Northern Hemispheric Cold Spells and their Tropospheric-Stratospheric Link

Kathrin Finke
Northern Hemispheric Cold Spells and their Tropospheric-Stratospheric Link

Kathrin Finke

Abstract

Cold spells have severe consequences for society. They require early warnings for elaborate mitigation strategies on sub-seasonal to seasonal time-scales. Intense stratospheric westerlies and a polar vortex breakdown (SSW) may enhance extended-range forecast skill for Eurasian and North American cold extremes through a dynamic coupling to the troposphere. Understanding the complex interplay remains a challenging task that requires further investigation.

Since fine-grained observational stratospheric data is limited to the satellite era, climate model simulations, such as atmosphere-only simulations (AMIP) from the Coupled Model Intercomparison Project Phase 6, can be considered. Application of the common empirical orthogonal function method in Paper II, a tool for multimodel comparison and evaluation, unveiled differences in daily winter 2m temperatures (T2m) across four reanalyses while stratospheric geopotential height varies across AMIP models. Results show a link between a weak polar vortex and cold T2m anomalies over Eurasia in reanalysis data.

In addition, quantile regression is a simple but proficient statistical method that neatly enables modeling the response variable's complete conditional distribution. Thereby, information about extremes, which hide in the distribution’s tails, is extracted. Application to boreal winter ERA5 reanalysis data and teleconnection indices in Paper I reveals significant asymmetries in duration, strength, and direction of the stratosphere-troposphere connection across quantiles.

Regionally specific, lagged composite analysis of ERA5 data in Paper III verifies the canonical warm stratosphere - cold Eurasia relation. However, persistent Eurasian cold spells may also coincide with a strong polar vortex. We find stratospheric reflection of upward propagating planetary waves toward the North Atlantic to potentially influence mid-tropospheric circulation anomalies that travel towards Eurasia. By interacting with a quasi-stationary anticyclone over the Barents Sea, which promotes a cold Eurasia, these circulation anomalies likely influence the persistence and strength of the cold spell.

Paper IV discusses the relationship between the 2018/2019 winter SSW and the subsequent North American cold spell using the JRA-55 reanalysis. An unusual wave number 3 planetary wave pulse in the stratosphere led to a polar vortex split. Further, wave reflection at the stratospheric Aleutian high likely fostered the circulation configuration, i.e., positive North Pacific and negative North American geopotential height anomalies that facilitated the cold temperatures.

Stockholm 2024

http://urn.kb.se/resolve?urn=urn:nbn:se:su:diva-224415


Department of Meteorology

Stockholm University, 106 91 Stockholm
NORTHERN HEMISPHERIC COLD SPELLS AND THEIR TROPOSPHERIC-STRATOSPHERIC LINK

Kathrin Finke
Northern Hemispheric Cold Spells and their Tropospheric-Stratospheric Link

Kathrin Finke
"Det ordnar sig."
Abstract

Cold spells have severe consequences for society. They require early warnings for elaborate mitigation strategies on sub-seasonal to seasonal time-scales. Intense stratospheric westerlies and a polar vortex breakdown (SSW) may enhance extended-range forecast skill for Eurasian and North American cold extremes through a dynamic coupling to the troposphere. Understanding the complex interplay remains a challenging task that requires further investigation. Since fine-grained observational stratospheric data is limited to the satellite era, climate model simulations, such as atmosphere-only simulations (AMIP) from the Coupled Model Intercomparison Project Phase 6, can be considered. Application of the common empirical orthogonal function method in Paper II, a tool for multimodel comparison and evaluation, unveiled differences in daily winter 2m temperatures (T2m) across four reanalyses while stratospheric geopotential height varies across AMIP models. Results show a link between a weak polar vortex and cold T2m anomalies over Eurasia in reanalysis data. In addition, quantile regression is a simple but proficient statistical method that neatly enables modeling the response variable’s complete conditional distribution. Thereby, information about extremes, which hide in the distribution’s tails, is extracted. Application to boreal winter ERA5 reanalysis data and teleconnection indices in Paper I reveals significant asymmetries in duration, strength, and direction of the stratosphere-troposphere connection across quantiles. Regionally specific, lagged composite analysis of ERA5 data in Paper III verifies the canonical warm stratosphere - cold Eurasia relation. However, persistent Eurasian cold spells may also coincide with a strong polar vortex. We find stratospheric reflection of upward propagating planetary waves toward the North Atlantic to potentially influence mid-tropospheric circulation anomalies that travel towards Eurasia. By interacting with a quasi-stationary anticyclone over the Barents Sea, which promotes a cold Eurasia, these circulation anomalies likely influence the persistence and strength of the cold spell. Paper IV discusses the relationship between the 2018/2019 winter SSW and the subsequent North American cold spell using the JRA-55 reanalysis. An unusual wave number 3 planetary wave pulse in the stratosphere led to a polar vortex split. Further, wave reflection at the stratospheric Aleutian high likely fostered the circulation configuration, i.e., positive North Pacific and negative North American geopotential height anomalies that facilitated the cold temperatures.
Kälteeinbrüche haben ernste Konsequenzen und erfordern zeitnahe Warnungen für ausgefeilte Mitigationsstrategien auf sub-saisonaler bis saisonaler Ebene. Starke stratosphärische Westwinde oder ein Polarwirbelzusammenbruch (SSW) verbessern die Vorhersagefähigkeit für eurasische und nordamerikanische Kälteextreme aufgrund dynamischer Kopplung zur Troposphäre. Das Verstehen dieses komplexen Zusammenspiels bleibt eine anspruchsvolle Aufgabe, die weitere Untersuchungen erfordert.


Verzögerte Kompositanalysen der ERA5-Daten in Paper III bestätigen die kanonische warme Stratosphäre - kaltes Eurasien Beziehung. Eurasische Kälteperioden können jedoch auch mit einem starken Polarwirbel zusammenfallen. In diesem Fall reflektiert die Stratosphäre die sich aufwärts ausbreitenden Rossby Wellen zurück Richtung Nordatlantik. Diese beeinflussen potenziell mitteltroposphärische Zirkulationsanomalien, die sich in Richtung Eurasien bewegen.

Durch die Wechselwirkung mit einem quasi-stationären Antizyklon über der Barentssee, der das kalte Eurasien begünstigt, beeinflussen diese Zirkulationsanomalien wahrscheinlich Kälteeinbruchdauer und -stärke.

belspaltung. Zusätzlich förderte die Wellenreflektion am stratosphärischen Aleutenhoch vermutlich die Zirkulationskonfiguration, d.h. jeweils positive und negative Anomalien der geopotentiellen Höhe über dem Nordpazifik und Nordamerika, die die Kälte begünstigten.
List of Papers

The following papers, referred to in the text by their Roman numerals, are included in this thesis.


Reprints were made with permission from the publishers.

[1] shared first author
The PhD project was initially conceptualized by my supervisor A. Hannachi. The scope and objectives of this research were refined by me throughout my time at MISU in collaboration with A. Hannachi and my co-authors.

A. Hannachi initially proposed the project for Paper I. The concept evolved whilst working on the project through numerous discussions between A. Hannachi and me, becoming more refined and specific. I did all data procurement, preprocessing, and analysis, as well as producing the plots and writing the manuscript. A. Hannachi reviewed and edited the manuscript. I conducted the revision based on the reviewers’ comments with input from A. Hannachi.

Paper II was primarily conceptualized and executed by A. Hannachi. I was responsible for data procurement, pre-processing, and composing of Chapter 3, which focuses on the data. Additionally, I produced Table 1. A. Hannachi conducted the main analysis, created the plots, and wrote the manuscript. I assisted in the interpretation of the plots, edited segments of the manuscript, and contributed to the revision process based on feedback from the reviewers.

The initial concept for Paper III originated from my ideas. The project underwent further refinement through discussions in the course of the research process involving A. Hannachi and T. Hirooka, who visited MISU as an IMI guest. The 3D wave activity flux code was provided by W. Iqbal. I undertook the tasks related to data procurement, data pre-processing, as well as their analysis, in addition to creating the plots and writing the manuscript. All co-authors reviewed and edited the manuscript. I revised the manuscript based on feedback received from the reviewers, incorporating input from all co-authors.

