occur without EEI. Figure 3-9 is the same as Figure 3-8, but for the runs with wave-2 forcing. The same conclusions can be drawn: splits and displacements are local reactions to wave-1 and wave-2 forcing, and EEI are needed for them to occur. While EEI in the troposphere and lower stratosphere affect the climatological wave forcing and hence the split and displacement ratios, low-level EEI are not necessary for splits to form.

3.5 Discussion and conclusions

In this chapter I have investigated the effects of EEI on SSW formation in an idealized GCM. I found that removal of EEI sometimes changes the SSW frequency drastically. SSWs have traditionally been thought of as interactions of anomalous tropospheric waves with the mean flow (Matsuno, 1971), and from that viewpoint it is reasonable that the SSW frequency should not change much when EEI are removed. From the perspective of more recent work, which has connected SSWs to resonance (Matthewman and Esler, 2011; Albers and Birner, 2014) and found that anomalous tropospheric wave flux is not a required for SSW formation (Birner and Albers, 2017), there is no guarantee that SSW frequencies should remain similar when EEI are removed. The results of this chapter show that EEI play an important role in SSW generation. On the one hand, removal of EEI everywhere and below 50 hPa with wave-2 tropospheric forcing (NE2 and NEt2) does not change the SSW frequency drastically, which seems to suggest SSWs are simply wave-mean flow interactions. This could be connected to the fact that the control run (H2) does not have much climatological tropospheric wave-1 forcing, which means that removal of EEI in the
troposphere and lower stratosphere does not change the wave forcing dramatically. On the other hand, removal of EEI in the upper stratosphere only (NEs2) increases the SSW frequency by 37% compared to the control run. It was found that the wave forcing and climatology of the troposphere and lower stratosphere was dependent on whether or not EEI were allowed there, and not on the upper stratosphere. Therefore, this 37% can be attributed entirely to changes in nonlinear interactions in the upper stratosphere. The SSW frequencies with wave-1 forcing are strongly dependent on EEI: even though H1 and NEs1 as well as NE1 and NEt1 have similar tropospheric and lower stratospheric wave forcings their SSW frequencies are very different, with a 53% decrease in NEs1 compared to H1 and a 45% decrease in NEt1 compared to NE1. 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 chapter show that the upper stratosphere is not a passive recipient of tropospheric and lower stratospheric wave forcing, and that upper stratospheric nonlinear processes are important for SSW generation.

How EEI affect SSW generation is not clear. Removal of EEI does not simply increase or decrease the SSW frequency: even with the same tropospheric forcing removal of EEI in the upper stratosphere can decrease SSW frequencies (H1 compared to NEs1) or increase it (NE1 compared to NEt1). However, some changes caused by removing EEI can be found in all model runs. The stratospheric wave flux, as measured by the vertical component of EP flux, is further equatorward when EEI are not allowed in the 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 stratospheric wave breaking, as measured by divergence of EP flux, also occurs further equatorward without EEI in the upper stratosphere.

It was found that EEI are required locally to create splits and displacements. Even though there are clear wave-1 and/or wave-2 structures in the 10 hPa geopotential height field during SSWs in runs without upper stratospheric EEI, the polar vortex
tends to cascade meridionally instead of splitting or being displaced. It was also found that splits and displacements occur even when EEI are not allowed in the troposphere and lower stratosphere, indicating that upper stratospheric EEI are responsible for split or displacement formation. It therefore seems that displacements and splits can be considered local nonlinear reactions to climatological wave-1 and wave-2 forcings, respectively.

With wave-2 forcing, all SSWs were splits when EEI were turned off in the troposphere and lower stratosphere (NEs2). This is likely due to the fact that all tropospheric forcing is in wave-2 when EEI are turned off in the lower levels, and the resulting SSWs are therefore splits. The fraction of splits increased when EEI were turned off in the lower levels with wave-1 forcing: 80% splits in NEt1 compared to 59% in H1. This is despite the fact that H1 has strong climatological tropospheric wave-1 and wave-2 wave flux, and NEt1 has almost only wave-1. The wave-2 forcing required to produce these splits has its origin in the transition region between no-EEI and EEI-allowed, where some of the wave-1 flux is transferred to wave-2. 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. It is possible that the split and displacement classification (WAC) skews the results towards more splits compared to the control run. The parameter $k = 1.0$ in Equation 2.2 was chosen because WAC and subjective analysis matched well with this value in the control run and in ERA-Interim reanalysis data (Dee et al., 2011), but this may no longer be the case in the mixed runs. Nevertheless, NEt1 produces a large amount of splits, showing that the wavenumber of the tropospheric wave flux is not necessarily the most important factor in determining the type of SSW produced. Interestingly, removing EEI everywhere or partially seems to increase the relative strength of wave-2 geopotential height anomalies (and hence split fractions) with both wave-1 and wave-2 forcing. The only exception was NE1, where the fraction of “splits” (57%) remained practically identical to that of the control run (59%). It should be reiterated, however, that splits and displacements do not occur without EEI.
locally, and that the 57% of "splits" in NE1 simply showed that wave-2 geopotential height anomalies were larger than wave-1 geopotential height anomalies at some point around 57% of SSWs.

In this chapter I have demonstrated that removing EEI can strongly influence SSW frequencies, even when the climatological wave forcing remains the same. This shows that SSWs are not simply wave-mean flow interactions, and that nonlinear interactions in the upper stratosphere play an important role in SSW generation and stratospheric dynamics. It was found that EEI are required locally for splits and displacements to form, but that splits and displacements can occur even if EEI are not allowed in the troposphere and lower stratosphere. This indicates that displacements and splits can be considered local reactions to wave-1 and wave-2 forcings.
Chapter 4

Effects of tropical wind variability on the stratospheric polar vortex

4.1 Introduction

In my analysis of the perpetual Northern Hemisphere winter runs forced with wave-2 heating perturbations I found tropical wind oscillations in the stratosphere. The amplitudes of the oscillations were 10-15 m/s below 10 hPa and up to 30 m/s above 8 hPa. The middle stratospheric magnitude of the oscillations is similar to that of the QBO (~ 15 m/s; see Figure 1-5). An example of these wind oscillations can be found in Figure 4-1, which shows the 10°S-10°N deviations from the climatological zonal mean zonal wind below 8 hPa during 3000 days of an extended version of the H2 model used in Chapters 2 and 3. Despite the fact that these wind oscillations occur in the tropical stratosphere there are two major differences between them and the QBO observed in the real atmosphere. First, the oscillations are not as periodic as the QBO: the timescale of the oscillations range from a few hundred days to well over a thousand days. Second, the oscillations start in the lower stratosphere and propagate upward, unlike the QBO that starts around 5 hPa and propagates downward. The oscillations seem to start around 50 hPa, which coincides with the area where the
Figure 4-1: 10°S-10°N anomalies from the climatological zonal mean zonal wind during 3000 days of H2. The anomalies are not scaled by latitude. Notice that the highest level of this figure is just above 8 hPa.

climatological zonal mean zonal wind approaches zero (Figures 2-2b or 3-3a). Like the QBO, these oscillations are driven by waves: in model runs without tropospheric wave forcing the oscillations do not appear.

Apart from these tropical wind oscillations I also observed other periodic behavior on timescales that were longer than one expects in a perpetual winter model. For one, the autocorrelation function (ACF) of zonal mean zonal wind at 10 hPa and 60°N exhibited periodic behavior with a period of just below 1000 days (see Section 4.3.1). This oscillation does not exist in model runs without any tropospheric forcing. Furthermore, while the SSW frequency of H2 was found to be around one SSW per 200 days there were intervals where no SSWs were observed for up to 2000 days. Long intervals between SSWs have been observed in the real atmosphere: there were nine consecutive Northern Hemisphere winters (90/91 through 98/99) without SSWs based on my analysis of the ERA-Interim dataset in Chapter 2. If the periodicity of the ACF and the long intervals without SSWs are real signals they must be forced by periodic behavior produced in the model itself, given that there is no induced
time-dependency in H2 (seasonal or otherwise).

In this chapter I use the idealized model with wave-2 tropospheric heating to investigate the sources and possible effects of the tropical wind oscillations. The advantage of this model compared to observations or more advanced climate models is its simplicity: equatorial wind variations can be damped away and long model runs can be performed at low cost. The three major questions that I will answer in this chapter are:

1. What causes the stratospheric tropical wind oscillations, and what causes their upward propagation?

2. Do the stratospheric tropical wind oscillations affect midlatitude winds, and if so, through which mechanism?

3. How do the results of this model relate to the connections between the tropical and midlatitude winds in the observed atmosphere?

Section 4.2 describes the methods used to remove tropical wind variations, and the statistical methods used to evaluate differences relating to SSWs. Section 4.3 contains the results obtained with model runs performed under perpetual Northern Hemisphere winter conditions, as well as a proposed mechanism for how the oscillations are formed and propagate. In Section 4.4 I investigate the effects of tropical wind damping in model runs with a seasonal cycle. Both the Northern and Southern Hemisphere are investigated. A discussion of the results can be found in Section 4.5.

4.2 Methods

The model used in Chapters 2 and 3 was used in this chapter as well. The wave-2 heating perturbation was used to induce Northern Hemisphere winter-like stratospheric variability. No tropospheric forcing was applied to the Southern Hemisphere. In the perpetual Northern Hemisphere winter runs the model setup is identical to that of H2
in Chapters 2 and 3, but the model run was extended to 71,100 days. Two additional runs of equal length but with tropical wind damping were also used. Only the zonal wind output was saved from these extended model runs, so any results showing EP fluxes come from the 31,100 day long H2 run from Chapters 2 and 3. Once again the first 1100 days were discarded from all runs.

To investigate the effects of tropical wind variability in the presence of seasonal variations, a 360 day long stratospheric seasonal cycle was added to the model. The seasonal cycle is identical to the one used by Sheshadri et al. (2015). There is no seasonal cycle below 200 hPa. The seasonal runs were 80,000 days long and the first 1170 days were removed (90 days plus three years), leaving 78830 days (almost 219 years) for analysis. In addition to a control run, a run of equal length with moderately damped tropical winds was also used.

