Sudden Stratospheric Warmings

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Key Points:

- Sudden stratospheric warmings (SSWs) are characterized by rapid temperature increases in the winter polar stratosphere (∼10–50km) and a reversal of the climatological westerly winds.
- SSWs affect not just the stratosphere, but the entire atmosphere from the surface to the ionosphere.
- Surface effects of SSWs include shifts of the jet stream, storm tracks, precipitation, and likelihood of cold spells.

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Abstract

Sudden stratospheric warmings (SSWs) are impressive fluid dynamical events in which large and rapid temperature increases in the winter polar stratosphere (∼10–50km) are associated with a complete reversal of the climatological wintertime westerly winds. SSWs are fundamentally caused by the breaking of planetary-scale waves that propagate upwards from the troposphere. During an SSW, the polar vortex breaks down, accompanied by rapid warming of the polar air column. This rapid warming and descent of the polar air column affects tropospheric weather, shifting jet streams, storm tracks, and the Northern Annular Mode (NAM), including increased frequency of cold air outbreaks over North America and Eurasia. SSWs affect the whole atmosphere above the stratosphere producing widespread effects on atmospheric chemistry, temperatures, winds, neutral (non-ionized) particle and electron densities, and electric fields. These effects span the surface to the thermosphere and across both hemispheres. Given their crucial role in the whole atmosphere, SSWs are also seen as a key process to analyze in climate change studies and subseasonal to seasonal predictions. This work reviews the current knowledge on the most important aspects related to SSWs from the historical background to involved dynamical processes, modelling, chemistry and impact on other atmospheric layers.

Plain Language Summary

The stratosphere is the layer of the atmosphere from ∼10–50km, with pressures decreasing to ∼1 hPa (0.1% of surface pressure) at the top. The polar stratosphere during winter is normally very cold, with strong westerly winds. Roughly every two years in the Northern Hemisphere, the quiescent vortex suddenly warms over a week or two, and the winds slow dramatically, resulting in easterly winds that are more similar to the summer. These events, known as sudden stratospheric warmings (SSWs) were discovered in the early 1950s, and today they are observed in detail by satellites. We have a good dynamical understanding of how and why SSWs occur, but our understanding of how they affect both surface weather and the upper atmosphere is incomplete. We observe that variability of the stratospheric circulation (SSWs being an extreme event) are associated with shifts in the jet stream and the paths of storms, with associated effects on rainfall and temperatures. The likelihood of cold weather spells and damaging wind storms is also affected. Almost all SSWs have occurred in the Northern Hemisphere, but there was one spectacular major SSW in 2002 in the Southern Hemisphere.

1 Introduction

The wintertime stratosphere is characterized by a strong, westerly, cold polar vortex. The polar vortex is formed primarily through radiative cooling and is characterized by a band of strong westerly winds at mid- to high latitudes. Typical temperatures are ∼-55 to -65°C in the polar Northern Hemisphere at 10 hPa. Roughly every two years, the wintertime vortex is disrupted by planetary-scale waves to an extent that this circulation breaks down, with westerly winds becoming weak easterly, and temperatures climbing to ∼-30°C—essentially summertime conditions. This phenomenon happens rapidly, and is known as a sudden stratospheric warming (SSW). Figure 1 illustrates a sudden warming event in 2018/19, together with the background climatology and variability of zonal wind and the average temperature from 65°-90°N at 10 hPa. Note that both the lowest and highest recorded temperatures occurred in mid-winter. Outside of winter, the stratosphere is quiescent. The warming event (red curve) was followed, after more than a month, by anomalously low temperatures and strong winds in the middle stratosphere. Figure 2 illustrates zonal mean temperature anomalies averaged over days 0-30 following SSW events. Note that the upper stratosphere cools, and that there is slight cooling in the mid-latitudes and tropics in compensation for the downward adiabatic warming over the polar cap. Vectors illustrate the approximate motion consistent with the temperature anomalies (and pressure anomalies, not shown). See Baldwin et al. (2020) for details of the calculation. In particular, note the poleward movement of mass near the surface at high latitudes. This leads to higher Arctic surface pressure following SSWs.

The effects of SSWs last much longer in the lower stratosphere and troposphere than they do in the uppper stratosphere. Figure 3a illustrates a lag composite of temperature anomalies for SSW events in JRA-55 data (1958–2015). Above 30 km, the SSW events end within two to three weeks, while in the lowermost stratosphere SSWs last more than two months, on average. This is largely due to the faster radiative time scale in the upper stratosphere. Pressure anomaly composites (Figure 3b) are similar to temperature, except that surface effects are clearly visible. The "lumpiness" of the surface signal is due to averaging of a relatively small number of SSWs. Averaged over days 0–60 the surface pressure anomaly is 2.1 hPa, but is only 0.74 hPa near the tropopause. This is called "surface amplification". The fact that the pressure anomaly from SSWs is largest at the surface is important. It means that tropospheric near-surface processes must be reinforcing the stratospheric signal, raising surface pressure over the polar cap (See Section 7).

SSWs are fascinating from a fluid dynamical perspective, and perhaps the simplest and most insightful way of viewing the dynamics is maps of potential vorticity (PV; see Section 4) (McIntyre & Palmer, 1983, 1984). Maps of PV in the middle stratosphere show that planetary-scale wave breaking erodes the polar vortex, sharpening its edge. All SSWs are preceeded by erosion of the vortex. The wave-breaking erosion forms a "surf zone" surrounding the vortex. With fine enough resolution, one can see filamentation—thin streamers of PV being stripped away from the vortex and mixed into the surf zone. This horizontal view stands in contrast to the zonal mean, which shows mainly rapid temperature rises as air descends over the polar cap, accompanied by slowing of the zonal winds. Different mechanisms to explain the occurrence of SSWs are discussed in Section 4.

An underlying question is whether or not SSWs are dynamically unique extreme events. Given the observed distributions of temperatures, winds, PV, etc., do SSWs stand out as outliers from the distribution? Or is it that SSWs simply occupy one tail of the distribution? In the Northern Hemisphere (NH), it appears that SSWs occupy one tail of the distribution. There is a broad continuum of warmings, from very minor to major deviations from climatology (Coughlin & Gray, 2009). Thus, defining an SSW as having occurred or not comes down to defining a fixed threshold (e.g., of absolute stratospheric fields such as polar wind and/or temperature at some level) or a relative field (e.g., based on the variability of the polar stratosphere such as the Northern Annular Mode or just the variability of the polar temperature (Butler et al., 2015)). There are several criteria for detecting major SSW events as will be described in Section 3. Different criteria often identify the same major disruptive events but differ in the quantitative size and timing of the events.

In the Southern Hemisphere (SH) there has been only one major SSW, and it was indeed spectacular (Kruger et al., 2005). In terms of daily zonal wind speeds, the event was approximately eight standard deviations from the mean. As rare as this event was, in early September 2019 a similarly anomalous event occurred, though it did not technically qualify as a major SSW by established criteria (Hendon et al., 2019). Southern Hemisphere warming events are important because they inhibit strong heterogeneous ozone depletion—essentially preventing the formation of the ozone hole—and because these events affect jet streams, precipitation (and droughts) especially across Australia (e.g Thompson et al., 2005; Lim et al., 2019).