The concept and initial draft of the manuscript for Paper IV originated from A. Hannachi. Based on discussions with T. Hirooka and Y. Matsuyama during my research visit at Kyushu University in Fukuoka, Japan, I revised and extended the analysis using a different data set than originally utilized. I further generated most plots and rewrote and extended the initial draft. W. Iqbal provided the 3D wave activity flux code and Figure 6. Y. Matsuyama provided the Eliassen-Palm flux data. All co-authors reviewed and edited the manuscript.
# Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>vi</td>
</tr>
<tr>
<td>Sammanfattning</td>
<td>vii</td>
</tr>
<tr>
<td>Zusammenfassung</td>
<td>ix</td>
</tr>
<tr>
<td>List of Papers</td>
<td>xi</td>
</tr>
<tr>
<td>Author’s contribution</td>
<td>xiii</td>
</tr>
<tr>
<td>Abbreviations</td>
<td>xvii</td>
</tr>
<tr>
<td>List of Figures</td>
<td>xix</td>
</tr>
<tr>
<td>1 Introduction</td>
<td>1</td>
</tr>
<tr>
<td>2 The Stratosphere</td>
<td>5</td>
</tr>
<tr>
<td>2.1 Essential Characteristics of the Stratosphere</td>
<td>5</td>
</tr>
<tr>
<td>2.2 The Stratospheric Polar Vortex and its Evolution</td>
<td>6</td>
</tr>
<tr>
<td>2.3 Stratospheric Polar Vortex Extremes</td>
<td>8</td>
</tr>
<tr>
<td>3 Stratosphere-Troposphere Coupling</td>
<td>13</td>
</tr>
<tr>
<td>4 Tropospheric Response to Stratospheric Variability</td>
<td>19</td>
</tr>
<tr>
<td>5 Research Summary</td>
<td>23</td>
</tr>
<tr>
<td>6 Outlook</td>
<td>29</td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>xxxi</td>
</tr>
<tr>
<td>References</td>
<td>xxxiii</td>
</tr>
</tbody>
</table>
## Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AMIP</td>
<td>Atmospheric Model Intercomparison Project Phase 6</td>
</tr>
<tr>
<td>AO</td>
<td>Arctic Oscillation</td>
</tr>
<tr>
<td>CMIP</td>
<td>Coupled Model Intercomparison Project Phase 6</td>
</tr>
<tr>
<td>DJF</td>
<td>December, January and February</td>
</tr>
<tr>
<td>ECMWF</td>
<td>European Centre for Medium-Range Weather Forecasts</td>
</tr>
<tr>
<td>EOF</td>
<td>empirical orthogonal function</td>
</tr>
<tr>
<td>EP</td>
<td>Eliassen-Palm</td>
</tr>
<tr>
<td>EUR</td>
<td>Eurasia</td>
</tr>
<tr>
<td>JMA</td>
<td>Japan Meteorological Agency</td>
</tr>
<tr>
<td>JRA-55</td>
<td>Japanese 55-year Reanalysis</td>
</tr>
<tr>
<td>LR</td>
<td>standard linear regression</td>
</tr>
<tr>
<td>MJO</td>
<td>Madden-Julian Oscillation</td>
</tr>
<tr>
<td>NAM</td>
<td>Northern Annular Mode</td>
</tr>
<tr>
<td>NAO</td>
<td>North Atlantic Oscillation</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanographic and Atmospheric Administration</td>
</tr>
<tr>
<td>PCE</td>
<td>persistent cold event</td>
</tr>
<tr>
<td>PV</td>
<td>potential vorticity</td>
</tr>
<tr>
<td>QBO</td>
<td>Quasi-Biennial Oscillation</td>
</tr>
<tr>
<td>QR</td>
<td>quantile regression</td>
</tr>
<tr>
<td>RI</td>
<td>reflective index</td>
</tr>
<tr>
<td>S2S</td>
<td>sub-seasonal to seasonal</td>
</tr>
<tr>
<td>SLP</td>
<td>sea level pressure</td>
</tr>
<tr>
<td>SPV</td>
<td>strong polar vortex</td>
</tr>
<tr>
<td>Abbreviation</td>
<td>Description</td>
</tr>
<tr>
<td>--------------</td>
<td>------------------------------</td>
</tr>
<tr>
<td>SSW</td>
<td>sudden stratospheric warming</td>
</tr>
<tr>
<td>T2m</td>
<td>2m temperature</td>
</tr>
<tr>
<td>WAF</td>
<td>wave activity flux</td>
</tr>
<tr>
<td>WN</td>
<td>wave number</td>
</tr>
<tr>
<td>Z10</td>
<td>10hPa geopotential height</td>
</tr>
<tr>
<td>Z500</td>
<td>500hPa geopotential height</td>
</tr>
</tbody>
</table>
List of Figures

1.1 The S2S prediction gap. (Figure taken from Mariotti et al. (2018)) 1

2.1 Layers of the atmosphere and characteristic temperature profile based on the U.S. Standard Atmosphere. The temperature [K] is on the x-axis and the altitude [km; hPa] is on the y-axis. Figure taken from Wallace and Hobbs (2006). Reprinted with permission from Elsevier.) 5

2.2 Latitude-height profile of the climatological zonal mean zonal wind (shading) and temperature (contours) based on DJF 9-day running mean JRA-55 data (1991-2020). 7

2.3 Climatological 10hPa geopotential height in gpdam (contours) and temperature in °C (shading) based on the ERA5 reanalysis from 1979-2022 for January-March. 10

2.4 10hPa geopotential height in gpdam (contours) and temperature anomaly (shading) based on ERA5 reanalysis data from 1979-2022 for March 15th 2019 (top), January 1st 2019 (bottom left), January 15th 2019 (bottom right). 11

3.1 Schematic representation of the positive (left) and negative (right) phase of the AO. (Figure provided by NOAA.) 13

3.2 Composite of the time-height development of the NAM index for A) weak vortex events and B) strong vortex events. (Figure taken from Baldwin and Dunkerton (2001). Reprinted with permission from AAAS.) 14

3.3 Geopotential height anomalies with respect to the zonal mean (contours), the WAF in the x-z-plane (arrows), both meridionally averaged (48°N-74°N), and the zonal wind at 60°N (shading) for January 15th, 2019 based on JRA-55 data. (Figure taken from Paper IV.) 17

4.1 Stratospheric drivers of Northern Hemispheric surface extremes. (Figure adapted from Domeisen and Butler (2020). 19
5.1 Schematic representation of homoskedastic (a) and heteroskedastic (b) data where x is the predictor and y the response variable. (Figure taken from Paper I) 24

5.2 Common EOF analysis in T-mode applied to ERA5 T2m, SLP, Z500 and Z10 anomalies showing the leading patterns of T2m (middle left), SLP (middle right), Z500 (bottom left) and Z10 (bottom right) anomalies. (Figure taken from Paper II) 25

5.3 QR of T2m for the lower quantiles a) 0.01, b) 0.05 and the median c) 0.5 based on ERA5 reanalysis data from 1979 to 2019. Figure taken from Paper III 26
1. Introduction

The inherent highly chaotic nature of the atmosphere imposes significant challenges to predicting the weather across the full spectrum of time-scales. As illustrated in Figure 1.1 in green, the prediction skill or, in other words, the reliability of short-range forecasts of individual weather events decreases drastically within two weeks. While it is currently unattainable to predict day-to-day weather conditions over the next few months, estimating long-lead seasonal outlooks (see Fig. 1.1 blue), such as the El Niño/Southern Oscillation, is achievable (Mariotti et al. 2018).