4.2.1 Tropical wind damping

The model runs without damping were performed first, after which the time mean zonal wind fields were calculated. These model runs will be referred to as the control runs, and the Northern Hemisphere perpetual winter control run and seasonal cycle control run will hereafter be referred to as NHC and SC, respectively. In the case of SC a 360-day seasonal mean zonal wind field was calculated.

To reduce tropical wind variability a damping term was added to the zonal wind tendency of the model. The variability was damped toward the means of the control runs, and applied around the equator and above 100 hPa. The format of the damping was

\[
\frac{\partial u_{damp}}{\partial t} (\lambda, \phi, p) = \begin{cases} 
-\frac{1}{\tau} \exp \left( -\frac{\phi^4}{2c^2} \right) \tanh \left( \frac{p_1 - p}{p_2} \right) (u (\lambda, \phi, p) - u_{mean} (\lambda, \phi, p)), & p < p_1 \\
0, & \text{otherwise,}
\end{cases}
\]  

(4.1)

where \(\lambda\) is longitude, \(\phi\) is latitude, \(p\) is pressure, \(c = 0.04\), \(p_1 = 1.0 \times 10^4\) Pa, \(p_2 = 2.0 \times 10^3\)
Figure 4-2: The vertical and latitudinal extent of the equatorial wind damping.

Pa, and \( \tau \) is the amplitude of the damping. The two different damping amplitudes used for the perpetual winter run were \( \tau = 1/100 \) days\(^{-1} \) and \( \tau = 1/40 \) days\(^{-1} \). The damped seasonal run used \( \tau = 1/100 \) days\(^{-1} \). For comparison, the radiative relaxation timescale is 40 days in the stratosphere of this model. In the case of the damped seasonal run the term \( u_{\text{mean}}(\lambda, \phi, p) \) was \( u_{\text{mean}}(\lambda, \phi, p, i) \), where \( i \) is one day of the seasonal mean. The damped runs will hereafter be referred to as NH100, NH40, and S100; for the perpetual winter runs with 100 day and 40 day damping, and the seasonal run with 100 day damping. The extent of the damping terms can be seen in Figure 4-2, which shows the product of the exponential and hyperbolic tangent parts of Equation 4.1.

4.2.2 Statistical methods for SSWs

A Monte Carlo approach was used to assess the statistical significance of differences associated with SSWs in the perpetual Northern Hemisphere winter model runs. After calculating the SSW frequency of a given model run 10,000 datasets of length
equal to the number of days in the model run were created. Each dataset consisted of ones and zeros, where the values were assigned randomly with a probability $p(1) = \frac{\text{number of SSWs}}{\text{length of dataset}}$. From the 10,000 datasets the standard deviation of number of SSWs (i.e., the number of ones) and the variability of intervals between SSWs could be estimated. These estimates assume that SSWs occur at random, and that the probability of producing a SSW is not affected by previous entries. Unless otherwise stated, the method described above is the source for statistical significance estimates dealing with SSWs in this chapter.

4.3 Results under Northern Hemisphere winter conditions

4.3.1 Changes in zonal mean zonal wind and ACF

The effect of the tropical wind damping can be seen in Figure 4-3. The left panels of the figure show 5000 days of NHC $10^\circ$S-$10^\circ$N zonal mean zonal wind deviations from the climatological mean. The top left panel shows the deviations for the full depth of the stratosphere. The deviations are largest in the uppermost stratosphere, so the bottom left panel shows the deviations up to 8 hPa (like Figure 4-1) to emphasize the structure near the source of the oscillations. The left panels show how the tropical zonal mean wind oscillations have their source in the lower stratosphere ($\sim 60$-70 hPa), and how they propagate upward with time. Oscillations formed in the lower stratosphere propagate up to the top of the model in 1000 to 2000 days.

The right panels show upper stratospheric (mean from 1 to 8 hPa; top right) and middle stratospheric (mean from 10 to 44 hPa; bottom right) zonal mean wind deviations for the three perpetual winter runs. The black lines show NHC deviations, and the same 5000 days as in the left panels are shown. The panels show how the long term (order 1000 day) oscillations disappear with damping. The differences between
Figure 4-3: Left: 5000 days of NHC 10°S-10°N zonal mean zonal wind deviations from the climatological mean for the full depth of the stratosphere (top) and up to 8 hPa (bottom). Top right: 5000 days of 10°S-10°N zonal mean zonal wind deviations averaged between 1 and 8 hPa for NHC (black line), NH100 (red line) and NH40 (blue line). Bottom right: same as top right but averaged between 10 and 44 hPa. The left panels and black lines show the same 5000 days. The means are not weighted by pressure or latitude.
Figure 4-4: Cross-correlations between 10°S-10°N zonal mean zonal wind at a given pressure level and the 10°S-10°N zonal mean zonal wind averaged between 28 and 44 hPa for NHC (a), NH100 (b) and NH40 (c). Notice the shorter lag axes for NH100 and NH40. Positive lag means that the 28-44 hPa winds lag the winds at that pressure level. The means are not weighted by pressure or latitude.

NH100 and NH40 are small compared to the differences between the damped runs and the control run. NH100 and NH40 show sharp drops in upper stratospheric tropical wind strength. These drops are associated with SSWs: the upper stratospheric tropical winds are strongly and almost instantaneously correlated with the winds at 10 hPa and 60°N (see Figure 4-7 further down in this section). This correlation exists in the control run as well: the sharp drop around 14,400 days, for example, is associated with an SSW. Variations associated with phenomena less extreme than SSWs are largely damped away in NH100 and NH40, but not in the control run.

The tropical wind variations in the middle stratosphere are not directly associated with SSWs (see below), and the variations in NH100 and NH40 are therefore more heavily damped here than in the upper stratosphere, when compared to the variations in the control run. Comparisons of the NHC wind variations in the upper and middle stratosphere of this figure suggests an anti-correlation between the two. This anti-correlation can be clearly seen in Figure 4-4, which shows the cross-correlations between lower stratospheric equatorial winds and the equatorial winds at any given pressure level in the stratosphere. In Figure 4-4 a positive lag means that the equatorial winds at 28-44 hPa lag the equatorial winds at a given pressure level. As can be expected, the cross-correlation maximizes around zero lag in the lower stratosphere,
Figure 4-5: ACF of zonal mean zonal wind at 10 hPa and 60°N (blue line) in NHC (a), NH100 (b) and NH40 (c), plus and minus three standard deviations (red lines). See text for details.

since that shows the instantaneous autocorrelation of lower stratospheric winds. The upward propagating anomalies seen in the control run in Figures 4-1 and 4-3 can be seen in panel a: anomalies in the lower stratosphere are correlated with anomalies of the same sign in the upper stratosphere 1000 to 2000 days later. However, this correlation is weaker than the instantaneous anti-correlation of lower and upper stratospheric winds. This anti-correlation is visible in all panels, although it is weaker in the damped runs. The x-axes are only a quarter of the length for NH100 and NH40 (panels b and c, respectively) compared to the control run. There are practically no correlations beyond 500 days for NH100, and for NH40 the correlations are mostly gone within 200 days. This shows that the equatorial wind damping removes the memory in stratospheric equatorial winds.

Figure 4-5 shows the ACF of zonal mean zonal wind at 10 hPa and 60°N in NHC, NH100 and NH40, plus and minus three standard deviations. The standard deviations were estimated with the assumption that the theoretical ACF goes to zero after $numD = 500$ days. The figure shows a clear oscillating pattern of just below 1000 days for NHC (panel a), where the statistical significance of the first peak exceeds three standard deviations. A peak similar to the first one in NHC can be seen the moderately damped run (NH100; panel b), although it is weaker and appears about 200 days earlier. The statistical significance of this peak exceeds two standard deviations (not shown), but not three. The statistical significances of these peaks exceed the thresholds mentioned above for any choice of $numD$ between 0 and

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Figure 4-6: Zonal mean zonal wind at 10 hPa and 60°N in bins of 5 m/s for NHC (black), NH100 (red) and NH40 (blue). See text for details.

2999 (not shown). No significant peak is visible in the strongly damped run (NC40; panel c). Overall, this indicates that there is periodic behavior in the stratospheric Northern Hemisphere midlatitudes in the control run, but that this periodic behavior disappears when stratospheric tropical wind variability is suppressed.

Since the tropical winds are damped towards the climatological mean of the control run the zonal mean zonal wind fields look very similar in most parts of the atmosphere (not shown). However, there are differences in the distributions of the midlatitude winds between the model runs. Figure 4-6 shows the zonal mean zonal wind distribution at 10 hPa and 60°N for the three runs, where the daily mean wind has been sorted into bins of 5 m/s. Weak winds (< 30 m/s) are more common in the damped runs compared to the control run, while strong winds (≥ 60 m/s) are more
Figure 4-7: Cross-correlations between 10°S-10°N zonal mean zonal wind at a given pressure level and the zonal mean zonal wind at 10 hPa and 60°N for NHC (a), NH100 (b) and NH40 (c). Notice the shorter lag axes for NH100 and NH40. Positive lag means that 10 hPa and 60°N winds lag the equatorial winds at that pressure level. The equatorial winds are not weighted by latitude.

common in the control run. The shift is rather large: there are 13,559 days with zonal mean zonal winds at 10 hPa and 60°N equal to or exceeding 60 m/s in NHC, while the corresponding numbers are 11,652 and 10,295 for NH100 and NH40, respectively, corresponding to 14% and 24% decreases.