SSWs are not only important for the polar stratosphere but for the whole atmosphere too. SSWs affect the circulation in the tropical stratosphere (e.g. Kodera et al., 2011) and beyond, mixing chemical constituents such as ozone, as indicated in Section 9. The large descent over the polar cap associated with the SSW is balanced by upwelling south of ∼50°N that extends into the Southern Hemisphere (Figure 2). Also visible is ascent (cooling) in the polar upper stratosphere, that extends into the mesosphere (Körnich & Becker, 2010). SSWs can affect thermospheric chemistry, temperatures, winds, electron densities, and electric fields, across both hemispheres (Chau et al., 2012). These effects are explained in Section 8

Nevertheless, the most important impact of SSWs occurs in the troposphere as summarized in Section 7. SSWs are observed to have substantial, long-lasting effects on surface weather and climate, especially on sea-level pressure (SLP) and the Northern Annular Mode (NAM), with associated shifts in the jet streams, storm tracks, and precipitation (e.g Baldwin & Dunkerton, 2001). These effects are much larger than can be explained by dynamical theories such as PV inversion (e.g. Charlton et al., 2005) or the tropospheric response to stratospheric wave forcing. Tropospheric processes, possibly involving low-level Arctic temperature anomalies, act to amplify the stratospheric signal (Baldwin et al., 2020).

Given the relevance of SSW events on the whole atmosphere, several efforts have been made in investigating their predictability. SSWs can be predicted relatively well 10-15 days in advance (Tripathi et al., 2015; Karpechko, 2018; Domeisen, Butler, et al., 2020a). Several phenomena outside the polar stratosphere have been identified, in the observations, as possible modulators of the likelihood of SSWs. Some of them are related to the tropical stratosphere such as the Quasi-Biennial Oscillation (QBO) and Semi-Annual Oscillation (SAO) of the equatorial stratosphere. Others are related to ocean-atmosphere system such as the El Niño-Southern Oscillation (ENSO) and Madden Julian Oscillation (MJO), and some others even refer to phenomenon outside the Earth such as the 11-yr solar cycle. With multiple possible influences, and only around 40 SSWs since 1958, quantifying these relationships is challenging.

In this study, we offer a review of our understanding of the main points of SSWs. In Section 2 a brief historical background is provided and Section 3 describes the classification of these events. Dynamical theories for the occurrence of SSWs are included in Section 4, and possible external factors driving SSWs are discussed in Section 5. The predictability of SSWs is discussed in Section 6, and their effects on climate and weather are presented in Section 7. Effects above the stratosphere are described in Section 8, and chemical/tracer aspects are shown in Section 9. Finally, the outlook/conclusion is provided in Section 10.

2 Historical background

SSWs were discovered by Richard Scherhag in radiosonde temperature measurements above Berlin, Germany. Scherhag started regular radiosonde measurements from the area of the former Tempelhof airport in Berlin in January 1951. As professor and head of the recently founded Institute of Meteorology at Freie Universität Berlin he was interested in exploring the stratosphere. With the help of the U.S. allies in post-war Berlin he was able to employ a new type of American radiosonde using neoprene balloons which provided regular measurements of the stratosphere up to ∼30 km altitude (∼10 hPa). In a first publication in spring 1952, Scherhag reported an "explosive-type warming of the high stratosphere" in January 1952 and concluded that the observed warming was too strong to be explained by advection (Scherhag, 1952a). This "Berlin Phenomenon", as Scherhag called the warming, developed as follows: "While all measured stratospheric temperatures ranged between –56 and –69°C on January 26, two days later only –37°C were measured at 13 hPa. This means, a sudden warming of 30 degrees had started on January 27. On January 30, the temperature reached –23°C in 10 hPa, followed by a rapid cooling." Scherhag also found that the warming slowly propagated downward to the 200 hPa pressure level within one week. This first warming pulse was followed by a second, even stronger warming about one month later, with a temperature maximum

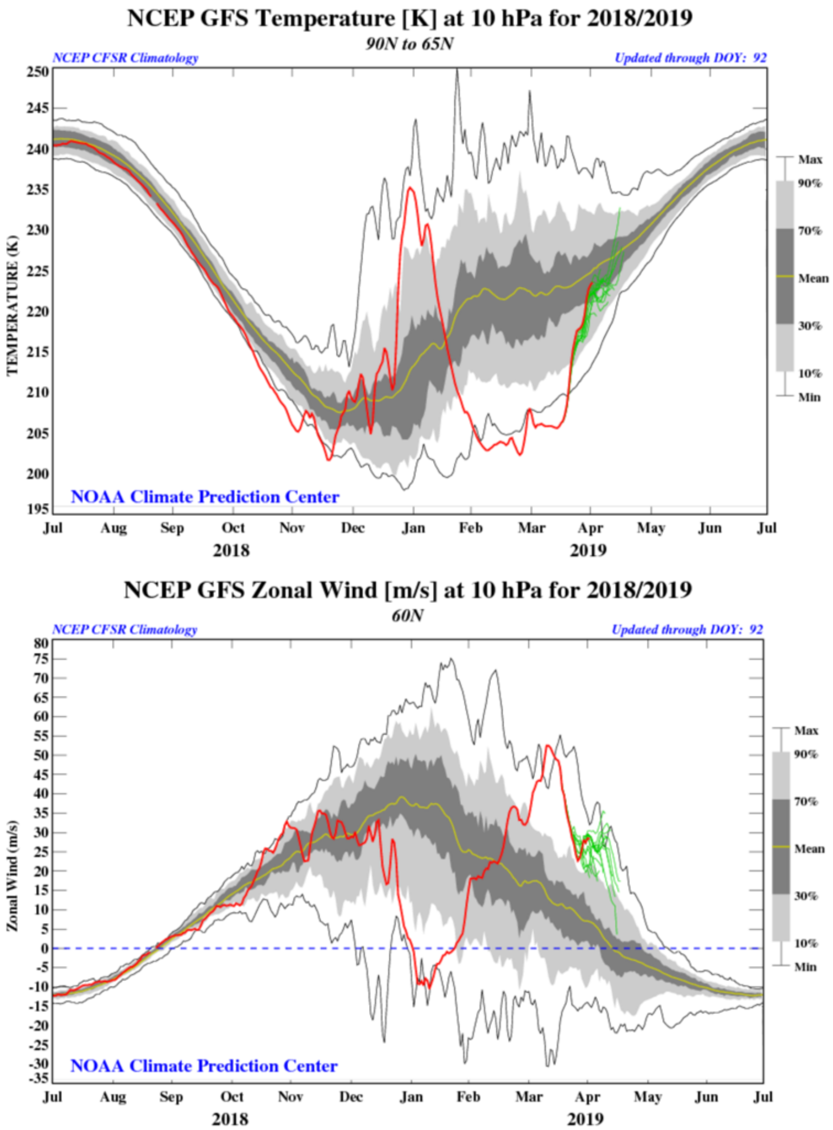


Figure 1. (top) The 10-hPa 65°–90°N observed zonal-mean temperatures and (bottom) zonal-mean wind at 60°N for 2018–19. An SSW event is seen as the upward spike in temperature (red) and the reduction to less than zero in zonal wind (easterlies). The yellow line signifies the average conditions in the stratosphere for that time of year, while the gray shadings show 70th and 90th percentiles. Solid black lines show the max/min for 1979–2019. The thin green lines are forecasts. [From Baldwin et al. (2019). Original source: NOAA/NWS/Climate Prediction Center, https://www.cpc.ncep.noaa.gov/products/ stratosphere/SSW/.]