The gap between these two time-scales, ranging from approximately two weeks to two months, is commonly referred to as the sub-seasonal to seasonal (S2S) prediction gap (illustrated in orange in Figure 1.1). Accurately forecasting atmospheric phenomena on S2S time-scales, particularly extreme weather events, has posed significant difficulties to the scientific community. Traditional weather forecast models that use an initial value problem approach would need to be run out long into the future. At the same time, coupled climate models, which
heavily rely on boundary conditions and are typically tailored for seasonal outlooks, would require finer temporal and spatial resolution to effectively capture extremes. Both of these model configurations strain current-day computational capabilities to their maximum capacities, emphasizing the necessity for a specialized forecasting framework. Bridging the S2S prediction gap promises substantial and essential advantages for today’s society. It enhances preparedness for extreme weather events and facilitates early warning systems, thereby significantly boosting the ability to protect against such occurrences (Mariotti et al., 2018). Thus, extensive research is dedicated to advancing S2S prediction through technical improvements as well as a better understanding of atmospheric phenomena that act on the time-scales in question (Vitart and Anderson, 2012).

The inherent interplay of all components of the Earth system offers a valuable opportunity for forecasting. Especially, when one component varies at a slower pace compared to another, it may exert a persistent influence on the latter. Specifically, the slower-varying components of the Earth system, such as the land and the ocean, have the potential to contribute to the prediction of S2S extreme events. The atmosphere, however, is a more complex medium. Particularly, the lowermost layer of the atmosphere, i.e. the troposphere, which is home to most weather phenomena, is predominantly driven by quickly evolving baroclinic waves and thus, characterized by rapid fluctuations. The variability in the stratosphere, however, dominantly arises from wave-mean flow interactions, meaning that relative to the troposphere, it evolves at a slower pace. Thereby, the interaction between the stratosphere and the troposphere has attracted increased scientific interest, especially from the S2S community.

One major challenge in deriving information about the coupling between the stratosphere and troposphere has been and still is the limited availability of years with fine-grained stratospheric observations. Before 1979, which marks the start of the satellite era, stratospheric measurements were done by balloons and radiosondes only at selected locations. Moreover, it’s only been in recent decades that the modeling community has recognized the significance of integrating a more finely resolved stratosphere (Gerber et al., 2012). Nowadays, a combination of balloons, rocket sondes, and remote sensing techniques through ground-based and satellite measurements leads to improved data availability (Braesicke, 2015). With the additional increase of computational power, researchers have shown that the stratosphere can indeed exert an influence on tropospheric mid-latitude weather for a period of up to two months (e.g. Baldwin and Dunkerton, 1999, 2001), thus encompassing S2S-timescales. Various extreme weather events, including strong winter storms, droughts, and severe cold spells in the mid-latitudes, can be attributed to and linked with stratospheric dynamics and their subsequent downward influences. Boreal
winter cold snaps in eastern North America or Eurasia often follow an extreme state of the stratospheric large-scale circulation. Given the significant stress these cold events induce on society, further investigation of this complex link is crucial. Emerging results have fundamental potential to improve extended-range predictability of cold spells, a highly sought-after proficiency by various industrial sectors and society.

This research project revolves around the relationship between the stratospheric and tropospheric circulation in the Northern Hemisphere during winter, specifically focusing on cold events in the Eurasian and North American sectors. Employing several statistical methods and diverse sets of observational and model data, we approach the project from various angles and a unique bottom-to-top perspective. Following this introduction in Chapter 1, Chapter 2 explains the characteristics of the stratosphere and its dynamic variability. Chapter 3 provides information on stratosphere-troposphere coupling processes. Potential surface impacts of extreme stratospheric states are demonstrated in Chapter 4. Concluding, a summary of Paper I-IV is given in Chapter 5, followed by an outlook on potential future research in Chapter 6.
2. The Stratosphere

2.1 Essential Characteristics of the Stratosphere

The Earth’s atmosphere is composed of several layers, which, from the ground up, are troposphere, stratosphere, mesosphere, thermosphere, and exosphere. The four lowermost layers are defined based on their distinct vertical temperature profiles, illustrated in Figure 2.1. Weather systems and convection dominate the well-mixed troposphere that spans from the ground to an altitude of approximately 8km to 15km (depending on latitude). Its characteristic decrease in temperature with increasing height up to the tropopause is clearly visible in Figure 2.1. The stratosphere, situated between the tropopause and an altitude of approximately 50km, belongs to the middle atmosphere.

![Figure 2.1: Layers of the atmosphere and characteristic temperature profile based on the U.S. Standard Atmosphere. The temperature [K] is on the x-axis and the altitude [km; hPa] is on the y-axis. Figure taken from Wallace and Hobbs (2006). Reprinted with permission from Elsevier.](image-url)
The strong stratification, winds, dryness, and low density within this layer effectively hinder the formation and intrusion of weather systems (e.g. Kidston et al. 2015). Consequently, its dynamic variability is considerably lower when compared to the scales observed in the troposphere. Further, the stratosphere is home to the ozone layer, a vital shield that protects the Earth from harmful radiation. The absorption of ultraviolet radiation emitted from the sun achieves this protective function and generates heat within the ozone layer. During boreal winter, however, the polar stratosphere receives limited sunlight. The Subsequently, the intermittent heat generation results in low stratospheric temperatures. Radiative cooling, through the emission of infrared radiation by gases such as water vapor, ozone, and carbon dioxide, yields further cooling (e.g. Waugh and Polvani 2010; Waugh et al. 2017).

2.2 The Stratospheric Polar Vortex and its Evolution

Large-scale motions in the stratosphere are approximately quasi-geostrophic, meaning that pressure gradient and Coriolis force are nearly balanced. During winter, the main driver of the characteristic westerly geostrophic wind is the pronounced temperature difference between polar and lower latitudes that creates a meridional pressure gradient. In combination with sinking motions over the polar cap, a movement of air towards the pole is initiated. This air mass is deflected eastward due to the Coriolis effect. In other words, the meridional temperature gradient balances the thermal wind, which essentially characterizes the change of the geostrophic wind with height. This so called thermal wind balance is represented by Eqn. 2.1 where \( u_g \) is the geostrophic wind, \( p \) is the pressure, \( R \) is the ideal gas constant, \( f \) is the Coriolis parameter and \( T \) is the temperature. It exemplifies that a more pronounced meridional temperature gradient \( \frac{\partial T}{\partial y} \) corresponds to a stronger vertical geostrophic wind shear \( \frac{\partial u_g}{\partial p} \).

\[
\frac{\partial u_g}{\partial p} = \frac{R}{fp} \frac{\partial T}{\partial y}
\]  

(2.1)

Figure 2.2 shows the latitude-height profile of the climatological zonal mean zonal wind (shading) and temperature (contours). The data are based on 9-day running means of the Japanese 55-year Reanalysis (JRA-55) from the Japanese Meteorological Agency (JMA) (Harada et al. 2016; Kobayashi et al. 2015), covering the months December, January and February (DJF) from 1991 to 2020. The notable horizontal temperature gradient between the warmer tropical and colder high-latitude stratosphere has a substantial dynamic consequence. According to the above theory, strong stratospheric circumpolar winds (blue) emerge in the winter hemisphere in response (e.g. Waugh and Polvani 2010).
The strength and position of the westerly wind band, referred to as the polar vortex or stratospheric polar vortex in the subsequent text, represents the primary mode of variability in the stratosphere (Butchart et al., 2011). It sets itself apart from the jet stream in the troposphere. The tropospheric peak of the climatological zonal mean zonal wind speed, is located at about 30° N and 300 hPa (see Fig. 2.2). The polar vortex, on the other hand, is situated further north at around 60° N (e.g., Waugh et al., 2017).