Figure 4-7 shows the cross-correlations between zonal mean zonal winds at the equator and the winds at 10 hPa and 60°N, which is used as a proxy for polar vortex strength. Like in Figure 4-4 the lag axes for NH100 and NH40 are only a quarter of the length of the axis for NHC. Positive lag means that the polar vortex winds lag the equatorial winds. All panels in Figure 4-7 show a strong instantaneous correlation between the upper stratospheric equatorial winds and the winds at 10 hPa and 60°N, which is the reason for the sharp drops in upper stratospheric equatorial winds speeds around SSWs seen in Figure 4-3. However, Figure 4-7 also shows a positive correlation with positive lag between upper stratospheric equatorial winds and the polar vortex strength, where the equatorial winds lead the winds at 10 hPa and 60°N by up to about 300 days. This correlation is there for all three runs, but it should be noted that any influence on the polar vortex strength by equatorial winds will be smaller in magnitude in the damped runs compared to the control run, since the equatorial wind variations are much smaller in the damped runs (see top right panel of Figure
4-3). In the control run (panel a) there is also a positive correlation with negative lag, although that disappears when equatorial wind damping is applied. In the most heavily damped run there are practically no correlations in negative lag. The positive correlation between upper stratospheric equatorial winds and polar vortex strength in the control run indicates that the same mechanisms responsible for modulating the polar vortex strength are also modulating the upper stratospheric equatorial winds. This does not exclude indirect effects associated with changing either the polar vortex strength or the equatorial wind strength: a change in either could influence the other region by changing the conditions for wave propagation and convergence. The reason why positive correlations at negative lag disappear could be that the influence of SSWs on upper stratospheric equatorial winds is very heavily damped. A SSW will decelerate the polar vortex rapidly and strongly, which will alter the conditions for wave propagation and wave convergence in the region greatly. This would affect the strength of upper stratospheric equatorial wind strength as well. However, Equation 4.1 shows how the damping is negatively and linearly correlated with wind anomalies, and the strong equatorial wind anomalies associated with SSWs in the control run will be weaker and disappear faster in the damped runs due to the heavy damping.

Figure 4-7 also shows a negative middle stratospheric correlation with negative lag in the control run, which could be interpreted as an influence of the polar vortex on the middle stratospheric equatorial region. However, no evidence for such an influence has been found. The mean wind speed in this region is not affected by SSWs in the control run (not shown), and SSWs cause the most dramatic changes in polar vortex strength. The fact that the negative anomaly largely disappears with damping (panels b and c) further indicates that it is not caused by SSWs. Another explanation for the negative correlation will be presented in the subsection below.

Figure 4-7 also indicates that the equatorial upper stratosphere may be the source of the oscillations in the ACF of zonal mean zonal wind at 10 hPa and 60°N seen in NHC in Figure 4-5: panel a in Figure 4-7 shows two weak local maxima of positive correlation centered around positive and negative lags of 1000 days. These peaks
are not present in 2000 day-lag cross-correlation figures of the damped runs (not shown). It therefore seems that it is the damping of the upper stratospheric equatorial winds that is responsible for the disappearing periodicity of the ACF of polar vortex strength.

4.3.2 Mechanisms behind the connections between tropical winds and polar vortex strength

The results of the control run and damped runs above show that tropical wind oscillations propagate upward, and that tropical wind variability in the upper stratosphere may influence the strength of the polar vortex. The QBO in the observed atmosphere is wave-driven, and it is reasonable to assume that the oscillations found in this model are as well. Therefore, to understand the upward propagation of the anomalies and the influence of the anomalies on the dynamics in the stratosphere one must consider the variability of wave activity in the model. Since the extended runs used in the subsection above only saved zonal wind output, the H2 model run from Chapters 2 and 3 will be used to explain the mechanisms behind tropical wind oscillations and their influence on the polar vortex. Chapters 2 and 3 used the quasi-geostrophic (QG) approximation for EP flux, but since we are interested the equatorial wave fluxes the full EP flux will be calculated in this chapter. The equations for EP flux were based on Andrews et al. (1983) and are:

\[
F(\phi) = \cos \phi \left( -\nu' \frac{\partial u'}{\partial p} + \frac{\nu' \theta'}{\partial_p} \frac{\partial \bar{u}}{\partial p} \right), \quad (4.2a)
\]

\[
F(p) = \cos \phi \left( -u' \omega' - \frac{\nu' \theta'}{\partial_p} \left[ \frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} \left( \bar{u} \cos \phi \right) - f \right] \right), \quad (4.2b)
\]

\[
\nabla \cdot \vec{F} = \frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} \left( F(\phi) \cos \phi \right) + \frac{\partial}{\partial p} F(p). \quad (4.3)
\]
\( \omega \) is vertical velocity in pressure coordinates, and the notation in the equations above is the same as for Equations 2.3 and 2.4. As before, the divergence of EP flux will be displayed as \( \frac{1}{\cos \phi} \nabla \cdot \vec{F} \). Zonal mean zonal wind acceleration and residual circulation were also calculated. The equations for these quantities, once again based on Andrews et al. (1983), are:

\[
\frac{\partial \vec{u}}{\partial t} = \left[ -\left( \frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} \left( \bar{u} \cos \phi \right) - f \right) \bar{v}^* - \frac{\partial \bar{u}}{\partial p} \omega^* \right] + \frac{1}{\cos \phi} \nabla \cdot \vec{F},
\]

(4.4)

\[
\bar{v}^* = \bar{u} - \frac{\partial}{\partial p} \frac{\bar{u} \bar{\theta}}{\bar{\theta}_p},
\]

(4.5a)

\[
\omega^* = \bar{\omega} + \frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\bar{u} \bar{\theta} \cos \phi}{\bar{\theta}_p} \right),
\]

(4.5b)

\[
\frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} \left( \bar{v}^* \cos \phi \right) + \frac{\partial \omega^*}{\partial p} = 0,
\]

(4.6)

where \( \bar{v}^* \) and \( \omega^* \) are meridional and vertical residual circulation, respectively. Non-conservative forces have been omitted in Equation 4.4.

Figure 4-8 shows the climatological zonal mean zonal wind, vertical EP flux and divergence of EP flux in H2, northward of 12°S. The vertical EP flux was scaled by pressure in the same way that Figures 3-6 and 3-7 were. The equatorial region where the zonal mean zonal wind approaches zero m/s can be seen in panel a. The region is centered around 50 hPa, and the oscillations seen in Figure 4-3 seem to originate just above it. The divergence of EP flux in the stratosphere (Figure 4-8c) shows that there is large EP flux convergence from about 15°N to poleward of 70°N, but also that there is weak EP flux divergence extending from the equator to about 30°N just above the tropopause as well as in the upper equatorial stratosphere. Equation 4.4 shows how EP flux divergence relate to zonal mean zonal wind acceleration and residual circulation, and the fact that the climatological lower stratospheric subtropical region contains both EP flux divergence and convergence indicates that this could be an
Figure 4-8: Climatological zonal mean zonal wind (a), vertical EP flux scaled by pressure (b), and divergence of EP flux (c), in H2. Panel a shows the full depth of the atmosphere while panels b and c only show the stratosphere.

An important region for lower stratospheric tropical wind oscillations.

The climatological vertical EP flux (Figure 4-8b) shows that although most vertical flux is located poleward of 40°N, there is still a considerable amount of vertical EP flux in the lower stratospheric subtropical region. Since the flux is scaled by pressure, the flux at the upper pressure levels looks larger than it actually is, and the strongest vertical wave flux is located in the lower stratosphere. The interactions between the stratospheric zonal mean flow and vertical EP flux are quite different depending on where the vertical EP flux is originating from. Figure 4-9 shows the cross-correlations between vertical EP flux averaged between 18°N and 32°N as well as 40°N and 80°N, and zonal mean zonal wind at the equator and 1-8 hPa along with zonal mean zonal wind at 10 hPa and 60°N.

The figure shows how flux anomalies between 40°N and 80°N lead to changes in stratospheric zonal mean zonal wind shortly after, and the influence is particularly strong on the polar vortex strength (panel d). The strongest correlation between the flux and polar vortex strength occurs at lags of 10 to 20 days and around 100 hPa, and shows how strong (negative) vertical flux anomalies in this region can lead to strong polar vortex decelerations such as SSWs. This is why the flux at 100 hPa is often highlighted as especially important for extreme events in the stratospheric polar vortex region (Polvani and Waugh, 2004; Birner and Albers, 2017). The influence of 40°N to 80°N vertical flux on upper stratospheric equatorial winds is slightly weaker.
Figure 4-9: Cross-correlations between vertical EP flux and stratospheric zonal mean zonal winds. Top row: vertical EP flux and 10°S-10°N zonal mean zonal wind averaged between 1 and 8 hPa. The wind is not weighted by latitude or pressure. Bottom row: vertical EP flux and zonal mean zonal wind at 10 hPa and 60°N. Left column: flux averaged between 18°N and 32°N. Right column: flux averaged between 40°N and 80°N. Positive lag means that flux anomalies lead wind anomalies. Notice the different limits of the colorbars.
(panel b). It is clear that the vertical flux in this region has a larger effect on the stratospheric zonal wind than the other way around: even though there are positive correlations at negative lags the correlations are much weaker than those at positive lags.

The cross-correlations for vertical EP flux between 18°N and 32°N and stratospheric zonal mean zonal winds are much more symmetric about zero lag. While flux anomalies between 40°N and 80°N clearly lead to changes in zonal mean zonal wind strength in the stratosphere, subtropical flux anomalies lead zonal wind strength anomalies and lag the zonal mean anomalies to similar degrees. The strongest correlations are found slightly higher up in the stratosphere at around 50 to 60 hPa. The correlation between subtropical flux anomalies and upper stratospheric equatorial wind extends longer for positive lags than negative lags (panel a), but the magnitude is similar on both sides of zero lag. It is worth noting that the correlation is slightly stronger compared to that found in panel b: this could indicate that subtropical flux anomalies exert a slightly larger influence on upper stratospheric equatorial winds than flux anomalies between 40°N and 80°N. The correlation between subtropical flux and polar vortex strength is very symmetric about zero lag (panel c), and its maximum value is lower than that found in panel d.