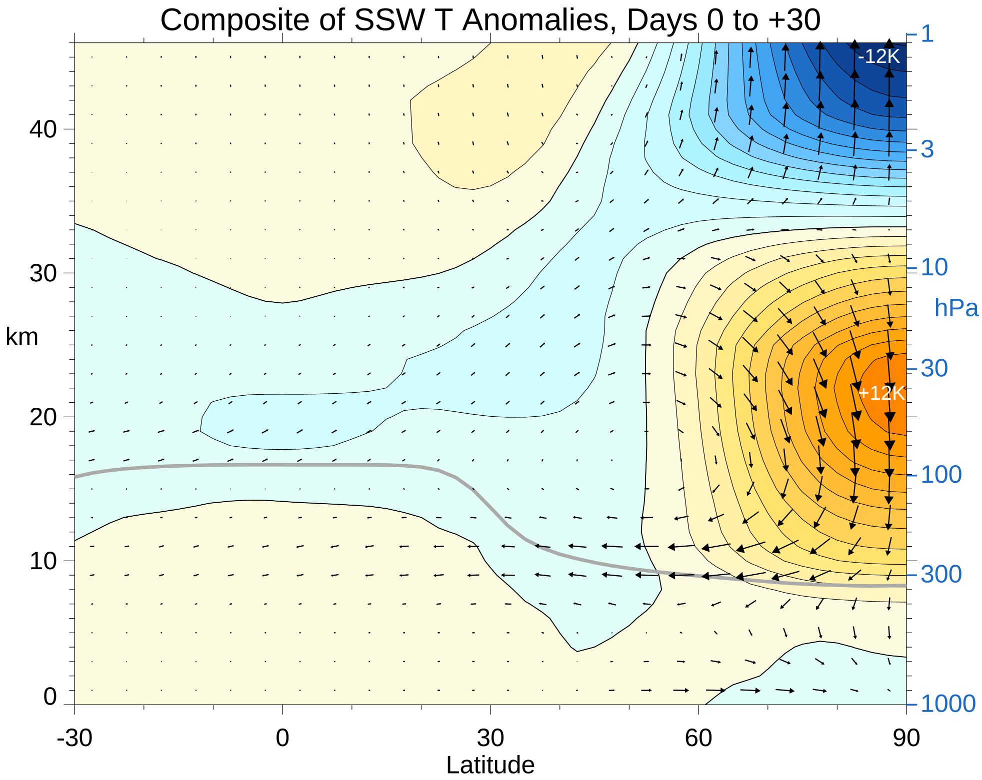
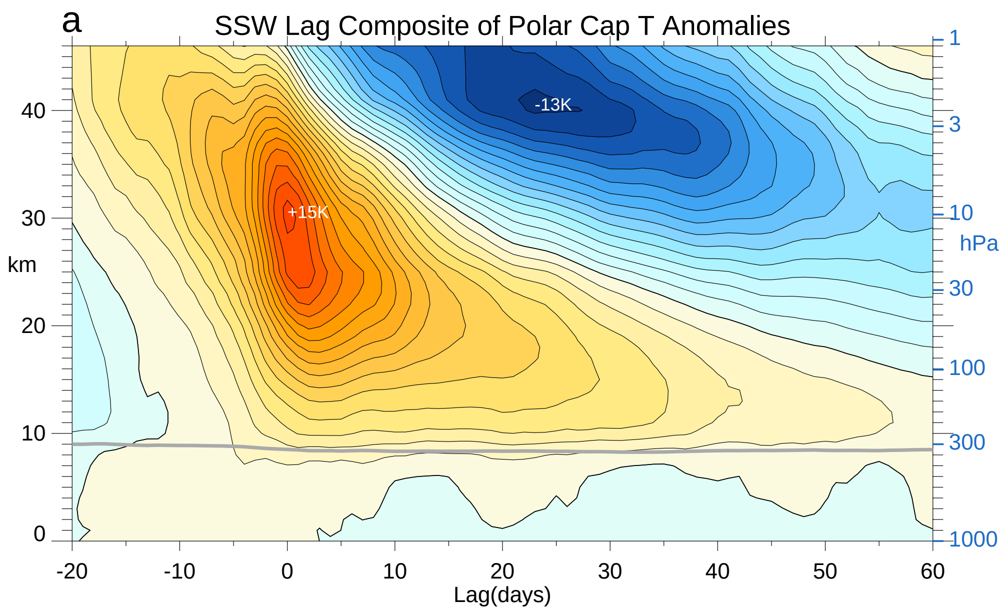


Figure 2. Composite temperature anomalies from the 0/30 days after 36 SSW events during 1958–2015 in JRA-55 data (1116 days). The SSWs dates are defined based on the reversal of the zonal mean zonal wind at 60°N and 10 hPa, applying the criterion of Charlton and Polvani (2007). The contour interval is 1K. The vectors represent the approximate movement of mass (from the climatological basic state) to reach the temperature anomalies and pressure anomalies (not shown). The calculation was performed in height coordinates (left axis). The pressure labels (right) are approximate. The lapse-rate tropopause (gray line) is shown for the days in the composite.



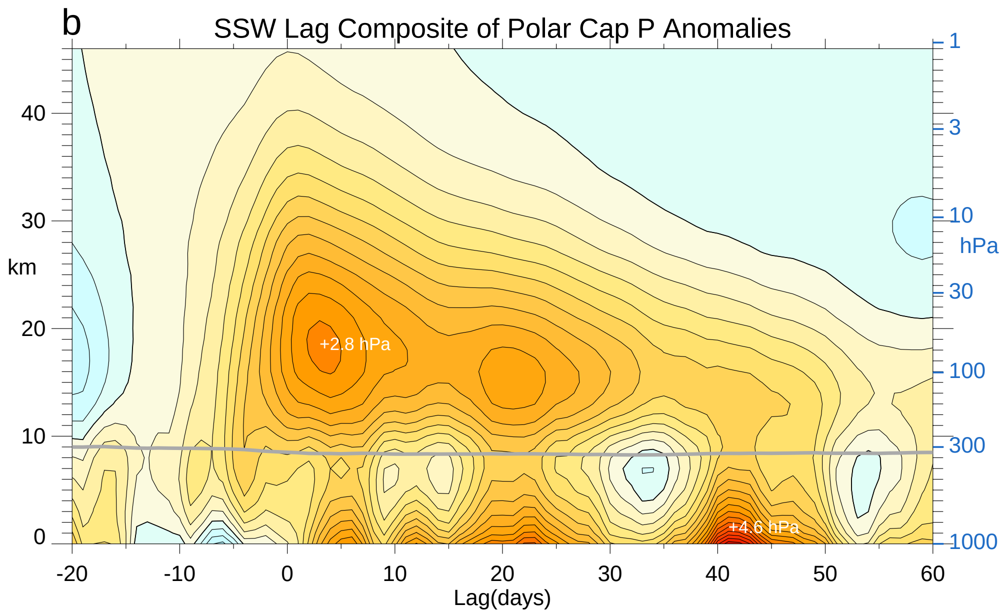


Figure 3. (a) Lag-composite polar cap (65-90°N) mean temperature anomalies from 36 SSW events during 1958–2015 in JRA-55 data. The SSWs dates are defined based on the reversal of the zonal mean zonal wind at 60°N and 10 hPa, applying the criterion of Charlton and Polvani (2007). Contour interval 1K. The tropopause (gray line) is depressed by ∼750m following the warmings. (b) as in (a) except for polar cap pressure anomalies. Contour interval 0.25 hPa