The latter extends from the tropopause to the stratopause with wind speeds of more than 45 ms$^{-1}$ in the zonal mean at approximately 50 km height. Typically, the sharp meridional gradient in wind speed acts as a barrier, isolating the cold polar air from the mid-latitude stratosphere (O’Neill et al., 2015). The boreal stratospheric polar vortex is present from about November to March (e.g., Schoeberl and Newman, 2015) and undergoes several stages throughout its presence. After the autumn equinox, marking the onset of the polar night, temperatures in the polar stratosphere begin to decline. Subsequently, the development of the polar vortex starts, typically reaching its climatological state some time in November. Usually, January is the month with fastest wind speeds and coldest polar temperatures in the stratosphere (Schoeberl and Newman, 2015). With the sunlight returning in spring, the meridional

![Image](image_url)
temperature gradient gradually relaxes. Eventually, a powerful dynamic event, referred to as final warming in the literature (e.g., Butler et al., 2015), rounds off the season. Subsequently, winds return to their summer easterly state. It is essential to note that deviations from the climatological evolution of the polar vortex constitute a significant part of its year-to-year variability. In fact, extreme states of the vortex hold the highest benefits for S2S forecasting.

2.3 Stratospheric Polar Vortex Extremes

Polar vortex extremes are of particular interest, as their signal may translate downward to the troposphere, influencing the circulation for up to two months (Baldwin and Dunkerton, 1999, 2001). These extreme states can be directly related to planetary wave propagation from the troposphere to the stratosphere and their interaction with the polar vortex, which will be explained in more detail in the following.

Atmospheric waves are deviations from the typical zonal (east-west), meridional (north-south), or vertical symmetries of the atmospheric flow. They introduce variations in, for example, atmospheric pressure, and can be of different origin and size and may propagate vertically, horizontally, or be stationary. Rossby waves are a specific type of wave occurring in rotating fluids, such as the Earth’s atmosphere. Typically, they have long wavelengths of thousands of kilometers and move relatively slowly. The rotation and curvature of the Earth and the concomitant Coriolis effect that varies with latitude, play a fundamental role in their generation and behaviour. This is because their restoring force is the north-south gradient of the background potential vorticity (PV). PV is an important conserved quantity under adiabatic and frictionless conditions. It describes a combination of the effects of spin and stratification on a fluid. Briefly, in a barotropic atmosphere, PV is given by the sum of the planetary \( f \) and relative \( \zeta \) vorticity, divided by the column depth \( H \).

\[
PV = \frac{f + \zeta}{H}. \tag{2.2}
\]

Climatologically, there is a gradual decline of PV from the pole toward the equator. When a parcel is displaced into a region with a higher \( f \) value (poleward), its relative vorticity decreases, leading to a tendency toward a clockwise movement. Conversely, a displacement into a region with a lower \( f \) value (equatorward) results in a counterclockwise motion of the parcel. Lower boundary conditions in the Northern Hemisphere, including large-scale topographical features and variations in thermal forcing due to the distribution of the ocean and land mass, excite quasi-stationary planetary-scale Rossby
waves in the troposphere (e.g. Schoeberl and Newman, 2015). The Charney-Drazin criterion (Charney and Drazin, 1961), given by Equation 2.3, indicates under which conditions those waves propagate vertically into the stratosphere. Here, $U$ represents the zonal background wind speed, $U_c$ is the critical background wind speed, $\beta$ refers to the beta-parameter, while $k$ and $l$ stand for the horizontal and meridional wave number (WN), respectively.

$$0 < U < U_c \quad \text{with} \quad U_c = \frac{\beta}{k^2 + l^2} \quad (2.3)$$

From the criterion it is clear that stationary waves only propagate vertically into moderate westerly (positive) background flow with $U$ that is slower than the critical background wind speed $U_c$. Further, a longer horizontal wavelength of the planetary wave, i.e., smaller zonal or meridional WN, results in an increased $U_c$. In that case, the background velocity window for which vertical propagation is possible is wider. Hence, if the stratospheric polar vortex is moderately westerly relative to the phase velocity of the wave, vertical propagation is possible. Typically, these large-scale waves have zonal WN1 and WN2 (Bancalá et al., 2012; Harada and Hirooka, 2017; Kidson et al., 2015) and only occasionally WN3 (Bancalá et al., 2012). Synoptic scale disturbances with higher WNs are trapped in the troposphere. When these large-scale planetary waves propagate upward, they increase in amplitude due to decreasing density with height to conserve energy (e.g. O’Neill et al., 2015). Approaching the critical level, where the zonal mean zonal wind and the zonal phase speed of the upward propagating waves are the same (or $0 \text{ms}^{-1}$ for stationary waves), the waves break and dissipate (e.g. Matsuno, 1971; Schoeberl and Newman, 2015). Thereby, a drag in the form of easterly momentum is transferred to the westerly flow (McIntyre and Palmer, 1983; Polvani and Waugh, 2004), leading to a decelerated and perturbed polar vortex. This wave drag can be evaluated by the divergence of the Eliassen-Palm (EP) flux (Eliassen and Palm, 1961) in the quasi-geostrophic approximation. Based on Edmon et al. (1980), the EP flux vector and its divergence in spherical coordinates and on the meridional plane are given by Equation 2.4 and Equation 2.5

$$\left( \begin{array}{c} F(\varphi) \\ F(p) \end{array} \right) = a \cos \varphi \left( \begin{array}{c} -\frac{u'v'}{\varphi f} \\ \frac{\theta f}{r^2} \end{array} \right) \quad (2.4)$$

$$\nabla \cdot \mathbf{F} = \frac{1}{a \cos \varphi} \frac{\partial}{\partial \varphi} \left( F(\varphi) \cos \varphi \right) + \frac{\partial}{\partial p} \left( F(p) \right). \quad (2.5)$$

Overbars and primes refer to the zonal mean and departures from it, respectively. Moreover, $a$ is the Earth’s radius, $\varphi$ is the latitude, $f$ is the Coriolis parameter, $\theta$ is the potential temperature, and $u$ and $v$ are the zonal and meridional wind,
respectively. The EP flux vector has two components that are proportional to the meridional flux of momentum (meridional component) and the meridional flux of heat (vertical component). Convergence of the EP flux indicates an acceleration of the zonal mean zonal wind while divergence reveals its deceleration. Further, the group velocity and thus, the direction of propagating planetary wave packets, is proportional to the direction of the EP flux vector.

The stratospheric polar vortex’s large-scale geostrophic winds follow the lines of constant geopotential height. Figure 2.3 shows the boreal winter (January-March) climatological (1979-2022) geopotential height (contours) and temperature (shading) in 10hPa based on ERA5 reanalysis data [Hersbach et al., 2020] that is produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). Low geopotential heights, associated with the center of the polar vortex, are located over Svalbard. The lowest temperatures reside on the eastern edge of the vortex.

**Figure 2.3:** Climatological 10hPa geopotential height in gpdam (contours) and temperature in °C (shading) based on the ERA5 reanalysis from 1979-2022 for January-March.

If upward planetary wave propagation and subsequent wave absorption is weak, the stratospheric zonal winds are undisturbed and reach peak strengths. In mid-March 2019, for example, the stratospheric polar vortex was exceptionally strong. The 10hPa geopotential heights (contours) and temperature anomalies (shading) for March, 15th 2019 are shown in Figure 2.4 (top) and can be compared to the climatology (Fig. 2.3). The negative anomalies indicate
remarkably cold temperatures and the geopotential heights are extremely low. Both variables are centered over the pole with densely packed isolines of the latter, suggesting a quasi-zonal arrangement of the polar vortex. On the other hand, if upward planetary wave propagation and subsequent absorption in the stratosphere is enhanced, the polar vortex may reduce in strength. A significant reduction can lead to an event known as Sudden Stratospheric Warming (SSW) \cite{Butler2015, Matuno1971}, often regarded as one of the most dramatic dynamical phenomena in the middle atmosphere that is associated with a rapid amplification of planetary waves. Characteristically, a zonal mean zonal wind reversal from westerly to easterly (commonly evaluated at 10hPa and 60°N \cite{Butler2015}) can be observed. In addition, as the name suggests, the mid-to upper stratospheric polar cap temperature increases sharply by several tens of degrees within just a couple of days. Eventually, the meridional temperature gradient reverses.