Overall, Figure 4-9 shows that midlatitude and polar vertical flux anomalies affect stratospheric zonal mean zonal winds but are less affected by zonal wind anomalies, while the interaction between subtropical flux anomalies and stratospheric zonal mean zonal wind anomalies is more two-sided: a strong subtropical flux will lead to weak stratospheric zonal winds, but weak zonal winds will also lead to strong subtropical flux. This does not necessarily indicate that the zonal mean zonal winds in these regions affect the subtropical flux directly, but could indicate that the same mechanism modulates both the flux and the zonal winds.

Given the correlation between subtropical wave flux and polar vortex strength seen in Figure 4-9c, one can expect that the cross-correlation between this wave flux and
Figure 4-10: Cross-correlations between 10°S-10°N zonal mean zonal wind at a given pressure level and the vertical EP flux at 66 hPa and 18°S-32°N (a) and the divergence of EP flux at 54 hPa and 18°S-32°N. Negative lag means that the 10°S-10°N winds lag the fluxes at that pressure level. The equatorial winds are not weighted by latitude, but the fluxes are. Notice the different limits of the colorbars.

The equatorial winds should be similar to the cross-correlation between polar vortex strength and equatorial winds seen in Figure 4-7a. This is indeed the case: Figure 4-10 shows the cross-correlations for subtropical vertical EP flux at 66 hPa and divergence of EP flux at 54 hPa. Vertical EP flux is strongly, positively and instantaneously correlated to divergence of EP flux at pressure levels above (not shown), and the similarity between the two panels reflect that. The fact that the correlation is slightly weaker for divergence of EP flux is due to the fact that not all anomalous vertical EP flux converges immediately above the vertical flux anomaly. The correlations seen in Figure 4-10 are not particularly sensitive to the pressure levels or latitudes chosen, as long as the pressure levels are between about 100 and 30 hPa and the latitude is between about 10°N and 40°N.

From Figure 4-7a one might think that changes in polar vortex strength alters the strength of the lower stratospheric tropical winds, but Figure 4-10 suggests that the modulation of lower stratospheric equatorial wind strength may be in part due to changing wave forcing in the lower stratospheric subtropics, which in itself is modulated by the polar vortex strength (Figure 4-9c) and upper stratospheric equatorial
winds (Figure 4-9a). The latter is likely why the negative correlation in the lower stratosphere shows up in Figure 4-7a. While the correlation between these fluxes and lower stratospheric equatorial winds is lower than that found in Figure 4-7a by a factor of about 2, the correlation between the fluxes and upper stratospheric equatorial winds is lower by a factor of about 3. This shows that the relative importance of lower stratospheric subtropical fluxes may be higher for lower stratospheric wind anomalies compared to upper stratospheric wind anomalies.

To understand how and why the equatorial wind oscillations form and propagate upward one can investigate the divergence of EP flux, residual circulation, and zonal mean zonal wind acceleration during times of anomalously strong and anomalously weak equatorial winds. This is done in Figures 4-11 and 4-12, which show anomalies during times when the zonal mean zonal wind at 10°S-10°N and 28-44 hPa is weaker and stronger than the mean by one standard deviation. The zonal mean zonal wind was not weighted by latitude or pressure. The residual circulation was obtained by calculating $\vec{V}^*$ from Equation 4.5a and $\vec{w}^*$ from Equation 4.6. The zonal mean zonal wind acceleration was calculated from Equation 4.4.

Figure 4-11 shows changes in residual circulation and $\partial u/\partial t$ during times when $u$ at 10°S-10°N and 28-44 hPa is anomalously easterly (panel a) and westerly (panel b). The upward propagation of the anomalies can be seen in the structure of the changes in $\partial u/\partial t$: the zonal mean zonal wind anomalies have acceleration anomalies of the same sign above their centers, while the acceleration anomalies at and below the centers are weaker or of the opposite sign of the zonal wind anomaly. It can be seen that the changes in residual circulation are substantial: an anomaly in the lower stratospheric equatorial winds is associated with an anomaly of the same sign in meridional residual circulation in the lower stratospheric subtropics. The structures of the anomalies themselves are partly responsible for the upward propagation of the anomalies. When the anomaly is westerly $\partial u/\partial p$ increases above the anomaly and decreases below the anomaly. While the strength of $\vec{w}^*$ varies with varying tropical winds (see below), it is always negative. Therefore, based on Equation 4.4, a westerly anomaly will
help induce acceleration above the peak of the anomaly and deceleration below. The induced accelerations are of opposite signs when the wind anomalies are easterly. While the strength of the residual circulation is different from the climatological mean when equatorial wind anomalies are present (see below), the effect of changes in $\frac{\partial \bar{u}}{\partial p}$ seem to be a major factor in the upward propagation of these anomalies.

As was seen in Figure 4-4a, the lower to middle stratospheric equatorial winds are anti-correlated to the upper stratospheric equatorial winds. Figure 4-12 shows the same things as Figure 4-11 but throughout the depth of the stratosphere, and with divergence of EP flux instead of zonal mean zonal wind acceleration. Unlike Figure 4-11, the climatological fields are also shown. Climatologically, there is EP flux convergence throughout most of the stratosphere (panels a and d), which induces poleward meridional circulation and deceleration of the mean flow. When the lower stratospheric equatorial winds are anomalously easterly, and therefore the upper stratospheric equatorial winds anomalously westerly, there is a substantial decrease in EP flux convergence throughout the stratospheric midlatitudes (Figures 4-12b and e). There is an increase in convergence of EP flux around the core of the polar vortex in the upper stratosphere, and a decrease in EP flux divergence poleward and upward.
of this increase. This could be due to changes in wave propagation caused by changes in the zonal mean zonal wind structure (Holton and Tan, 1980). This decrease in EP flux convergence will weaken the poleward meridional circulation and the deceleration of midlatitude winds. The positive zonal mean zonal wind anomalies in the upper stratospheric equatorial region coincide with an unusually strong polar vortex, as can be expected from Figure 4-7a. When the zonal mean zonal winds in the lower stratospheric equatorial region are anomalously westerly (Figures 4-12c and f) the signs of the anomalies are opposite to those found in Figures 4-12b and e: there is an increase in EP flux convergence in the midlatitudes, and the meridional residual circulation is strengthened.

Figure 4-13 shows the same thing as Figure 4-11 but for means composited on upper stratospheric anomalies instead of lower stratospheric anomalies. A similar pattern can be found: zonal mean zonal wind anomalies have acceleration anomalies of the same sign at and above their centers and acceleration anomalies of opposite sign in other areas, indicating an upward propagation. The upper stratospheric equatorial wind anomalies are caused by a combination of changes in wave forcing, residual circulation and zonal wind gradients. First, Figures 4-12b and c show that there are EP flux divergence anomalies of the same sign as the zonal wind anomalies in the upper stratospheric equatorial region. This indicates that at least part of the anomaly may be caused by changes in wave convergence. Second, upper stratospheric equatorial wind anomalies are associated with rather large changes in residual circulation. When the equatorial wind anomalies between 1 and 8 hPa exceed one standard deviation the strength of the residual circulation in this region weakens by $\sim 10\%$. The decrease in the strength of the meridional residual circulation is one of the contributing factors to the acceleration of upper stratospheric equatorial wind strength, as can be seen in Equation 4.4. Third, the altered zonal wind gradients will affect the zonal wind anomalies. The westerly anomaly acts to increase (decrease) $\frac{\partial \overline{u}}{\partial p}$ above (below) the center of the positive anomaly, and decrease (increase) $\frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (\overline{u} \cos \phi)$ in the Northern (Southern) Hemisphere, which will help decelerate (accelerate) the posi-
Figure 4-12: Left: climatological $\bar{u}$ (magenta contours), $\frac{1}{\cos \phi} \nabla \cdot \bar{F}$ (filled contours) and residual circulation (green lines) in H2. Middle columns: deviations from the climatological $\bar{u}$ (magenta contours) and $\frac{1}{\cos \phi} \nabla \cdot \bar{F}$ (filled contours), and full residual circulation (green lines) when $\bar{u}$ at 10ºS-10ºN and 28-44 hPa is more than one standard deviation below the mean. Right columns: same as middle column but when $\bar{u}$ at 10ºS-10ºN and 28-44 hPa is more than one standard deviation above the mean. Top rows show the upper stratosphere, and bottom rows the lower stratosphere. Negative zonal mean zonal wind values are dashed. The right hand side colorbars pertain to the deviations.
Figure 4-13: Deviations from the climatological $\bar{u}$ (magenta contours) and $\partial \bar{u}/\partial t$ (filled contours), and full residual circulation (green lines) when $\bar{u}$ at 10°S-10°N and 1-8 hPa is more than one standard deviation above the mean (a) and more than one standard deviation below the mean (b). Negative zonal mean zonal wind values are dashed.

The lower stratospheric anomalies are also likely to be caused in part by changes in upwelling. The lower stratospheric anomaly starts around 60 hPa (see, for example, Figure 4-10), and this is the location where $\partial \bar{u}/\partial p$ changes sign (see Figure 4-8a). Figures 4-12c and f showed that strong EP flux convergence coincides with a weak polar vortex and anomalously easterly (westerly) upper (lower) stratospheric equatorial winds. The strong EP flux convergence increases the upwelling in the stratospheric equatorial region, and this will lead to acceleration where $\partial \bar{u}/\partial p$ is positive and deceleration where $\partial \bar{u}/\partial p$ is negative, since $\bar{w}^*$ becomes more negative. However, the cross-correlation figures show that changes in lower stratospheric equatorial winds lag changes in polar vortex strength and upper stratospheric equatorial winds, and that the anomalies tend to last slightly longer than upper stratospheric wind anomalies (Figure 4-7a). The different structures of the cross-correlations make it unlikely that both the lower and upper anomalies would be governed by the same mechanisms. Instead, it seems the lower stratospheric anomalies are induced partly by changes in upwelling, but that lower
stratospheric subtropical flux anomalies play a bigger role in creating and maintaining the lower stratospheric anomaly compared to the upper stratospheric anomaly, based on Figure 4-10.