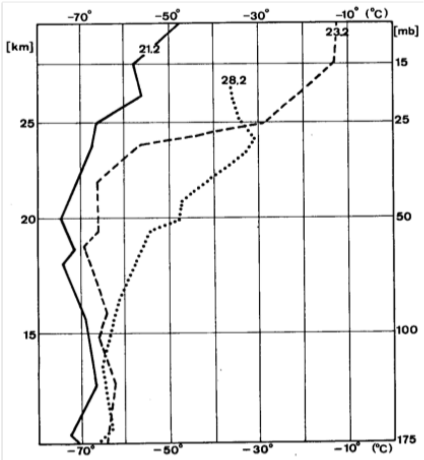


Figure 4. Radiosonde temperature measurements in Berlin-Tempelhof during the first recorded Sudden Stratospheric Warming in February 21-28, 1952. Figure from Wiehler (1955).

of −12.4°C (a warming of ∼ 37°C within 2 days) at 10 hPa on February 23 and a change in circulation to south-easterly winds in the middle stratosphere. Figure 4 shows the Tempelhof radiosonde temperature measurements of February 21 (before the warming), February 23 (at the peak of the SSW), and on February 28 (after the peak) (Wiehler, 1955). Also in February 1952, upper-level wind data from radiosondes over the northern U.S. indicated an increase of the frequency of easterly winds at 50 hPa associated with a closed persistent anticyclonic circulation northwest of Hudson Bay and a warming over Canada and Greenland (Darling, 1953).

In a first attempt to explain the unexpected warming of the winter stratosphere, Scherhag (1952b) and Willett (1952) suspected a severe solar eruption on February 24 to be the source. While we now know that solar effects are not strong enough to force individual SSWs, a statistical relation between the occurrence of SSWs and solar activity is actively discussed until present day. A similar stratospheric warming had also been noted the year before, in February 1951, from British Meteorological Office radiosonde and radar measurements over England and Scotland. It was accompanied by a reversal of the lower stratospheric winds to easterlies which were followed again by westerlies before the transition to summertime easterlies (Scrase, 1953). It then took until the winters 1956/57 and 1957/58 that similarly strong SSWs were analysed in maps which had been specifically produced on stratospheric pressure levels (Teweles, 1958; Teweles & Finger, 1958). Figure 5 shows the evolution of 50 hPa temperature over Alert, Ellesmere Land, during 3 winters with stratospheric warmings in the 1950s.

With the start of the International Geophysical Year (IGY) in July 1957, the number of radiosonde balloons reaching altitudes above 30 km increased. Regular daily or 5-daily stratospheric maps (100, 50, 30 and 10 hPa) for the Northern Hemisphere were published by several centers, e.g., the US-Weather Bureau, the Arctic Meteorology Research Group of McGill University Montreal and the Stratospheric Research Group of Freie Universität Berlin. Meteorological rocketsondes provided new insights: It was found, for example, that the strong stratospheric warming over Fort Churchill in January 1958 occurred a couple of days earlier in the altitude region above 40 km than in the layers below (Stroud et al., 1960). Moreover, intense warmings were detected in the upper stratosphere which were never detected below 10 hPa. In order to obtain an increased number of high-altitude soundings during stratospheric warmings, the STRATWARM warning system was established by the WMO in 1964. These alerts included information on

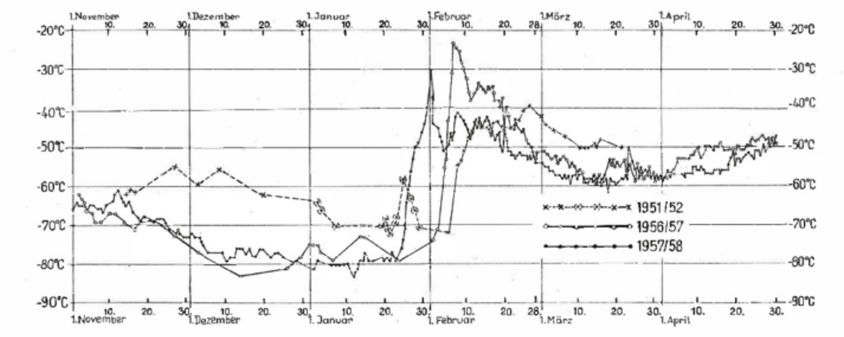


Figure 5. 50 hPa temperature time series over Alert, Ellesmere Land, during the 3 winters with stratospheric warmings in the 1950s. Figure from Warnecke (1962).

the intensity and movement of the warmings and were distributed internationally from the meteorological centers at Melbourne, Tokyo, Berlin and Washington D.C. As suggested in the WMO/IQSY (1964) report, SSWs were classified according to their time of occurrence ("mid-winter warmings" versus "final warmings" in late winter) and their intensity. While "minor mid-winter warmings" were characterized by a strong warming of the Arctic stratosphere at 10 hPa and higher levels, "major mid-winter warmings" had to be additionally accompanied by a complete circulation reversal from westerlies to easterlies at 60°N and polewards. Alternative SSW definitions that were developed later are discussed in Section 3.

In a plea for additional upper air data, Scherhag et al. (1970) raised the question of "whether an intimate knowledge of the stratospheric circulation would prove valuable in, for example, forecasting." He stated that phase relationships between events in the stratosphere and troposphere must be known for a full exploration of forecast probabilities. In fact, Scherhag had speculated about the impact of SSWs on surface weather as early as in his initial 1952 report, in which he pointed out a drop in forecast skill score following the February 1952 SSW (perhaps related to the fact that stratospheric information was not included in the forecasts). Indeed, some early studies had pointed at a potential interaction of tropospheric and stratospheric zonal wavenumber 2 during the 1967/68 warming (Johnson, 1969) and the role of tropospheric blockings for the onset of stratospheric warmings (Julian & Labitzke, 1965). An early example of stratosphere-troposphere coupling is illustrated in Figure 6 which shows 10 hPa height maps at the beginning (January 18, left panel) and peak (January 27, middle panel) of the 1963 stratospheric warming, and the surface pressure map of January 31 (Fig 6, right panel) (Scherhag, 1965). A few days after the stratospheric warming, a surface anticyclone developed over Greenland which extended through the troposphere up to the middle stratosphere. This was consistent with Labitzke (1965) who described the occurrence of a tropospheric blocking about ten days after a stratospheric warming and Quiroz (1977) who found tropospheric temperature changes after a stratospheric warming.