\begin{figure}[h]
\centering
\includegraphics[width=0.45\textwidth]{figure2a.png}
\includegraphics[width=0.45\textwidth]{figure2b.png}
\caption{10hPa geopotential height in gpdam (contours) and temperature anomaly (shading) based on ERA5 reanalysis data from 1979-2022 for March 15th 2019 (top), January 1st 2019 (bottom left), January 15th 2019 (bottom right).}
\end{figure}
An increase in temperature without the reversal in zonal mean zonal wind is called minor and otherwise major SSW. While there are several definitions of major and minor warmings, the ones explained above are the official ones given by the World Meteorological Organization. Nonetheless, climatological features of SSWs across definitions show only minor differences (Palmeiro et al., 2015). On average, major SSWs persist for 14 days based on the time the zonal mean zonal wind is easterly (Lee and Butler, 2019) and, statistically, happen every other winter. While these events can be generally predicted with lead times of 10-20 days, this time range may vary on an event-to-event basis (e.g. Tripathi et al., 2015). The 2019 season was quite remarkable. In addition to the example of an extremely strong polar vortex event mentioned earlier, an unusual SSW happened in the beginning of the year. Exemplary, Figure 2.4 (bottom left) shows the 10hPa geopotential height (contours) and temperature anomalies (shading) on the day of zonal mean zonal wind reversal, i.e. January 1st, 2019 (e.g. Paper IV). Positive polar cap anomalies indicate the drastic increase in temperature and, compared to the climatology (Fig. 2.3), the meridional temperature gradient is reversed. Further, the Aleutian High extends up into the stratosphere, facilitating a displacement of the low geopotential heights towards Eurasia and the North Atlantic sector. Subsequently, a WN1 structure appears. Some SSW events are associated with a pure displacement of the vortex. However, the polar vortex may even split. After preconditioning, i.e., vortex stretching, the polar vortex eventually tears apart, resulting in the emergence of two or more vortices. Commonly, this is imprinted in the data as a WN2 pattern or, occasionally, a WN3 pattern. Ehrmann and Colucci (2019) find vortex displacements to happen twice as often as well as to be twice as persistent compared to split events. The latter are most common in January and February (Charlton and Polvani, 2007) and are less predictable compared to the former type (e.g. Taguchi, 2018). Exemplary, in 2019, an unusual WN3 pulse resulted in a polar vortex split in January. Figure 2.4 (bottom right) shows the 10hPa geopotential heights (contours) and temperature anomalies (shading) for January, 15th 2019. Two centers of low geopotential heights, one over North America and one over Eurasia, are clearly visible. Further, stratospheric air masses over the polar cap are warmer than usual. Eventually, the polar vortex radiatively relaxes and recovers towards its climatological winter state.
3. Stratosphere-Troposphere Coupling

The stratosphere and the troposphere are not just two layers of the atmosphere situated on top of each other, acting independently. Instead, a dynamic coupling can be observed from about November to April that potentially leads to an impact of the stratosphere on the tropospheric circulation. The exact mechanisms at work are not yet fully understood. The dynamic coupling is often described in a zonal mean sense, such as the impact on the Northern Annular Mode (NAM), also called Arctic Oscillation (AO). The AO is the leading mode of variability in the Northern Hemisphere and influences the weather substantially (e.g. Thompson and Wallace, 1998) by setting the tone for the behavior of the mid-latitude jet stream. It essentially describes a quasi-zonally-symmetric, meridional see-saw of sea level pressure (SLP) anomalies between polar- and mid-latitudes. Figure 3.1 taken from the National Oceanographic and Atmospheric Administration (NOAA), schematically shows the two phases of the AO and how they may affect the troposphere dynamically.

![Figure 3.1: Schematic representation of the positive (left) and negative (right) phase of the AO. (Figure provided by NOAA.)](image)

The positive phase (Fig. 3.1, left) refers to low polar SLP and a quasi-zonal...
arrangement of the tropospheric jet stream that locks the cold polar air in high
latitudes. The negative AO phase (right), on the other hand, describes high polar
SLP and weaker zonal winds. In that case, the wavy tropospheric jet stream
allows cold polar air to advect further southward into the mid-latitudes, while
warmer mid-latitude air is transported poleward.
In case of an SSW or a particularly strong polar vortex (SPV), the stratospheric
imprint of the NAM shifts towards the negative or positive state, respectively.
These signals may progress downward into the troposphere and reach the
surface (Baldwin and Dunkerton[1999]2001). Figure 3.2 taken from Baldwin and
Dunkerton [2001], depicts a composite of the time-height development
of the NAM for weak (A) and strong (B) vortex events. Red colors indicate
negative values, while blue refers to positive ones. A lag of 0 days corresponds
to the day the NAM index crosses -3.0 and +1.5, respectively.

In both cases, the signal initially occurs in the stratosphere around 10hPa and
progresses down towards the tropopause (approximated by the horizontal line).
After reaching the tropopause, the signal "drips down" into the troposphere,
hence the common name "dripping paint plot" for these types of Figures.
Physically, Kuroda and Kodera [1999], for example, suggest that the grad-
ual downward progression of the weakened stratospheric winds is related to
changes in the stratospheric background flow through wave absorption. The
subsequent altered wave propagation properties drag the mean-flow signal down.
More specifically, as specified in Chapter 2.3, upward traveling waves may only propagate through layers that meet the Charney-Drazin criterion ([Charney and Drazin](1961)), given in Equation 2.3. Once they reach the critical layer, they break and deposit momentum, thereby changing the characteristics of the background flow of the layer just below the critical one. Thus, the critical layer is successively dragged down, causing the following waves to break at progressively lower levels (e.g. [Kuroda and Kodera](1999); [Schoeberl and Newman](2015)). This downward progression is not only visible in the zonal wind, but its imprint is observable in temperature and pressure fields. Nonetheless, linear theory cannot explain nonlinear processes that are involved. Particularly, in the highly variable extratropical troposphere, the progression of the signal from the stratosphere toward the surface is more diffusive and dependent on the tropospheric state. Weather systems and tropospheric eddies may strongly influence the downward progressing anomalies and, in turn, also be influenced or even generated through the stratospheric signal (e.g. [Song and Robinson](2004)).

Instead of being absorbed by the mean flow, the upward propagating planetary waves may be refracted or reflected back down by the stratosphere. Thereby, the stratosphere may alter the tropospheric circulation indirectly. Firstly, a direct consequence of wave reflection is a decrease in upward wave fluxes. The resulting reduction in wave absorption hinders the downward progression of the NAM. Further, the downward reflected waves may interact with tropospheric ones by influencing, for example, their meridional orientation and their amplitudes. This interaction may, subsequently, cause circulation changes in the troposphere (e.g. [Perlwitz and Harnik](2003, 2004)). Note, however, that finding significant effects of stratospheric wave reflection is not an easy task as the dominant sources of tropospheric waves are tropospheric dynamics itself.

The allocation of a stratospheric reflective layer is strongly dependent on the structure of the vertical and meridional background wind. Generally, reflecting surfaces form in case the meridional PV gradient above the jet peak becomes weak or negative due to a vertical curvature in the zonal and vertical wind profiles and negative vertical wind shear (e.g. [Matsuno](1970); [Perlwitz and Harnik](2003, 2004)). One possible region that meets these criteria is within the negative wind shear above the polar vortex maximum that is co-located with a minimum in the meridional PV gradient ([Matsuno](1970)). Further, the waves need to be hindered from dispersing in the meridional direction by a meridional waveguide that forms through a meridional curvature of the wind. Corresponding to the above, the high-latitude vertical wind shear in the stratosphere may indicate whether the stratospheric basic state is reflective or not. The reflective index (RI) provides for a broad evaluation and is, essentially, given by the vertical wind shear between 2hPa and 10hPa averaged over the mid to high latitudes
(58°N–74°N) ([Perlwitz and Harnik] 2003, 2004). Following [Nath et al.] (2014), the zonal variation of the RI over time is given by Equation 3.1:

\[
RI(\lambda, t) = U_2(\lambda, t) - U_{10}(\lambda, t),
\]

where \(U_2\) and \(U_{10}\) are the zonal wind in 2hPa and 10hPa, \(\lambda\) the longitude and \(t\) the time. Negative values indicate negative wind shear, translating to a negative meridional gradient of PV, and, broadly speaking, to the formation of a reflective layer. Positive RI values refer to increasing wind with increasing height, which indicates a non-reflective basic state. It is commonly believed that wave reflection predominantly occurs during periods of strong vortex conditions (e.g. [Perlwitz and Harnik] 2003, 2004). Note that the RI is often negative during or just before SSWs without a reflective surface to form. Nevertheless, [Kodera et al.] (2013) and [Kodera et al.] (2016) find wave reflection to also be possible during the recovery phase of an SSW, leading to an abrupt ending of the stratospheric warming. Compared to the wave-mean flow interaction described above, the downward interaction through wave reflection happens on shorter time-scales in the order of a week (e.g. [Perlwitz and Harnik] 2004).