The top right panel in Figure 4-3 shows how the upper stratospheric equatorial winds become strongly easterly during SSWs even when the tropical winds are damped. It is of interest to understand why this response remains even in the presence of damping. Figure 4-14 shows anomalous zonal mean zonal wind, divergence of EP flux and zonal mean zonal wind acceleration when the zonal mean zonal wind at 10°S-10°N and 1-8 hPa is more than two standard deviations below the mean. Such large deviations in upper stratospheric equatorial wind strength coincide with a weakening of the polar vortex strength by about 50 m/s. Since the mean strength of the polar vortex is about 55 m/s this shows that these anomalies are associated with SSWs. Figure 4-14a shows that a strong decrease in EP flux convergence throughout the upper stratosphere occurs when the polar vortex has collapsed. This decrease in divergence of EP flux will lead to an anomalous acceleration of the zonal mean zonal winds in
the extratropics, which can be seen northward of 20°N in Figure 4-14b. Figure 4-14b also shows the strong deceleration of zonal mean winds in the upper stratospheric equatorial region that we can expect from the cross-correlations between equatorial winds and polar vortex strength seen in Figure 4-7. Since the divergence of EP flux does not change much in the equatorial region, Equation 4.4 tells us that this deceleration must be a result of changes in residual circulation and zonal mean zonal wind gradients. The deceleration is likely a two-step response: first, a strong increase in midlatitude convergence of EP flux will initiate the SSW and initially increase the strength of the overturning circulation. When the convergence of EP flux in the stratosphere and at 40°N-60°N is more than two standard deviations above the mean the strengths of \( \vec{v}^* \) and \( \vec{\omega}^* \) in the upper stratospheric equatorial region both increase: \( \vec{v}^* \) increases by a factor of about 4.5 and \( \vec{\omega}^* \) by a factor of about 2.5. This initial increase in the strength of meridional residual circulation will, in the absence of changes in zonal wind gradients and divergence of EP flux, decelerate the zonal winds in the upper stratospheric equatorial region according to Equation 4.4, since 
\[
\frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (\bar{u} \cos \phi)
\]
is positive throughout the upper stratosphere. The strengthening of the vertical residual circulation will decrease (increase) the magnitude of this change below (above) about 2 hPa, since \( \frac{\partial \bar{u}}{\partial p} \) is positive (negative) there (see Figure 4-8a). The result is that the deceleration is strongest above 2 hPa. The second step of the response is what can be seen in Figure 4-14. At this time the polar vortex has already collapsed and the residual circulation is weakened due to the decrease in wave driving in midlatitudes. During times when the zonal mean zonal wind at 10°S-10°N and 1-8 hPa is more than two standard deviations below the mean the change in upper stratospheric equatorial \( \vec{\omega}^* \) is only few percent compared to the mean. The strength of \( \vec{v}^* \), however, is about 25% below the mean. If zonal wind gradients did not change this should lead to an acceleration of the zonal mean winds. However, the easterly anomaly already present in the region acts to increase \( \frac{\partial \bar{u}}{\partial p} \) below the center of the anomaly and decrease 
\[
\frac{1}{\cos \phi} \frac{\partial}{\partial \phi} (\bar{u} \cos \phi)
\]
in the Southern Hemisphere, which will help maintain the easterly anomaly. This second step of the response will be weaker in the damped runs, where the tropical winds are damped.
Figure 4-15: Cross-correlations between 10°S-10°N $\bar{u}$ (a), 21°N-40°N $\bar{u}$ (b), and 10 hPa and 60°N $\bar{u}$ (c) at a given pressure level with divergence of EP flux at 29°N-49°N and 1-66 hPa. Negative lag means that the winds lag the fluxes at that pressure level. The winds are not weighted by latitude. Notice the different ranges of the colorbars.

While Figure 4-12 shows how the state of the stratosphere changes based on equatorial wind oscillations, it does not say anything about the order in which the changes happen. This is especially true for the divergence of EP flux in midlatitudes and the polar vortex strength: one can expect an increase in convergence of EP flux to weaken the polar vortex, but a decrease in polar vortex strength is also likely to affect wave propagation in the region. Based on Figure 4-12, the changes in divergence of EP flux that are likely to be most strongly correlated to zonal wind strength throughout the stratosphere are those found between about 30°N and 50°N: they are located where they can influence both polar vortex strength and subtropical winds, and the signs of the anomalies are the same throughout most of the stratosphere. Apart from the upper stratospheric equatorial winds and polar vortex strength, zonal winds between about 20°N and 40°N are also of interest: zonal mean zonal wind anomalies in the lower equatorial stratosphere coincide with anomalies of the same sign at these latitudes. These anomalies are located close to the zero wind line, and could therefore have a strong influence on wave propagation. Figure 4-15 establishes how the zonal winds in these regions are affected by and affect the divergence of EP flux.

Figure 4-15a shows how increased EP flux convergence (anomalously negative EP flux divergence) leads to anomalously weak upper stratospheric equatorial winds. However, there is also a weak positive correlation at positive lag, indicating that upper
stratospheric equatorial winds could influence divergence of EP flux at midlatitudes. A possible mechanism for this is that anomalous upper stratospheric equatorial winds are carried poleward, where they alter the zonal wind structure in the midlatitudes, which changes the conditions for wave propagation. Figure 4-15b shows that the 21°N-40°N zonal mean zonal winds are negatively correlated with divergence of EP flux at short, positive lags. This is likely due to modulation of wave propagation by the winds. At short negative lags the correlation is positive, which makes sense: anomalously strong divergence of EP flux will decelerate the flow, and vice versa. The negative correlation at longer negative lags is more curious. More detailed analyses of this region show that the wind anomalies in this region have their origin close to the equator and are transported poleward by the overturning circulation (not shown). This is what the negative correlation at longer lags show. While Figure 4-15c indicates that polar vortex strength weakly influences the divergence of EP flux in the region of interest, the influence of EP flux divergence is larger on the polar vortex strength, with strong, positive and instantaneous correlations at short negative lags.

A mechanism for the influence of equatorial wind anomalies on midlatitude winds can be proposed based on the information presented in this chapter. That there is an influence of equatorial wind anomalies on midlatitude winds seems highly likely based on two results: first, that there is an oscillation in the ACF of polar vortex strength in the control run that disappears with increasing damping, and second; that there is a large change in polar vortex strength distribution with increasing damping, where the control run has substantially more days with a strong polar vortex compared to the damped runs. Since the only change between the model runs is the applied tropical damping in the stratosphere, the source of these differences should be searched for in that region. The mechanism must be able to explain the two features mentioned above for it to be feasible. The mechanism must show why the there is a positive peak in the ACF after a lag of just below 1000 days, and it must be able to explain why a strong polar vortex is more likely to develop in the control run compared to the damped runs. The first of the two features is a slow but transient response, while
the second feature is a steady-state response to first order. A mechanism that can explain these features is shown in Figures 4-16 and 4-17. The first figure shows a transition from a climatological mean state to a steady strong-vortex state, while the second figure shows a transition from a strong-vortex to a weak-vortex state. The choices of climatological-to-strong and strong-to-weak are arbitrary: these figures simply illustrate how the mechanism might work during specific situations.

After starting in a climatological mean state (Figure 4-16a) a sudden decrease in subtropical and midlatitude EP flux convergence throughout the depth of the stratosphere (see Figures 4-12b and e) will allow the polar vortex to grow strong and weaken the overturning circulation, and create a westerly anomaly in the upper stratospheric equatorial region (Figure 4-16b). The upper stratospheric equatorial response is immediate (see cross-correlations between equatorial winds and polar vortex strength in Figure 4-7a), and likely a result of changes in the residual circulation that will cause an acceleration in the region as described by Equation 4.4. When the zonal mean zonal wind at 1-8 hPa and 10°S-10°N is more than one standard deviation above the mean the strength of the overturning circulation decreases by ~ 10%. Shortly after the westerly anomalies in the upper stratospheric equatorial winds and polar vortex strength form, the easterly lower stratospheric equatorial wind anomaly will form (see Figure 4-7a; illustrated in Figure 4-16c). The lower stratospheric anomalies are likely influenced by changes in lower stratospheric subtropical EP flux (Figure 4-10). The lower stratospheric anomaly will take longer to form than the upper stratospheric anomaly, but a few hundred days after the initial change in EP flux convergence a strong-vortex state similar to the one seen in the top left panel of Figure 4-3 will have been reached (Figure 4-16d). During this state, the westerly upper stratospheric anomalies will be carried poleward by the overturning circulation, which helps maintain the unusually weak EP flux divergence in midlatitudes (Figure 4-15a). The residual circulation will accomplish this on a timescale of about 200 days. This is the reason for the higher counts of zonal mean zonal winds at 10 hPa and 60°N above 60 m/s in the control run seen in Figure 4-6. Since the equatorial wind anomalies are
Figure 4-16: Mechanism describing a transition from a steady climatological mean state to a steady strong-vortex state. Orange arrows show the overturning circulation, red (blue) colors denote positive (negative) anomalies.  

a) A climatological mean state. 

b) Immediately following a decrease in EP flux convergence. 

c) Shortly (∼100 days) after initial change. 

d) Steady strong-vortex state. Black text shows the time it takes for the residual circulation to flow from one end of the bracket to the other. See text for further details.
much smaller in the damped runs this effect will disappear with increasing damping. While equatorial wind anomalies will propagate upward on average, this state will be maintained as long as the subtropical and midlatitude wave forcing that created them is maintained. If advection of equatorial anomalies by the residual circulation was the only factor in the upward propagation of the anomalies, one would expect to see the lower stratospheric anomaly in the upper stratosphere after about 800 days.