With the beginning of the satellite era in 1979 much improved data coverage allowed new breakthroughs in our understanding of stratospheric dynamics and SSWs. McIntyre and Palmer (1983) showed the first observationally derived hemispheric scale maps of PV on a mid-stratospheric potential temperature surface (850 K) based on the then newly available radiance data from the Stratospheric Sounding Unit (SSU). These maps clearly demonstrate the existence of large amplitude planetary wavenumber 1 preceding the 1979 SSW event with subsequent evolution showing wave breaking. The maps furthermore illustrate the split up of the vortex during the 1979 major SSW in term of PV at 850 K. Satellite data have continued to provide valuable observational constraints on the dy-

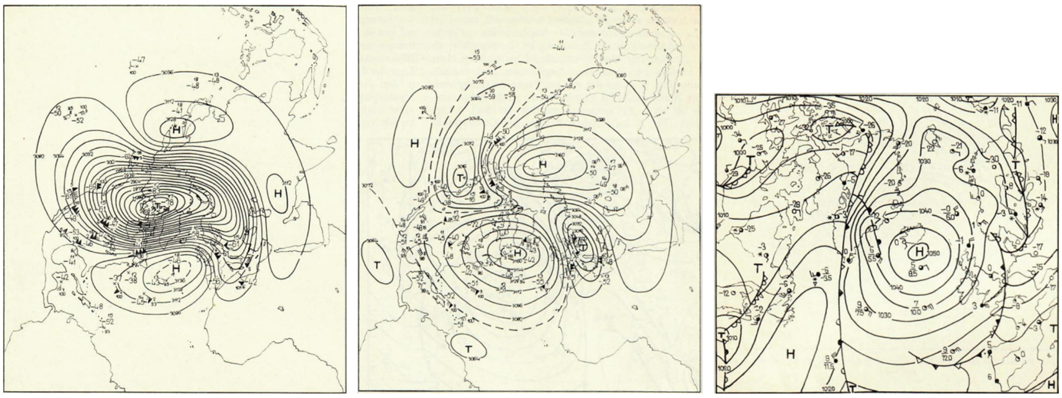


Figure 6. 10 hPa height map on January 18, 1963 (left) and January 27, 1963 (middle) and sea level pressure on January 31, 1963 (right) (From Scherhag (1965). ©Springer. Used with permission.)

namics and transport characteristics surrounding SSW events (e.g., Manney, Harwood, et al. (2009); see also Section 9).

3 Types and classification of SSWs

In the early decades after the discovery of SSWs, the WMO developed an international monitoring program called STRATALERT, led by Karin Labitzke of Freie Universität Berlin, to detect SSWs. Early metrics to measure these events were based on temperature changes, as the sudden and rapid warming of the stratosphere were key features measurable by radiosondes and rocketsondes. WMO/IQSY (1964) established that "major" SSWs were separated from more minor events by requiring a reversal (from westerly to easterly) of the zonal winds poleward of 60° latitude and an increase in the zonal-mean temperature polewards of 60° at 10 hPa (WMO/IQSY, 1964; McInturff, 1978; Labitzke, 1981). The inclusion of a zonal circulation reversal criteria stems from wave-mean flow theory which stipulates that quasi-stationary planetary-scale waves cannot propagate into easterly flow (Charney & Drazin, 1961; Matsuno, 1971; Palmer, 1981). Thus, an obvious dynamical distinction between a major and minor SSW is that vertical wave propagation from the troposphere is prohibited beyond the middle stratosphere following a major event. A remarkable aspect of these early metrics is the extent to which they still form the basis of SSW detection, despite being based on a very small number of observations.

The most commonly-used metric to detect major SSWs was proposed by Charlton and Polvani (2007) (hereafter CP07) and adapted from earlier definitions: the reversal of the daily-mean zonal-mean zonal winds from westerly to easterly at 60°N latitude and 10 hPa from November to April[1]. The earlier criteria for a temperature gradient increase was found to be largely redundant since, by thermal wind balance, this occurs in almost all cases of a zonal wind reversal. While the detection of major SSWs using the CP07 definition is sensitive to the particular latitude, altitude, and threshold of the zonal wind weakening (Butler et al., 2015), the choice of a reversal at 10 hPa and 60°N optimizes key features and impacts of major SSWs (Butler & Gerber, 2018). Having a common metric for major SSWs allows for consistent intercomparison of models (Charlton-Perez

[1] By CP07, wind reversals must be separated by 20 consecutive days of westerly winds, and must return to westerly for at least 10 consecutive days prior to 30 April, to be classified as a mid-winter SSW.

et al., 2013; Kim et al., 2017; Ayarzagüena et al., 2018) and reanalyses (Palmeiro et al., 2015; Butler et al., 2017; Martineau et al., 2018; Ayarzagüena et al., 2019).

It should be noted that the CP07 metric was developed during a time when the increased availability of global climate model simulations necessitated the evaluation of the model stratosphere in large gridded datasets (Charlton-Perez et al., 2013). Thus, a major criterion for the CP07 metric was that the data request needed for the calculation should be as small as possible. In the current era, with greater availability of dynamical metrics output from model simulations (Gerber & Manzini, 2016), this requirement is not as stringent. Thus, it is worth emphasizing the intended use of the CP07 definition as a simple metric for polar vortex weak extremes, rather than as an infallible selection of events that should be deemed "important". This metric yields on average 6 major SSWs per decade in the NH. There is however significant decadal variability in the frequency of SSW events (Reichler et al., 2012), with the 1990s exhibiting only two SSWs (in 1998 and 1999) and the 2000s exhibiting 9 events according to the CP07 metric. Recent decades show stronger decadal variability in SSW frequency than earlier decades, with the 1990s likely representing the longest absence of SSW events since 1850 (Domeisen, 2019).

The application of the CP07 metric to the SH polar vortex (where zonal-mean zonal wind reversals at 60°S and 10 hPa between May-October are considered) reveals only one major SH SSW in the reanalysis back to 1958, which occurred on 26 Sep 2002 (Shepherd et al., 2005). This highlights important differences in dynamics and climatology between the NH and SH. However, in mid-September of 2019 an extremely anomalous weakening of the SH vortex occurred (Hendon et al., 2019) that did not meet the CP07 criterion for a major SSW. Nonetheless, this event should not be disregarded simply because the circulation failed to meet one metric; significant and persistent impacts on SH surface climate followed this SSW, such as extensive Australian bushfires (Lim et al., 2019). Further diagnostics should thus be considered for evaluating the relevance of extreme vortex events in both hemispheres for surface weather effects; a so-called minor SSW can have major societal impacts.

In addition to major versus minor SSWs, there is also classification of the morphology of the event. During a SSW, the polar vortex can either be displaced off the pole or split into two sister vortices. Several different methods have been developed to classify split versus displacements (CP07; Mitchell et al., 2011; Seviour et al., 2013; Lehtonen & Karpechko, 2016). About a third of the observed 36 major SSWs in the 1958-2012 period can be unanimously classified across all methods as splits and another third as displacements (Gerber et al., 2020). The rest of the events are more ambiguous across methods, perhaps because in some cases the polar vortex both displaces and splits within a period of several days (Rao, Garfinkel, et al., 2019).

Furthermore, SSWs have been classified by the zonal wavenumber of the tropospheric precursor patterns leading up to the SSW. These predominantly wave-1 and wave-2 patterns tend to precede SSWs (Tung & Lindzen, 1979a; Woollings et al., 2010; Garfinkel et al., 2010; Cohen & Jones, 2011). In particular, blocking (a persistent anomalous high pressure) over the Pacific region and North Atlantic/Scandinavian region has been tied to wave-2 driving of split vortex events (Martius et al., 2009). Anomalous low pressure over the North Pacific/Aleutians with Euro-Atlantic blocking has been tied to wave-1 driving of primarily displacement vortex events (Castanheira & Barriopedro, 2010). Note that while displacements of the vortex are nearly always preceded by wave-1 forcing, splits of the vortex can be preceded by either wave-1 or wave-2 forcing (Bancalá et al., 2012; Barriopedro & Calvo, 2014) and often proceed with an increase in wave-1 followed by a subsequent increase in wave-2.