A way of evaluating the movement of wave packets, i.e., the energy the wave transports, is the three-dimensional Plumb wave activity flux (WAF) ([Plumb] 1986). It directly visualises the wave path and is given by the WAF vector \(\mathbf{F} = (F_x, F_y, F_z)^T\):

\[
\mathbf{F} = p \cos \phi \begin{pmatrix}
\frac{1}{2a^2 \cos^2 \phi} \left( \frac{\partial \psi}{\partial \lambda} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \lambda^2} \\
\frac{1}{2a^2 \cos^2 \phi} \frac{\partial \psi'}{\partial \lambda} \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} \\
\frac{2a^2 \sin^2 \phi}{N^2 a \cos \phi} \frac{\partial \psi'}{\partial \lambda} \frac{\partial^2 \psi'}{\partial \lambda \partial z} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial z}
\end{pmatrix}.
\]

Primes denote departures from the zonal mean, and \(a, \Omega, \psi, f\) and \(N\) represent respectively the Earth’s radius, the Earth’s rotation rate, the three-dimensional quasi-geostrophic stream function, the Coriolis parameter, and the Brunt-Väisälä frequency. Further, \(p\) is the scaled pressure (hPa/1000hPa) and \(\phi\) and \(\lambda\) are latitude and longitude, respectively. By definition, the wave packets follow the geopotential height gradient. Figure 3.3 shows an example of the 3-day running mean geopotential height anomalies with respect to the zonal mean (contours), the WAF in the x-z-plane (arrows), both meridionally averaged (48°N–74°N), and the zonal wind at 60°N (shading) for January 15th, 2019 based on JRA-55 data (Figure taken from Paper IV). Between 180°E and 90°W the structure of the positive geopotential height anomaly features a westward tilt with height on its western flank and an eastward tilt on the eastern one. This configuration indicates upward and downward propagation of wave packets,
respectively (see arrows). In this case, the upward traveling waves are reflected in the lower stratosphere and channeled back down into the troposphere. 

Figure 3.3: Geopotential height anomalies with respect to the zonal mean (contours), the WAF in the x-z-plane (arrows), both meridionally averaged (48°N-74°N), and the zonal wind at 60°N (shading) for January 15th, 2019 based on JRA-55 data. (Figure taken from Paper IV.)

In recent decades, scientific interest in the exact mechanisms of the stratosphere-troposphere coupling increased because a better understanding helps towards enhanced S2S predictability of extremes. While not every stratospheric event has consequences for the tropospheric circulation (e.g. Karpechko et al., 2017), some may persistently influence tropospheric weather (e.g. Baldwin and Dunkerton, 2001), and can thus be valuable harbingers for our weather.
4. Tropospheric Response to Stratospheric Variability

In Chapter 3, the dynamic coupling between the stratosphere and the troposphere has been described. This Chapter discusses its impact on the tropospheric circulation and surface weather, particularly regarding extreme events (e.g. Baldwin and Dunkerton, 2001; Cohen et al., 2010; Finke and Hannachi, 2022; Finke et al., 2023; Kidston et al., 2015; Kolstad et al., 2010; Kretschmer et al., 2018a; Thompson et al., 2002; Woollings et al., 2010). Figure 4.1 was adapted from Domeisen and Butler (2020). It lists a summary of tropospheric extreme events in the Northern Hemisphere, which are associated with stratospheric drivers such as SSWs, SPV events and wave reflection, as well as their societal impacts and affected regions. Various extremes, including storms, droughts, and cold spells, are associated with stratospheric variability. The immense consequences for society span from flooding, infrastructure and wind damage to impacts on shipping, agriculture and health (Domeisen and Butler, 2020).

<table>
<thead>
<tr>
<th>stratospheric precursor</th>
<th>tropospheric extreme</th>
<th>impact</th>
<th>affected region</th>
</tr>
</thead>
<tbody>
<tr>
<td>sudden stratospheric warming</td>
<td>(marine) cold air outbreak</td>
<td>infrastructure damage, health impacts</td>
<td>Arctic, northern Europe, North Atlantic</td>
</tr>
<tr>
<td></td>
<td>increased storminess</td>
<td>flooding, wind damage</td>
<td>southern Europe</td>
</tr>
<tr>
<td></td>
<td>regional sea ice changes</td>
<td>shipping impacts, resource extraction</td>
<td>Arctic</td>
</tr>
<tr>
<td>strong vortex event</td>
<td>storm series</td>
<td>flooding, wind damage</td>
<td>northern Europe, North Atlantic</td>
</tr>
<tr>
<td></td>
<td>drought</td>
<td>agricultural damage</td>
<td>southern Europe</td>
</tr>
<tr>
<td>wave reflection</td>
<td>cold air outbreak</td>
<td>health impacts</td>
<td>North America</td>
</tr>
</tbody>
</table>

**Figure 4.1:** Stratospheric drivers of Northern Hemispheric surface extremes. (Figure adapted from Domeisen and Butler, 2020)

Due to the effect of the stratosphere on the NAM (e.g. Baldwin and Dunkerton, 2001; Black, 2002), surface weather is ultimately affected (e.g. Scaife et al., 19...
The strongest coupling to the troposphere and the surface is usually observed at the locations of the storm tracks, particularly over the North Atlantic (e.g. Garfinkel et al., 2013; Iqbal et al., 2019). This translates to an influence on the North Atlantic Oscillation (NAO) (Hurrel, 1995; Wallace, 2000), which is a see-saw of SLP between the Icelandic Low and the Azores High. Thereby, the latitudinal position of the tropospheric jet stream (e.g. Iqbal et al., 2019) and storm track (e.g. Baldwin and Dunkerton, 2001) is affected.

The focus of the research included in this PhD thesis is on Northern Hemispheric mid-latitude cold spells, i.e., extended periods with unusually cold temperatures, as stratosphere-troposphere coupling may offer feasible predictability for these events (Kidston et al., 2015; Shtanina et al., 2022; Thompson et al., 2002). The exact involved mechanisms are, however, still a subject of ongoing research and the prediction skill varies hugely on an event-to-event basis. Generally, cold temperatures may arise in eastern Asia, northern Europe (e.g. Baldwin and Dunkerton, 2001; Chen et al., 2005; Kolstad et al., 2010; Thompson et al., 2002) and eastern North America in the aftermath of an SSW. Opposite temperature signs have been related to SPV events (Thompson et al., 2002). However, it has to be noted that the surface response may differ for different types of SSWs (e.g. Mitchell et al., 2013; White et al., 2021) and that the impact is dependent on the tropospheric background state into which the stratospheric signal progresses (Afargan-Gerstman et al., 2020).