Figure 4-17 provides an explanation for the $\sim 1000$ day oscillation in the ACF of polar vortex strength in the control run. After a long period in a strong-vortex
state (Figure 4-17a) the subtropical and midlatitude divergence of EP flux suddenly increases (Figure 4-17b). One reason for such an abrupt change can be anomalous tropospheric wave forcing. Immediately following the increase in divergence of EP flux, an easterly anomaly in the upper stratospheric equatorial region will form, while the polar vortex decreases in strength. A positive anomaly will also start to form in the lower equatorial stratosphere, but this will occur $\sim 100$ days later and will start in the lower regions. The winds immediately above are still anomalously easterly (in blue circle). The residual circulation will carry the easterly lower stratospheric anomaly upward. After $\sim 400 – 800$ days the easterly anomaly found in the lower stratosphere will have been carried upward by the overturning circulation (Figure 4-17c; blue circle). Note that this “anomaly” could be westerly compared to the climatological mean, but it will be less westerly than air that is advected following this since the zonal winds in this parcel started as anomalously easterly. After $\sim 600 – 1000$ the anomaly will have been advected to midlatitudes by the overturning circulation (Figure 4-17d; blue circle), where it will act to further decrease EP flux divergence in the region, thus acting as a positive feedback on the initial change $\sim 600 – 1000$ days earlier. This creates the peak in the ACF of the control run seen in Figure 4-5a. As above, this effect will disappear with increasing tropical wind damping and decreasing equatorial wind anomalies.

The modulation of the residual circulation by changes in wave convergence is somewhat similar to the “downward control” principle presented by Haynes et al. (1991), where the meridional residual circulation in the extratropics is determined by the vertically integrated wave forcing above, and results in changes in upwelling and downwelling equatorward and poleward of these regions. The changes in upwelling and downwelling will result in temperature anomalies that alter the meridional temperature gradient and hence the zonal mean zonal winds through thermal wind balance (if the anomalies are found outside the tropics, where the Coriolis parameter is significant). A similar mechanism was proposed by Kodera and Kuroda (2002) and Kodera and Shibata (2006) to explain an influence of the solar cycle on lower strato-
spheric equatorial temperature anomalies. According to these authors, an increase in upper stratospheric temperature gradient in the winter hemisphere will increase the strength of the zonal winds in the subtropics, which will deflect planetary waves from higher latitudes and decrease the EP flux convergence in the region. This will result in decreased downwelling in the polar region and decreased upwelling in the tropics (and hence a positive lower stratospheric equatorial temperature anomaly).

While the mechanisms shown in Figures 4-16 and 4-17 are different from those proposed by Kodera and Kuroda (2002) and Kodera and Shibata (2006) in terms of the order of events, there are some similarities. The most important similarity is that all mechanisms argue that changes in upper stratospheric subtropical zonal mean zonal wind will affect EP flux convergence in the region, and that the changes in EP flux convergence will result in changes in stratospheric residual circulation.

The upper stratospheric equatorial winds are of course not the only influence on mid-latitude EP flux convergence, but the mechanisms described above show how equatorial wind oscillations can affect the polar vortex, and they explain the differences in polar vortex strength in the control run compared to the damped runs.

### 4.3.3 Effects on SSW frequencies and intervals between SSWs

The SSW frequencies for the three runs are show in Table 4.1. SSWs were defined using the Charlton and Polvani (2007) method described in Chapters 2 and 3. The SSW frequency for the control run was slightly higher in this longer run than the H2 run used in Chapters 2 and 3 as well as the previous subsection (0.53 SSWs per 100 days compared to 0.48). The SSW frequency increases slightly with damping, but only the SSW frequency in NH40 is more than one standard deviation higher than that of the control run. The slightly increased SSW frequencies are likely one reason for the higher counts of weak winds at 10 hPa and 60°N in the damped runs compared to the control run, as seen in Figure 4-6. However, the differences in SSW frequencies between the model runs are too small to draw any definite conclusions.
Table 4.1: SSW frequencies and standard deviations for the three perpetual winter runs.

<table>
<thead>
<tr>
<th></th>
<th>NHC</th>
<th>NH100</th>
<th>NH40</th>
</tr>
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<tbody>
<tr>
<td>Total SSWs (SSWs per 100 days)</td>
<td>370 (0.53)</td>
<td>388 (0.55)</td>
<td>392 (0.56)</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>19</td>
<td>20</td>
<td>20</td>
</tr>
</tbody>
</table>

about whether or not tropical wind variations affect SSW frequencies in this model.

Figure 4-18 shows the binned number of intervals between SSWs in the three runs, along with the 1st and 99th percentiles of the bins. The figure mainly illustrates one thing: that SSW generation is not completely random. This is not unexpected since SSWs in a zonal mean sense is a reversal of the westerlies at 10 hPa and 60°N, and a weaker polar vortex will require less wave forcing to decelerate to this threshold. Figure 4-18 shows that SSW intervals below about 100 days are much more common in all three runs than one would expect from a random process. For reference, the 99th percentiles of the sums of SSW intervals below 100 days are 145, 156 and 158 for NHC, NH100 and NH40, respectively. The observed numbers are 172, 166 and 183, indicating that SSWs are more likely to occur when the polar vortex is weakened following a preceding SSW.

Figure 4-18 also shows that the number of long intervals between SSWs in all model runs may be longer than expected from a random process. The numbers of intervals above 700 days are 17, 14 and 10, for NHC, NH100 and NH40, respectively. This should be compared to the 95th and 99th percentiles of intervals above 700 days, which are 13 and 16 for NHC, and 12 and 14 for both NH100 and NH40. This is to be expected based on the fact that intervals below 100 days are also higher than the 99th percentiles: since SSW generation is not completely random and a preceding SSW within about 100 days increases the chance of producing another SSW, the long intervals should be more numerous compared to those of a random process with the same probability of producing a “SSW.” However, the number of intervals above this threshold exceeds the 99th percentile in NHC, reach the 99th percentile in NH100, but does not reach the 95th percentile in NH40. This is a slight indication that long
intervals may be more common in NHC compared to the damped runs, especially NH40. However, given the small number of intervals above these levels the results are sensitive to the value of the interval chosen. The number of intervals above 600 days, for example, does not exceed the 95th percentile for any of the model runs. As with the SSW frequencies, no strong conclusions can be drawn from these results alone.

There are, however, other indications that long intervals between SSWs may be connected to tropical wind variations in this model. Figure 4-19 shows the intervals between SSWs versus mean zonal mean wind speed at 10°S-10°N and 1-8 hPa during the intervals for the three model runs. The solid green line shows the mean wind strength and the dashed green lines show one standard deviation of the wind strength. There is a clear correlation between the zonal mean zonal wind at this level and the interval between SSWs, with the strength of the equatorial easterlies decreasing with increasing SSW intervals. For the control run (Figure 4-19a) some of the longest intervals between SSWs are associated with mean upper stratospheric equatorial wind anomalies exceeding one standard deviation. While the two damped runs (Figures 4-19a and b) also show correlations between the upper stratospheric equatorial wind strength and interval between SSWs, the differences in equatorial wind strength are smaller due to the equatorial wind damping. The correlations between these two variables are to be expected based on Figures 4-7 and 4-12, which showed that positive upper stratospheric equatorial wind anomalies are associated with a strong polar vortex. The mean polar vortex strength will be anomalously
strong during long periods without SSWs. Therefore, the correlations between upper stratospheric equatorial wind speed and polar vortex strength found in Figure 4-19 support the results of Gray et al. (2001a) and Gray et al. (2001b), who found that upper stratospheric wind anomalies were correlated with and unusually strong Northern Hemisphere polar vortex. The results also agree with Gray (2003), who found that model runs with westerly upper stratospheric winds produced fewer SSWs compared to model runs with easterly anomalies.

Figure 4-19 only shows that there is a correlation between the two variables and says nothing about the causality, but the mechanism through which upper stratospheric equatorial wind anomalies reinforce positive EP flux anomalies in midlatitudes described in Section 4.3.2 could be why long intervals between SSWs seem slightly more common in the control run compared to the damped runs. Since the stratospheric equatorial wind anomalies are damped in NH100 and NH40 the influence of upper stratospheric equatorial wind anomalies on the midlatitude will be weaker than in the control run.


4.4 Results with a seasonal cycle

The zonal mean zonal winds for the seasonal cycle control run (SC) can be seen in Figure 4-20. The zonal mean zonal winds are practically identical for the damped seasonal run (S100; not shown). The figure shows that the winds during Northern Hemisphere winter (panel a) are similar to those of the H2 run, although the polar vortex is about 15 m/s weaker during December through February. This discrepancy is not unexpected since H2 is run under perpetual midwinter conditions. In contrast, the polar vortex during Southern Hemisphere winter (panel c) is very strong and approaches 90 m/s at its strongest level. Another difference in zonal mean zonal wind structures between the Northern Hemisphere winter runs and the seasonal runs is the equatorial wind at the highest pressure levels, which is less easterly in the seasonal runs. Overall, Figure 4-20 shows that the model does a reasonably good job of reproducing the seasonal zonal mean zonal wind structure of the observed atmosphere, with a disturbed polar vortex during Northern Hemisphere winter and a very strong polar vortex during Southern Hemisphere winter.

The equatorial wind oscillations also exist in SC. Figure 4-21 shows the upward propagation of deseasonalized equatorial wind anomalies during 5000 days in SC. The format of the figure is identical to that of Figure 4-3. The magnitudes and timescales of the anomalies in the left panels are similar to those of NHC, and the strong anti-correlation between upper and lower stratospheric equatorial winds can be clearly seen. The right panels show that the effect of the damping on the equatorial winds is similar to that of the Northern Hemisphere winter run, with a significant decrease in equatorial wind variability in S100 compared to SC. The effect of SSWs on upper stratospheric equatorial winds can be seen in the sharp drops in wind strength for S100 in the top right panel.