While the focus of this review is on SSWs, which represent the weakest polar vortex extremes, SSWs are just one extreme within a broad spectrum of polar stratospheric dynamic variability. A wide range of variations (see Figure 1, daily maximum and minimum values in black lines)– from more minor deviations from climatology, to the strongest polar vortex extremes– can influence stratosphere-troposphere coupling, transport, and chemical processes. Polar stratospheric variability peaks from January-March in the Northern Hemisphere, and from September–November in the Southern Hemisphere (though variability is less). Early winter extremes may evolve differently than late winter extremes; for example, *Canadian Warmings* are amplifications of the Aleutian High in the lower and middle NH stratosphere, and are the dominant type of stratospheric warming in early boreal winter (Labitzke, 1977). Additional metrics have been proposed to better capture the full spectrum of polar stratospheric variability. A number of studies consider metrics based on empirical orthogonal functions (EOFs). For example, the first EOF of geopotential height anomalies, also known as the "annular mode", (Baldwin & Dunkerton, 1999, 2001; Baldwin & Thompson, 2009; Gerber et al., 2010) captures mass fluctuations between the polar cap and extratropics. EOFs of vertical polar-cap temperature profiles have been used to identify weak vortex extremes (SSWs) that have the most extended recovery periods, called "Polar-night Jet Oscillations" (PJO) (Kuroda & Kodera, 2004; Hitchcock & Shepherd, 2012; Hitchcock et al., 2013). An advantage to EOF-based techniques is that thresholds for extremes are based on anomalies (deviations from the climatology) rather than absolute values, as in the CP07 zonal wind metric. Thus, EOF metrics can capture anomalous events relative to any changes in the climatology (McLandress & Shepherd, 2009a; Kim et al., 2017).

4 Development of dynamical theories

SSWs are a manifestation of strong two-way interactions between upward propagating planetary waves and the stratospheric mean flow. The polar vortex can be disrupted by large wave perturbations, primarily planetary-scale zonal wave-number 1–2 quasi-stationary waves. Sufficient wave forcing of the mean flow by these waves can result in an SSW, with the breakdown of the westerly polar vortex, and easterly winds replacing westerlies near 10 hPa, 60°N. When the winds in the polar vortex slow, air is forced to move poleward to conserve angular momentum, with descent over the polar cap (arrows in Figure 2). The adiabatic heating associated with this descent results in the observed rapid increases in polar cap temperatures on time scales of just a few days.

Strong westerly winds in the polar night jet inhibit all but the largest, planetary scale waves from propagating into the stratosphere (Charney & Drazin, 1961). While planetary scale waves can spontaneously be generated by baroclinic instability or via upscale cascade from synoptic scale waves (Scinocca & Haynes, 1998; Domeisen & Plumb, 2012), they are chiefly forced by planetary scale features at the surface: topography and land-sea contrast. The relative zonal symmetry of the austral hemisphere explains why SSWs are almost exclusively a boreal hemispheric phenomena, but this does not imply that the stratosphere just passively responds to wave driving from the troposphere.

The diversity of observed SSWs demonstrates that some SSWs appear to be forced by anomalous bursts of planetary wave activity from the troposphere, while in other SSWs the stratosphere itself acts to regulate upward wave propagation. All theories agree, however, that it is the sustained dissipation of wave activity in the stratosphere, chiefly through nonlinear wave breaking and irreversible mixing (Eliassen-Palm flux convergence), that generates a deep, sustained warming of the polar vortex. Once the vortex is destroyed, strong radiative cooling helps to rebuild the vortex provided there is time before the end of winter, but this radiatively controlled process can take several weeks (see Figure 3). Rotation and stratification couple the poleward transport of heat by waves to a downward transport of westerly momentum. Thus, the warming of the polar stratosphere occurs in concert with an eradication of the climatological vortex in a major warming event.

4.1 Wave-mean flow interactions, dissipation, and SSWs

The wintertime stratospheric polar vortex is formed primarily through radiative cooling, as absorption of UV radiation by ozone shuts off in the polar winter. Much of the theory of how SSWs occur relies on the basic assumption of waves propagating on a zonal mean flow. Although this assumption is violated during the extreme flow disruptions of SSWs (particularly at high latitudes), wave mean-flow interaction theory has been remarkably successful in explaining (at least qualitatively) the dynamics of how SSWs occur. Upward propagation of a Rossby wave on a zonal-mean flow is associated with a poleward heat flux, $\overline{v'\theta'}$ (e.g. Vallis, 2017, Chpt. 10). Warming of the vortex could then, in principle, be provided by convergence of the heat flux on the poleward flank of an upward propagating planetary wave. However, an opposing tendency arises due to the fact that the wave also induces vertical advection, producing adiabatic cooling where the heat flux would otherwise warm the air. (Likewise, the air on the equatorward side, which would be cooled by the poleward heat flux, sinks and adiabatically warms.) For conservatively propagating waves, i.e., a case with no dissipation, the two tendencies exactly cancel and no net warming or cooling occurs:

$$\overline{\omega}_r \frac{\partial \overline{\theta}}{\partial p} = -\frac{\frac{\partial}{\partial \varphi}(\cos\varphi \overline{v'\theta'})}{a\cos\varphi}\,.$$

Here, $\overline{\omega}_r$ refers to the reversible component of zonal mean vertical motion that arises due to conservatively propagating waves.

The calculus changes when the waves are allowed to dissipate, either damped by radiation and/or friction, or more cataclysmically, through non-linear breaking (though dissipation still plays a role, as breaking simply moves energy to smaller scales). Rossby waves carry easterly momentum owing to their intrinsic easterly phase speed; this easterly momentum is transferred to the mean flow during dissipation. The resulting easterly body force not only decelerates the vortex but also causes poleward flow, due to the Coriolis torque, and downwelling over the polar cap. This downwelling opposes the wave-induced upward motion described above. With extreme wave dissipation, it completely overwhelms the upwelling tendency and drives the spectacular warming of the polar stratosphere characterized by an SSW.

From this perspective, formalized in the "Transformed Eulerian Mean" representation of atmospheric dynamics (Andrews & McIntyre, 1976; Edmon et al., 1980), it is the residual downwelling that gives rise to warming of the polar cap when planetary waves dissipate. Neglecting diabatic heating during the onset of the warming, this can be written as in equation 1:

$$\frac{\partial \overline{\theta}}{\partial t} \approx -\overline{\omega}\frac{\partial \overline{\theta}}{\partial p} - \frac{\frac{\partial}{\partial \varphi}(\cos\varphi \overline{v'\theta'})}{a\cos\varphi} = -\overline{\omega}\frac{\partial \overline{\theta}}{\partial p} + \overline{\omega}_r \frac{\partial \overline{\theta}}{\partial p} = -\overline{\omega}^* \frac{\partial \overline{\theta}}{\partial p}, \tag{1}$$

where $\overline{\omega}^* \equiv \overline{\omega} + \frac{\frac{\partial}{\partial \varphi}(\cos\varphi \overline{v'\theta'})}{(a\cos\varphi)\frac{\partial \overline{\theta}}{\partial p}}$ is a modified vertical velocity that incorporates the effect of reversible wave-induced vertical motion and therefore corresponds to the net, residual vertical motion that gives rise to adiabatic warming (residual downwelling) or cooling (residual upwelling). Note that the full temperature tendency needs to also take into account diabatic (radiative) heating.

Planetary wave dissipation gives rise to polar cap warming. However, in part because the fundamental assumption of waves propagating on a zonal mean flow is violated, it falls short of explaining the explosive warming associated with SSWs. During an SSW, the vortex may be displaced from the pole or split in two, clearly violating the assumption of waves propagating on a zonal mean flow. The wave-induced deceleration of the vortex and the associated polar cap warming are at extreme levels; exactly how such extreme interactions between the waves and mean flow get triggered and unfold to the point of complete breakdown of the vortex is still not fully understood.