Regionally more specific, after an SSW, the frequency and strength of cold air outbreaks may increase in the northern part of Eurasia (Kretschmer et al., 2018a; Scaife et al., 2008) and eastern North America (Kretschmer et al., 2018b; Thompson and Wallace, 2001). Cold spells in these two regions have been analyzed in Paper III and Paper IV, respectively. Eurasian temperatures are under local as well as remote influences. A local mechanism, leading to cold temperatures in winter is linked to a reduction of sea ice in the Arctic Ocean. This reduction contributes to an intensification of geopotential heights over Siberia (e.g. Honda et al., 2009; Panagiotopoulos et al., 2005; Wu et al., 2011) and a blocked flow over the Ural region (e.g. Yao et al., 2017). Subsequently, this pattern facilitates the movement of cold air masses toward Eurasia (Mori et al., 2014). Besides this local influence, the North Atlantic circulation has consequences for Eurasian weather (e.g. Barnett et al., 1984). For example, the negative phase of the NAO, which often develops in response to SSWs, has been related to cold surface temperature anomalies over Northern Europe and Eurasia (see Paper I-III and e.g. King et al., 2019; Kretschmer et al., 2018a,b; Thompson and Wallace, 2001). In addition, warming over the Arctic reduces the meridional temperature and geopotential height gradient. Consequently, the propagation speed of large-amplitude Rossby waves is decreased over the
North Atlantic which may facilitate more persistence of cold spells that develop downstream \cite{Francis2012}. The mechanism of planetary wave reflection in the stratosphere has been adequately documented in the literature, yet its contribution to promoting cold spells has yet received limited investigation. However, it has gained increased attention in recent years. Using statistical analyses with reanalysis data, cold spells over eastern North America have been related to wave reflection events \cite{Kretschmer2018, Matthias2020, Messori2022}. Several studies find a connection of an anomalous Alaskan ridge \cite{Matthias2020, Messori2022} and the negative phase of the Western Pacific Oscillation \cite{Kretschmer2018} to wave reflection toward eastern North America and subsequent cold surface temperatures in that region. These studies find this link to rather occur under neutral or strong stratospheric polar vortex conditions. In particular, a strong Aleutian High in the stratosphere reflects the upward propagating waves downward towards Canada \cite{Matthias2020, Messori2022} which yields a development of a strong ridge/trough pair over the eastern North Pacific and North America \cite{Kodera2013}. This circulation configuration subsequently promotes cold air advection towards eastern North America. The described mechanism has, however, also been found to similarly occur in the aftermath of SSW \cite{Kodera2013}.

Note that cold air outbreaks are inherent to tropospheric variability and may occur without stratospheric involvement \cite{Waugh2017}. However, if there is a stratospheric influence, it is likely that their predictability is increased, yielding a potential for better preparedness and mitigation strategies by society.
5. Research Summary

This research project consists of several parts, jointly aiming for an enhanced understanding of the connection between Northern Hemispheric cold spells and stratospheric variability. Common empirical orthogonal function (EOF) analysis in Paper II and quantile regression (QR) in Paper I are two sophisticated statistical methods. Both facilitate the path to efficiently analysing the data needed for a better understanding of the stratosphere-troposphere link. The connection between surface temperature and stratospheric variability is complex, not least due to the inherent internal variability of the troposphere. In Paper I and Paper III, we approach the task from a unique bottom-up perspective, i.e., focusing on surface temperature extremes before linking them to the stratosphere instead of facing the problem from a top-down angle, commonly used in previous studies. The identified distinct lagged effect in extreme surface temperatures to changes in the stratosphere gives straightforward access to invaluable information usable for S2S prediction. A case study of the link between the cold North American winter of 2018/2019 and the early 2019 SSW in Paper IV complements the project. Summaries of Paper I-IV are given in the following.

Paper I: Tropospheric Response to Stratospheric Variability via Lagged Quantile Regression

The effect of stratospheric variability on the tropospheric circulation and the surface weather is a widely researched topic since the mechanisms operate on S2S time scales. Standard linear regression (LR) is often used as the simplest way to find a relation between two or more variables. However, LR is based on the conditional mean of the response variable, meaning that the modeled relation between the predictor and the response variable is based on the mean of the response variables distribution. While this is not an issue for data with constant variance and few extremes, LR is unsuitable if the response in outer quantiles differs from the mean response. This is schematically shown in Figure 5.1. Homoskedastic data (a) has constant variance across the distribution, i.e., the data in the outer quantiles behave similar to the mean. For heteroskedastic data (b), the variance is not constant. Then, the relation between the predictor
(x-axis) and the response variable (y-axis) in the outer quantiles is different to the one in the mean.

Figure 5.1: Schematic representation of homoskedastic (a) and heteroskedastic (b) data where x is the predictor and y the response variable. (Figure taken from Paper I)

In this project, the focus is on the stratospheric connection to tropospheric extreme events. As extremes are depicted in the outer quantiles, it is difficult to simply model this relation by looking at the mean when the data is heteroskedastic. QR bypasses this shortcoming by modeling the whole conditional distribution of the response variable. Applying the method to the ERA5 reanalysis, the connection between teleconnection indices, such as the NAO and AO index, as well as gridded mid-tropospheric and surface fields to the boreal winter stratosphere is evaluated. We find significant differences in duration, strength, and direction of the relation between the stratosphere and the troposphere across the response variable’s quantiles. Focusing on S2S time-scales, the link between tropospheric extremes and stratospheric variability is analysed using a lagged version of the QR model. Thereby, we gain invaluable additional information not derivable by using simple LR.

**Paper II: Common EOFs: a tool for multi-model comparison and evaluation**

Identifying common modes of variability between various variables from given reanalysis data or between climate model simulations at different atmospheric levels is a challenging problem. This includes the identification of their individual explained variance. **Paper II** introduces common EOF analysis with
a step-wise algorithm. The method has the benefit of combining information from, for example, a given variable from different models or different variables from a given dataset. Further, the method can be applied in the S- or T-mode, i.e., identifying common spatial or common temporal patterns across multiple time series. In the S-mode, common EOF analysis is applied to four reanalyses products and to simulations of the Atmospheric Model Intercomparison Project Phase 6 (AMIP6). The AMIP is a companion project of the Coupled Model Intercomparison Project Phase 6 (CMIP6) using atmospheric general circulation models instead of ones coupled to the ocean. Thereby, the complexity introduced by ocean-atmosphere interactions is not considered. Monthly means and winter daily gridded data of several tropospheric and stratospheric fields over the Northern Hemisphere are used. The reanalyses are found to be consistent in the mid-tropospheric and stratospheric fields. However, the explained variance of the 2m surface temperature (T2m) shows significant differences across the data sets. On the contrary, stratospheric fields of AMIP6 data exhibit significant deviations in the explained variance of around 25%. In the T-mode, the method was applied to 4 variables (T2m, SLP, 10hPa (Z10), and 500hPa (Z500) geopotential height anomalies) based on ERA5 reanalysis data.

Figure 5.2: Common EOF analysis in T-mode applied to ERA5 T2m, SLP, Z500 and Z10 anomalies showing the leading patterns of T2m (middle left), SLP (middle right), Z500 (bottom left) and Z10 (bottom right) anomalies. (Figure taken from Paper II.)
Figure 5.2 shows anomaly patterns of the leading common mode of T2m (top left), SLP (top right), Z500 (bottom left) and Z10 (bottom right). The results indicate that a weak polar vortex (positive Z10 anomalies) is associated with cold surface temperature anomalies over northern Eurasia and Siberia, with a signature of the negative phase of the NAO/AO (positive SLP and Z500 anomalies over the North Atlantic/North pole and negative anomalies over the mid-latitudes).

**Paper III: Exceptionally persistent Eurasian cold events and their stratospheric link**

Extreme cold temperatures, which endure for a long time period, strongly affect society. Initially, we applied QR using time as the predictor and gridded T2m as the response variable. Here, ERA5 reanalysis data from 1979-2019 is used. Shown in Figure 5.3, we find a negative trend in the lowermost quantiles (a,b) in the southern region of the Eurasian study area (black box), corresponding to a cooling trend of exceptionally cold temperatures. This trend is not visible for the median (c).