Despite the similarities between Figures 4-3 and 4-21, the cross-correlations of mid-stratospheric equatorial winds with equatorial winds are quite different in SC compared to NHC. Figure 4-22 shows that while there is an upward propagation of anoma-
Figure 4-20: Zonal mean zonal winds during DJF (a), MAM (b), JJA(c) and SON (d) in SC.
Figure 4-21: Same as Figure 4-3, but for the seasonal cycle run. The wind anomalies have been deseasonalized. Black lines in the right panels show SC, while red lines show S100.
Figure 4-22: Same as Figure 4-4, but for the seasonal cycle run. The wind anomalies have been deseasonalized. Panel a shows SC, and panel b shows S100. Notice the different lengths of the lag axes.

lies from the lower stratosphere to about 10 hPa in the first 500 days of SC, a weak anti-correlation between upper and lower stratospheric equatorial winds remains for lags of over 6000 days. This is different to NHC, where anomalies at lower pressure levels led to anomalies of the same sign in the upper stratosphere 1000-2000 days later (Figure 4-4a). Based on Figure 4-21 the equatorial anomalies clearly propagate to the uppermost stratosphere in SC, but the upward propagation seems to be masked by anti-correlations of very long timescales. The damped seasonal run (Figure 4-22b) shows that the upward propagation seen in the first 500 days of SC remains with damping, but that correlations at longer lags disappear. The cross-correlation between upper and lower stratospheric equatorial winds is also weaker than in the control run.

Figure 4-23 shows the full time series of upper and middle stratospheric equatorial winds for SC, with SSWs marked on the time axis (SSWs are discussed in more detail below). The figure shows two things: first, that even though short oscillations (less than 2000 days) in the wind occur the long-lived anomalies that the cross-correlations in Figure 4-22a alluded to do exist. Second, that the periods of unusually large equatorial wind oscillations seem to occur during periods with no SSWs (see, for
example, the periods around 15,000 and 65,000 days). Long periods without SSWs do not lead to equatorial wind anomalies of a given sign, but only seem to increase the magnitude of the variability of the winds. As seen in Figure 4-14, SSWs exert a strong influence on upper stratospheric equatorial winds. There are, however, much fewer SSWs in SC compared to NHC, since SSWs only occur during Northern Hemisphere winter in this model. It could be that SSWs keep the equatorial winds from growing too anomalous by damping the anomalies, but another explanation is that strongly anomalous equatorial winds decrease SSW generation. However, this seems less likely since the long periods without SSW contain both anomalously westerly and easterly winds.

Figure 4-24 shows the cross-correlations between zonal mean zonal winds at 10 hPa and 60° and equatorial winds, after the winds have been deseasonalized. An instantaneous cross-correlation between stratospheric equatorial winds and winds at 10 hPa
and 60°N can be seen in panels a and b. This is likely connected to SSWs and possibly to stratospheric final warmings (final warmings discussed in more detail below), in the way that Figure 4-14 showed. The correlation is weaker in SC (Figure 4-24a) compared to S100 (Figure 4-24b), but the magnitudes of the oscillations are stronger in the control run compared to the damped run, and the large long-term oscillations seen in Figure 4-23a are likely to result in a smaller values for the cross-correlations. Both SC and S100 show a weak negative correlation at negative lag between about 10 and 2 hPa, and since it cannot be found in NHC or NH100 it is likely connected to the final warmings. The structure of the correlations between Northern Hemisphere polar vortex strength and equatorial winds are similar between the two runs, and most correlations disappear within one year. Since the mechanisms through which equatorial winds and polar vortex strength influence each other occur during Northern Hemisphere winter, the magnitudes of the correlations are much lower than those found in panels a and b in Figure 4-7. Overall, Figures 4-24a and b do not show any correlations that indicate that the equatorial winds influence or are influenced by the Northern Hemisphere polar vortex on timescales longer than about 200 days.

The correlations between equatorial winds and Southern Hemisphere polar vortex strength differ quite a lot between the control run and the damped run. The control run (Figure 4-24c) shows very weak positive (negative) correlations between lower (upper) stratospheric equatorial winds and Southern Hemisphere polar vortex strength. These anti-correlations likely originate from the long-lived anomalies seen in Figure 4-22a, rather than an actual connection to the Southern Hemisphere polar vortex strength.

The damped run does not have the long-term oscillations found in the control run, and shows a positive, instantaneous correlation between upper stratospheric winds and Southern Hemisphere polar vortex strength. This is likely a result of the final warmings. A final warming is defined as the day when the polar vortex breaks up for the season, and it is dynamically rather similar to a SSW. It is therefore reasonable to assume that a final warming should have an influence on the upper stratospheric
Figure 4-24: Cross-correlations between 10°S-10°N zonal mean zonal wind at a given pressure level and the zonal mean zonal winds at 10 hPa and 60°N (top) and 60°S (bottom) for SC (left) and S100 (right). Notice the longer lag axis in panel c. Positive lag means that 10 hPa and 60° winds lag the equatorial winds at that pressure level. The equatorial winds are deseasonalized and not weighted by latitude. Notice the different ranges of the colorbars.
equatorial winds similar to that of SSWs. The effect of final warmings will largely be removed during deseasonalization, but some influence will remain due to the variability in the timing of the final warming (see below). Since the correlation between upper stratospheric equatorial winds and Southern Hemisphere polar vortex strength is weaker than that found between the same winds and the Northern Hemisphere polar vortex strength, it seems SSWs exert a larger influence on equatorial wind variability compared to final warmings. There is a weak, negative correlation between the lower stratospheric winds and the polar vortex, which extends for a few hundred days into positive lag. The negative correlation maximizes around zero lag, but is weaker than its counterparts in Figure 4-7. There is an indication of a positive correlation between lower stratospheric equatorial wind strength and polar vortex strength at negative lag, but it is very weak. Overall, any correlations that exist between equatorial winds and Southern Hemisphere polar vortex strength outside of about 100 days of zero lag are too weak to suggest that equatorial winds influence the polar vortex strength on longer timescales.

Table 4.2 shows SSW frequencies and final warming (FW) dates for SC and S100. The FW dates were defined as the last day of the winter that the zonal mean zonal wind strength at 10 hPa and 60° became easterly without returning to a strength exceeding 5 m/s before the next winter. The differences in SSW frequencies between the two runs are negligible. The frequencies of 0.47 and 0.48 SSWs per year for SC and S100, respectively, are lower than the 0.63 per year found in ERA-Interim reanalysis data (Dee et al. (2011); see Chapter 2). The final warming dates for SC and S100 are also very similar, both in terms of dates and standard deviations. For comparison, the same definition applied to the ERA-Interim dataset used in Chapter 2 gives a Northern Hemisphere (Southern Hemisphere) FW date of April 9 (November 20), with a standard deviation of 21 days (12 days). The only difference in SSWs and FW dates between SC and S100 worth noting is that SC had seven years with two or more SSWs, while S100 had 13. It is difficult to determine whether or not this difference is statistically significant, but if it is then it indicates that damping of
Table 4.2: SSW frequencies and final warming (FW) dates for SC and S100, with standard deviations for the final warmings.

<table>
<thead>
<tr>
<th></th>
<th>SC</th>
<th>S100</th>
</tr>
</thead>
<tbody>
<tr>
<td>NH SSWs (SSWs per year)</td>
<td>103 (0.47)</td>
<td>106 (0.48)</td>
</tr>
<tr>
<td>Mean day of NH FW (standard deviation)</td>
<td>90 (17)</td>
<td>92 (17)</td>
</tr>
<tr>
<td>Mean day of SH FW (standard deviation)</td>
<td>340 (14)</td>
<td>339 (13)</td>
</tr>
</tbody>
</table>

equatorial winds increases the probability of producing more than one SSW in a given year. The mechanism by which that would happen is unclear; the mean polar vortex strength recovers at practically the same rate for both runs, and there is no difference in the mean time of the first SSW (not shown). Furthermore, the mechanism for how upper stratospheric equatorial winds can influence polar vortex strength presented in Section 4.3.2 suggests that anomalously easterly upper stratospheric equatorial winds, which one would expect after a SSW, can decrease the polar vortex strength. This would suggest that multiple SSWs in a year should be more likely in the control run. However, the mechanism in Section 4.3.2 describes the influence on the mean circulation, and not on SSW generation. In Section 4.3.3 it was shown that the number of SSW intervals below 100 days for the three Northern Hemisphere winter runs were 172 (NHC), 166 (NH100) and 183 (NH40), which indicates that the probability of producing a SSW shortly following a previous SSW does not change in a consistent manner with increasing equatorial wind damping. Based on this, it seems likely that the increased number of years with multiple SSWs found in S100 is not indicative of an increase in probability of producing more than one SSW in a given year.

The results of this section show that the equatorial wind oscillations in this model are not sufficient to significantly alter the dynamics of the polar vortex when the model is run with a seasonal cycle. The structures and magnitudes of the oscillations for SC and NHC are very similar. While there were some indications that equatorial wind oscillations affected the polar vortex under Northern Hemisphere winter conditions, it seems the effects are small enough to disappear when a seasonal cycle is added. Alternatively, the effects of equatorial wind oscillations on polar vortex strength seen under Northern Hemisphere winter conditions take place on timescales that are longer
than a season, which would diminish or remove their effect when the model is run with a seasonal cycle. The upward propagation of anomalies takes 1000-2000 days (Figure 4-4a), and the timescales on which equatorial wind oscillations affect midlatitude EP flux divergence are less than 300 days (Figure 4-15a). Another explanation is that the long-lived oscillations in SC mask the effect on SSWs that the shorter (order 1000 day) oscillations have on the polar vortex.

4.5 Discussion

This chapter investigated the effects of tropical wind oscillations on polar vortex strength. By suppressing tropical wind variability I showed that the Northern Hemisphere polar vortex is slightly stronger and experiences slightly fewer SSWs when large tropical wind oscillations are present in the perpetual winter run, but that no differences in zonal wind structure, SSW frequencies or timing of final warming dates could be found when a seasonal cycle was added to the model. I also showed that the oscillations in polar vortex strength found in NHC were likely connected to upper stratospheric equatorial wind variations, and that the oscillations disappeared when the equatorial winds were damped.