Two different perspectives exist in the literature regarding the role of the troposphere (see section 4.2 below). Early work focused on the role of anomalous wave fluxes from the troposphere that drive the SSW, i.e., provide sufficient additional wave drag in the stratosphere to destroy the vortex, especially if it accumulates over a sufficiently long period of time. A second view holds that, given a wave field provided by the troposphere—which does not need to be anomalously strong—the stratospheric polar vortex may spontaneously feed back onto the wave field such that both get mutually amplified, reminiscent of resonance phenomena (e.g Plumb, 1981; Albers & Birner, 2014).

Regardless of the perspective on the triggering mechanisms of SSWs, once the primary circulation breaks down and easterlies ensue, vertical propagation of stationary Rossby waves is inhibited. (Stationary wave can only exist if there are mean westerlies to offset their intrinsic easterly propagation.) The resulting "critical line" drives an accumulation of wave dissipation just below it, associated with more easterly acceleration and rapid lowering of the critical line (Matsuno, 1971). The corresponding downward progression of easterly zonal wind anomalies is mechanistically similar to the QBO (Plumb & Semeniuk, 2003), but acts on a much faster timescale, on the order of days, not years.

Another way of viewing sudden warmings is by viewing of potential vorticity (PV) on isentropic surfaces is shown in equation 2

$$PV = -g\frac{\partial\theta}{\partial p}(\zeta_\theta + f) \tag{2}$$

where g is gravity, θ is potential temperature, p is pressure, ζ_θ is relative vorticity perpendicular to an isentropic surface, and f is the Coriolis parameter. PV combines the conservation of mass and angular momentum, and PV is materially conserved in the absence of diabatic processes. Thus, it is extremely powerful as a diagnostic tool on the time scales associated with SSWs. By examining maps of PV on isentropic surfaces, it is possible to observe the breaking of planetary-scale Rossby waves in the "surf zone" (McIntyre & Palmer, 1983, 1984). SSWs can be seen to arise as a consequence of planetary-scale wave breaking, which causes the polar vortex to be eroded, and, ultimately dissipated. During early winter radiative cooling causes the vortex to strengthen. As winter progresses, wave breaking in the surf zone sharpens the edge of the vortex, and if the wave breaking persists, the vortex can be displaced from the pole or even split in two. This can be viewed on horizontal maps of PV, as seen in Figure 7, or simply by measuring the size of the polar vortex in terms of PV (e.g. Butchart & Remsberg, 1986; Baldwin & Holton, 1988).

4.2 Bottom up or top down: An evolving understanding on the mechanism(s) driving SSWs

A "bottom up" perspective, focused on the role of enhanced tropospheric wave forcing, is inherent in Matsuno's seminal work on showing that SSWs are dynamically forced. Matsuno (1971) prescribed a switch-on planetary wave 2 forcing at the lower boundary (approximately the tropopause) of a general circulation model. The model produced a strong split SSW in response to this pulse from below.

Matsuno's work suggests two key criteria for forcing an SSW. (1) SSWs only happen with sufficiently strong planetary wave forcing from the troposphere, and (2) SSWs require a pulse of anomalously strong wave forcing from the troposphere to initiate. Support for the first criterion includes the simple observation that warming events are much more prevalent in the Northern versus Southern Hemisphere. Additional support for a necessary minimum amount of wave forcing from the troposphere was established in a conceptual model developed by Holton and Mass (1976), who sought to distill an SSW down to its most basic elements.

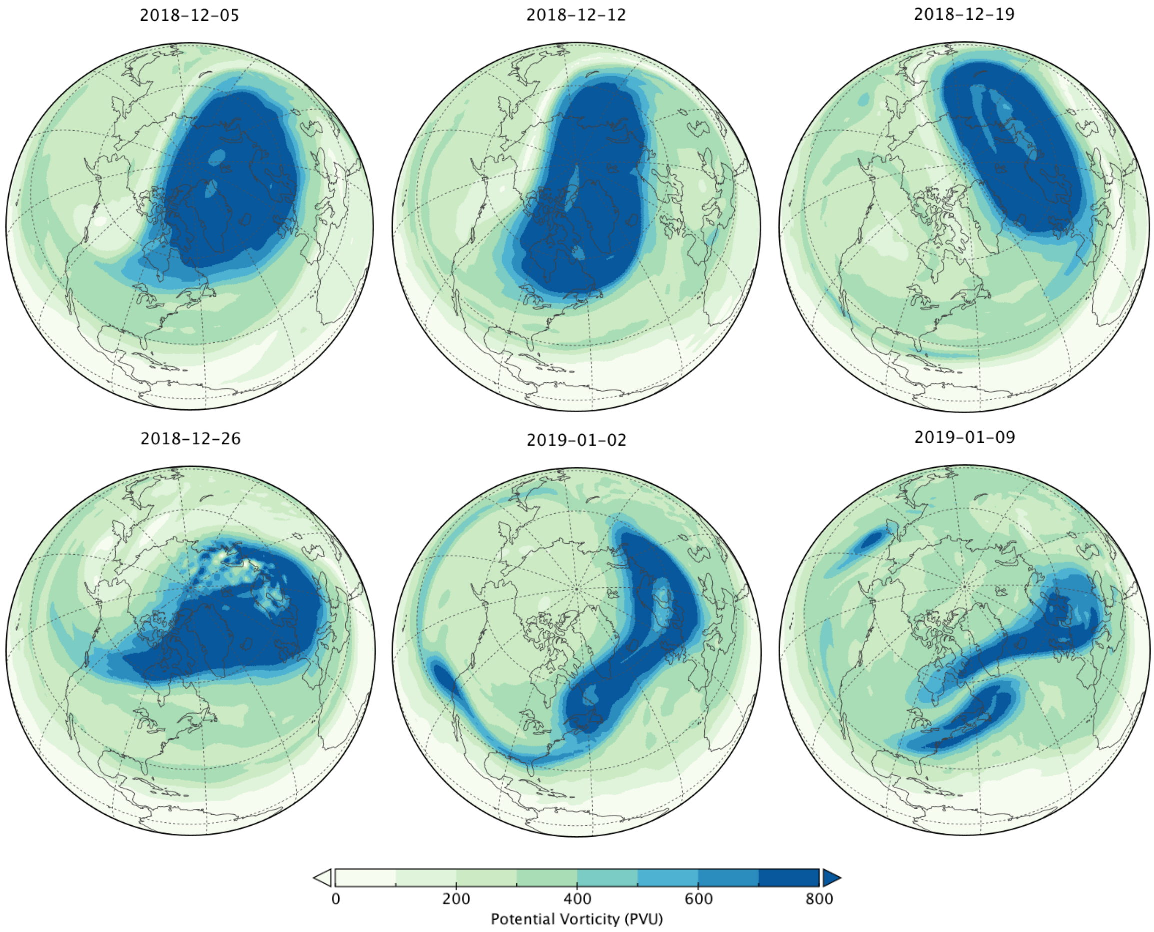


Figure 7. Illustration of the evolution of the polar vortex during an SSW in the winter 2018/19. Panels show PV on the 850 K isentropic surface on six dates, showing a sequence illustrating a displacement of the vortex off the pole with concomitant stripping away of vortex filaments into the surf zone. Once the vortex is fully displaced off the pole (bottom middle) it then further splits into two small daughter vortices (bottom right). From Baldwin et al. (2019) ©American Meteorological Society. Used with permission.