![Figure 5.3: QR of T2m for the lower quantiles a) 0.01, b) 0.05 and the median c) 0.5 based on ERA5 reanalysis data from 1979 to 2019. Figure taken from Paper III.](image)

The canonical impact of SSWs is the negative phase of the AO/NAO and subsequent cold surface temperatures in Eurasia. However, Eurasia has seen persistent cold events (PCE) during the full range of polar vortex variability. Lagged composite analysis based on ERA5 reanalysis data reveals that PCEs, which follow extremely weak (SSW PCE) and strong (SPV PCE) stratospheric winds, exhibit similarities but also differences. SPV PCEs set in more suddenly, have colder temperatures, and are of shorter duration than SSW PCEs. In both cases, a quasi-stationary anticyclone in the mid-troposphere over the Arctic Ocean blocks warm air advection towards Eurasia. For SSW PCEs,
the blocking is related to the downward progression of the negative phase of the AO - a well-known response. In the case of SPV PCEs, a Rossby wave, potentially formed over the North Atlantic, is co-located with negative vertical WAF anomalies related to stratospheric wave reflection. This wave shares a ridge with the anticyclone over the Arctic Ocean. Further, Arctic surface warming and unusually weak zonal winds over the study area likely contribute to its slow eastward progression.

**Paper IV: The stratospheric polar vortex and surface effects: The case of the North American 2018/19 cold winter**

Severely cold surface temperature records in eastern North America, for example, −32°C in Chicago, led to serious societal and economic consequences during the 2018/2019 winter. These temperatures have been considered a consequence of the SSW that happened in early 2019 (e.g. Lee and Butler, 2019). Using JRA-55 data, we find upward propagating planetary waves with zonal WN1-3 to cause the polar vortex to shift away from its climatology (see Figure 2.4 bottom left), resulting in a major SSW. A strong WN3 pulse led to a vortex split with one of the centers residing over eastern North America (see Figure 2.4 bottom right). The configuration of the Aleutian High, imprinted in the geopotential height anomalies with respect to the zonal mean, i.e., the westward tilt with height on its western flank, allowed for upward wave propagation into the stratosphere (see Fig. 3.3). The lower stratospheric eastern edge is eastward tilted with height corresponding to a downward propagation of wave packets towards eastern North America. The latter promotes a ridge-trough pair over the eastern North Pacific and North America. Further, the tropospheric circulation is blocked over the North Pacific, which has been related to wave reflection events in the literature (e.g. Kodera et al., 2013). This setup promoted the cold surface temperatures through cold air advection.
6. Outlook

It is difficult to derive a causal relation between the stratosphere and the troposphere by only using observational data due to the short period of the satellite era and the additional scarcity of stratospheric events. While AMIP6 model simulations show significant differences in the stratospheric fields (see Paper II), it is still a valuable database for increasing statistical power. Thereby, the relation between cold spells over EUR/North America and stratospheric variability that was found in Paper I-IV can additionally be assessed. On the other hand, the substantial variance in the stratospheric fields may hint at difficulties to represent coupling processes such as wave reflection or the accurate position of the polar vortex during, e.g., SSWs. Numerical sensitivity experiments with a climate model, where, for example, reflecting surfaces are manually implemented to specific stratospheric levels, may yield valuable insights into underlying processes of the connection between wave reflection and tropospheric cold spells that have been identified in Paper III and IV.

Processes within the stratosphere can alter the occurrence frequency of stratospheric dynamic extremes. Baldwin and Dunkerton (2001), for example, find that due to the alteration of the waveguide for upward propagating wave packets, i.e., the northward shift of the critical latitude where planetary waves break in the stratosphere, SSWs are twice as likely in case the Quasi-Biennial Oscillation (Ebdon 1960; Reed et al. 1961; QBO) is in its easterly phase compared to westerly. Further, an SPV has three times the chances of occurring if the QBO is westerly (Baldwin and Dunkerton 2001). This may have direct implication for the likelihood of cold spells in Eurasia and North America and would thus be interesting to research in regard to the results of, for example, Paper III. Another possible investigation point could be to focus on dynamical processes in the troposphere that are strongly interconnected with extreme weather events at the surface. For example, breaking Rossby waves in the troposphere can have a major impact on the circulation and can affect atmospheric blocking and subsequent cold snaps (e.g., Martius and Rivière 2016). Stratospheric variability may influence the occurrence frequency of both Rossby wave breaking life cycles, i.e., the anticyclonic and cyclonic type (e.g., Kunz et al. 2009), through the stratospheres impact on the latitudinal position of the tropospheric jet stream. Moreover, the downward reflected planetary waves from the stratosphere may also influence the tropospheric wave breaking process. Hence, an investigation
of the impact of stratospheric reflection events on the Rossby wave breaking life cycles could improve the understanding of the complex processes that are involved.

Finally, S2S predictability can benefit from other atmospheric phenomena such as the Madden-Julian Oscillation (Madden and Julian (1971); MJO), whose teleconnections act on intraseasonal time-scales. It has been shown that the MJO signal in the extra tropics in terms of circulation anomalies (Green and Furtado 2019) as well as its impact on Rossby wave breaking (Finke 2019) is sensitive to the variability of the polar vortex. These findings encourage further research on the influence of the MJO on the connection between the stratosphere and cold surface extremes.
Acknowledgements

Completing this PhD wouldn’t have been achievable without the support of many of my colleagues, my friends, and my family.
First of all, I would like to thank my main supervisor, Abdel Hannachi, for giving me the opportunity to pursue my PhD at MISU. I am very grateful for your continuous support and for always finding time to help and encourage me when needed. You helped me become an independent researcher, and for that, I am immensely thankful.

I would also like to express my gratitude to my second supervisor Jörg Gumbel. You are an inspiring teacher, and I really enjoyed all the discussions in the Middle Atmosphere and Atmospheric Waves course. The visit to the Esrange Space Center in the Arctic was a remarkable and memorable experience.

Additionally, I am very grateful for the support and enriching discussions offered by my co-authors and the scientists I met at conferences and workshops. Special thanks to Toshihiko Hirooka for enabling my research visit to Kyushu University in Fukuoka, Japan. Additionally, I am grateful to Yuya Matsuyama for showing me around the campus and being incredibly helpful. I thoroughly enjoyed my time there.

Further, I am grateful for the input and support of my committee, Annica Ekman and Jonas Hedin. And, of course, thanks to all the administrative staff at MISU and the Bolin Centre. I enjoyed working here and appreciated the welcoming work and social environment.

Special thanks to my office mates Aiden, Thomas, and Joakim for many insightful discussions, Swedish lessons, penguin watching, and the non-work related chitchat. Thanks to Sonja and Björn, as well as Alejandro, Ezra, and Ole Martin, for being great company in the office and beyond. Further, I want to thank Lutz, Matti, Clara, Milla, and the Potatis Klubb, particularly Henni, Vanessa, Dagmar, and Robert, for all the great fun after work.

David, the numerous fun climbing sessions have been a much-needed break from work. Thank you!

Thanks to Jonathan, the time at MISU wouldn’t have been the same without you! Not least because of all the spontaneous road trips to Luleå, Sarek, and the North Kap, the fun outdoor activities and many cooking sessions beyond the PhD work.

Tack, Emil, for always having my back, and, particularly, for cultivating my
green thumb by passing on your ‘Don’t be stressed, just get a new plant’ lifestyle. Your support has been invaluable!
Wolfgang, I really appreciate your constant help and support. Thank you!
Big thanks to my incredible mom! Your endless support, encouragement, and cheers have been invaluable. None of this would have happened without you and I am truly blessed to have you in my corner.
Lastly, a heartfelt thanks to Thorsten for your unwavering support. Our countless silly and non-silly adventures have been an absolute blast, whisking me away from all the work-related stress. I know this sounds a bit cheesy now but you truly are my rock. (: ❤️
References


Mariotti, A., P. M. Ruti, and M. Rixen (2018), Progress in subseasonal to seasonal prediction through a joint weather and climate community effort, *npj Climate and Atmospheric Science*, 1(1), 4, doi: 10.1038/s41612-018-0014-z. xix, 1, 2