I identified a mechanism through which variability of EP flux divergence in the subtropics and midlatitudes created the tropical wind anomalies. A possible mechanism for equatorial upper stratospheric influence on polar vortex strength was also established, where the overturning circulation carries westerly equatorial anomalies poleward, thereby decreasing the wave convergence in the region and strengthening the anomalously strong polar vortex. This was identified as a likely source for the lower SSW frequency and the higher number of days with a strong polar vortex in NHC compared to NH100 and NH40. Interestingly, no direct relationship where lower stratospheric equatorial winds influence polar vortex strength was found. This contrasts the observed relationship between subtropical and midlatitude wind strength and the QBO, where tropical wind anomalies at lower pressure levels have been found
to be correlated with anomalies at higher latitudes (Holton and Tan, 1980) and important for reproducing observed connections between the tropics and polar vortex (Gray et al., 2001a).

The positive correlation between upper stratospheric equatorial winds and polar vortex strength in the perpetual winter runs agree with the findings of Gray et al. (2001a), Gray et al. (2001b) and Gray (2003). The slightly lower SSW frequency found in NHC agreed with the findings of Gray (2003), since NHC produced stronger westerly upper stratospheric wind anomalies compared to NH100 and NH40. The equatorial winds did not seem to affect the polar vortex when a seasonal cycle was added, even though positive, instantaneous correlations between upper stratospheric equatorial winds and Northern Hemisphere polar vortex strength were found (Figures 4-24a and b). It is possible that the influence on Northern Hemisphere polar vortex strength by upper stratospheric equatorial winds found in NHC acts on timescales longer than a typical season, and that the influence is diminished when a seasonal cycle is added. Another explanation is that SC exhibited tropical wind variations on timescales much longer than those found in NHC (Figure 4-4 versus 4-22 and 4-23), and that these long term oscillations masked the effect of the upward propagating signals clearly seen in Figure 4-21. This seems to be what is happening for the correlations between equatorial wind and Southern Hemisphere polar vortex strength in SC (Figure 4-24c). The fact that the tropical wind oscillations in this model were spontaneous and not prescribed as in Gray et al. (2001a) and Gray (2003) likely made the effects of tropical wind oscillations on midlatitude winds more difficult to distinguish.

Despite the negative results of the seasonal runs, the possible connection between upper stratospheric equatorial winds and polar vortex strength is intriguing. It supports Gray et al. (2001a), Gray et al. (2001b) and Gray (2003) in highlighting the fact that the upper stratosphere plays a role in modulating the polar vortex strength in the winter hemisphere. Since the upper stratosphere is strongly influenced by solar radiation, this also supports the conclusion that the solar cycle could influence the midlatitude winter stratosphere. The results of this chapter confirm the importance
of resolving upper stratospheric wind variations in order to capture stratospheric midlatitude variability.
Chapter 5

Conclusions and outlook

In this thesis I have investigated the effects eddy-eddy interactions (EEI) and stratospheric tropical wind variability on sudden stratospheric warming (SSW) formation. Chapter 2 introduced a tropospheric heating perturbation that produces splits and displacements in comparable amounts in an atmospheric GCM, both when the forcing is of wavenumber 1 and of wavenumber 2. These model runs were referred to as H1 and H2, respectively. H1 and H2 were compared to a commonly used topographic wavenumber 2 forcing, which produces almost only splits. The results of the three model runs were used to investigate the extent to which splits and displacements are forced by anomalous tropospheric wave activity. It was found that the fractions of SSWs forced by strongly anomalous tropospheric wave activity in this simple model were similar to the fraction found in reanalysis data (Birner and Albers, 2017). This indicates that the model may be producing SSWs for reasons similar to those of the observed atmosphere, despite its simplicity. Furthermore, I found that the fractions of splits and displacements forced by strongly anomalous tropospheric wave flux were dependent on the wavenumber of the forcing, with displacements being more likely than splits to be forced by anomalous tropospheric wave flux at a 95% significance level in H1. I also discovered that a quarter to a third of SSWs produced in the model were associated with tropospheric flux anomalies never exceeding one standard devi-
ation, which strongly suggests that some SSWs may be produced by internal stratospheric processes that increase the propagation of climatological wave flux to higher levels, rather than anomalous forcing from the troposphere. Finally, zonal mean polar geopotential height anomalies associated with splits were found to be different from those of displacements in H1, with structures similar to those found in reanalysis data (Mitchell et al., 2013). However, the effects of splits and displacements, as measured by the difference between polar geopotential height anomalies preceding and following SSW, were very similar. I concluded that the state of the polar region may influence the type of SSW produced in this model setup.

An important conclusion from Chapter 2 is that SSW generation is far from a simple interaction between anomalous tropospheric wave forcing and stratospheric mean flow (Matsuno, 1971). That a large amount of SSWs occur without significant anomalous tropospheric wave flux anomalies indicates that the state of the winter stratosphere plays a crucial role in SSW generation. The tropospheric forcings used were of one wavenumber only, and yet both displacements (wavenumber 1 events) and splits (wavenumber 2 events) were produced. This indicates that significant transfers of energy between the two main stratospheric wavenumbers occur from the wave source in the troposphere to the region of wave breaking in the stratosphere. These observations motivated the focus of Chapter 3, where EEI were turned off in parts of the atmosphere during model runs using the wavenumber 1 and wavenumber 2 tropospheric heating. I showed that SSW frequencies could be highly sensitive to EEI throughout the atmosphere: even in model runs where EEI were only turned off above 50 hPa the SSW frequency was significantly altered compared to the control run. Since the majority of climatological wave forcing originates in the troposphere the model runs without EEI in the upper stratosphere show that stratospheric climate is highly dependent on internal stratospheric processes, and the stratosphere is not simply a passive recipient of wave forcing from below. These results contrast previously published work that has highlighted the importance of tropospheric and lower stratospheric wave flux for SSW generation (Polvani and Waugh, 2004; Birner and
Albers, 2017). Exactly how EEI affect SSW generation is not clear, since the SSW frequencies of the different runs varied greatly and in non-predictable ways. However, removal of EEI in the upper stratosphere always resulted in more equatorward vertical EP flux and EP flux divergence, something that was not related to latitudinal changes in the polar vortex. This could be one of the reasons for the changes in SSW frequencies associated with removing EEI.

Chapter 3 also showed that EEI are required locally for splits and displacements to occur. Without EEI in the upper stratosphere, SSWs result in meridional cascades of waves unlike anything seen in the observed atmosphere. This is despite the fact that large amounts of wavenumber 1 or wavenumber 2 climatological forcing existed in the upper stratosphere even in the absence of EEI. However, splits and displacements at 10 hPa did occur when EEI were allowed in the upper stratosphere only, showing that tropospheric and lower stratospheric wave-wave interactions are not necessary for the formation of splits and displacements. This shows that displacements and splits can be considered local reactions to wavenumber 1 and 2 forcings, respectively.

Chapters 2 and 3 both highlight the complexity of SSW generation. Even though SSWs can be considered wave-mean flow interactions to first order, it is clear that much of the dynamics behind SSW generation is lost when higher order processes are ignored. To improve our understanding of SSWs future work should focus on the importance of stratospheric transmission of climatological wave forcing from the troposphere, and the role of EEI in stratospheric dynamics.

In Chapter 4 I investigated the effects of the model’s spontaneous tropical wind oscillations on the polar vortex, with a particular emphasis on SSWs. By damping the tropical wind oscillations in the model, I showed that SSWs are slightly less common in the control run compared to model runs with damped tropical winds. Long intervals between SSWs (leading to an anomalously strong polar vortex) were found to be correlated with anomalously westerly upper stratospheric equatorial wind anomalies, which agreed with both observations (Gray et al., 2001b) and previous model results.
A mechanism explaining the formation and propagation of the equatorial anomalies was presented, in which the anomalies are caused by changes in subtropical and midlatitude EP flux convergence. To explain the higher counts of days with a strong polar vortex found in the control run, I suggested that the overturning circulation carried westerly upper stratospheric equatorial wind anomalies poleward, where they strengthened the polar vortex and decreased convergence of EP flux, thereby reinforcing the conditions that lead to an anomalously strong polar vortex. However, when adding a seasonal cycle to the model no significant differences in SSW frequencies or final warming dates could be found between the control run and the damped run. This could be because the upper stratosphere affects the polar vortex on timescales longer than a typical winter, whereby the effect of the upper stratosphere on the polar vortex would be diminished in the presence of a seasonal cycle, or because the control run produced extremely long tropical wind oscillations that masked the correlation between the upper stratospheric equatorial region and the polar vortex caused by oscillations of shorter periods.

Even though the results with the seasonal cycle were negative, the connection between the upper stratospheric equatorial winds and polar vortex strength found in model runs under perpetual Northern Hemisphere winter conditions suggest that resolving upper stratospheric stratospheric wind variations could be important for capturing stratospheric midlatitude variability. The drastic changes in SSWs frequencies seen when EEI were turned off linearly from 50 hPa to 30 hPa in Chapter 3 indicate that internal stratospheric processes are crucial above levels that are usually considered most important (Polvani and Waugh, 2004; Birner and Albers, 2017), and reinforce this conclusion. The results of this thesis therefore indicate that future research relating to SSWs should be open to possible connections between upper stratosphere wind variability (in the tropics as well as further poleward) and the stratospheric polar vortex strength.
Appendix A

Supplementary figures for Chapter 2
Figure A-1: Vertical wave-1 (left) and wave-2 (right) EP flux anomalies around displacements (top) and splits (bottom) in the H1 run. The EP flux anomalies were averaged between 40°N and 80°N. Areas within green lines are statistically significant at a 95% confidence level. Notice the different ranges of the colorbars.
Figure A-2: Same as Figure A-1 but for H2.

Figure A-3: Vertical EP flux wave-1 (left) and wave-2 (right) anomalies around splits in the T2 run. Areas within green lines are statistically significant at a 95% confidence level. Notice the different ranges of the colorbars.
References