The Holton and Mass model consists of a single planetary wave of constant amplitude, prescribed as input forcing to the stratosphere at its lower boundary. The mean flow (i.e, the vortex) exists either in a strong state with weak wave amplitudes (corresponding to weak wave-mean flow interaction), or a weak state with strong wave amplitudes (corresponding to strong wave-mean flow interaction similar to the dynamics involved in SSWs). More recently, idealized GCM studies have found a sharp increase in SSW frequency as planetary scale zonal asymmetries in the underlying flow are increased, either by topography (e.g., Taguchi and Yoden (2002); Gerber and Polvani (2009)) or thermal perturbations (Lindgren et al., 2018).

The second criterion in the Matsuno model—that SSWs are driven by an exceptional pulse of wave activity from the troposphere—is supported by the fact that SSWs are often preceded by blocking events, which amplify the tropospheric wave activity (e.g. Quiroz, 1986; Martius et al., 2009). This has led researchers to look for tropospheric precursor events that potentially give rise to additional planetary wave fluxes entering the stratosphere (e.g. Garfinkel et al., 2010; Cohen & Jones, 2011; Sun et al., 2012).

Palmer (1981) suggested that the stratospheric vortex may need to be "pre-conditioned" to accept a pulse of wave activity, based on observations of the 1979 event, a topic further explored by McIntyre (1982). Various studies have suggested that the strength and size of the vortex play a critical role in allowing wave activity to penetrate deep into the stratosphere (Limpasuvan et al., 2004; Nishii et al., 2009; Kuttippurath & Nikulin, 2012; Albers & Birner, 2014; Jucker & Reichler, 2018).

Newman et al. (2001) and Polvani and Waugh (2004) pointed out that a single precursor event will likely not cause sufficient deceleration of the stratospheric polar vortex; rather it is the accumulated wave forcing over 40-60 days that needs to be anomalously strong to cause enough deceleration to reverse the zonal mean flow around the polar cap. Sjoberg and Birner (2012) further pointed out that sustained forcing that lasts for at least 10 days, but does not need to be anomalously strong, is crucial for forcing SSWs. Processes that can lead to such a sustained increase in wave forcing from the troposphere are discussed in Section 5.

Preconditioning suggests that the state of the stratospheric vortex impacts its receptivity to accept waves from the troposphere. The "top down" perspective takes this view to the extreme, supposing that that fluctuations in tropospheric wave forcing do not play an important role at all. Rather, as long as the background wave fluxes entering the stratosphere are strong enough (such as provided by the climatological conditions in Northern Hemisphere winter) the stratosphere is capable of generating SSWs on its own.

The top down perspective has often been framed in the context of resonant growth of wave disturbances (e.g. Clark, 1974; Tung & Lindzen, 1979b). In a particularly insightful incarnation of this mechanism, the wave-mean flow interaction causes the vortex to tune itself toward its resonant excitation point (Plumb, 1981; Matthewman & Esler, 2011; Scott, 2016). Support for this perspective comes from idealized numerical model experiments that show that the stratosphere is capable of controlling the upward wave activity flux near the tropopause (Scott & Polvani, 2004, 2006; Hitchcock & Haynes, 2016) and that stratospheric perturbations can trigger SSWs even when the tropospheric wave activity is held fixed (Sjoberg & Birner, 2014; de la Cámara et al., 2017).

Preconditioning of the polar vortex, i.e., wave driving that brings it to the critical state, would clearly play a key role in this mechanism, suggesting that SSWs could potentially be predicted in advance, even in the limit where they are entirely controlled by the state of the stratospheric vortex.

The bottom up and top down SSW mechanisms are associated with a different expected lag-lead relationship in upward wave energy propagation (i.e., the EP-flux) between the tropospheric source and stratospheric sink. Events forced by tropospheric waves will be preceded by a build up of wave activity over time, while self-tuned resonant SSWs would be characterised by nearly instantaneous wave amplification throughout an extended deep layer, and no lag between troposphere/tropopause and stratosphere.

In this context it is important to note that fluctuations in the upward wave flux at 100 hPa are not generally representative of fluctuations in the troposphere below (Polvani & Waugh, 2004; Jucker, 2016; de la Cámara et al., 2017). The typical tropopause pressure over the extatropical atmosphere during winter is around 300 hPa, as shown in Figures 2 and 3. That is, wave flux events at 100 hPa can generally not be interpreted as tropospheric precursor signals because $\sim$2/3 of stratospheric mass is below 100 hPa. Nevertheless, enhancements of upward wave fluxes from the troposphere at sufficiently long time scales (e.g., associated with climate variability extending over the whole winter season) tend to cause enhanced wave flux across 100 hPa into the polar vortex, which increases the likelihood for SSWs.

Evidence supporting both the bottom up and top down pathways has been observed, but it has become clear that the second criterion suggested by the Matsuno (1971) model—that the troposphere must drive an SSW with a pulse of enhanced wave activity—is not necessary. Birner and Albers (2017) found that only 1/3 of SSWs can be traced back to a pulse of extreme tropospheric wave fluxes. Roughly 2/3 of observed SSWs are more consistent with the top-down category or do not fit into either prototype (i.e. tropospheric wave fluxes are anomalously strong but not extreme). Similar ratios have been observed in modeling studies by White et al. (2019) and de la Cámara et al. (2019).

It also appears that mechanism may vary with the type of warming. While Matsuno (1971) prescribed a wave 2 disturbance, it appears that wave 1 (displacement) events tend to be associated with the slow build up of wave activity, better matching the bottom-up paradigm, although resonant behavior has also been suggested for displacement events (Esler & Matthewman, 2011). Split, or wave 2, events are more instantaneous in nature (Albers & Birner, 2014; Watt-Meyer & Kushner, 2015), more closely matching the top-down paradigm.

5 External influences on SSWs

Because there have only been around 40 observed SSWs between 1958 and 2019, it is challenging to quantify and/or establish statistically robust changes in frequency of SSWs from external influences, especially if the observations show a subtle effect. Despite this difficulty, a range of external influences have been connected to SSWs, including the Quasi-Biennial Oscillation (QBO), ENSO, 11-year solar cycle, the Madden-Julian Oscillation, and snow cover. Confidence in the robustness of such relationships is increased if there is a well described physical mechanism that is expected to produce the observed effect, for example through changes in the propagation and breaking of Rossby waves in the stratosphere or the generation of planetary Rossby waves in the troposphere. Similarly, confirmation of observed relationships in modelling studies also increases confidence that they are robust. Even more challenging is establishing relationships in the observations whereby two or more external influences act in concert (Salminen et al., 2020).

It has been recognized for 40 years that the stratospheric polar vortex is weaker during the easterly QBO winter than during the westerly QBO winter, known as the Holton-Tan relationship (Holton & Tan, 1980; Anstey & Shepherd, 2014). The frequency of occurrence of SSW during each QBO phase is shown in Table 1 based on NCEP-NCAR reanalysis. The SSW occurence is more likely during easterly QBO winters than during westerly QBO phase (0.9/yr vs 0.5yr). Therefore, SSW events occur less frequently during the westerly phase of the QBO, consistent with early studies (Labitzke, 1982; Naito et al., 2003). Models also simulate a weakened vortex and more SSWs during easterly

QBO as compared to westerly QBO, though the magnitude of the effect tends to be somewhat weaker than that observed (e.g. Anstey and Shepherd (2014); Garfinkel et al. (2018)). At least four different mechanisms have been proposed linking the QBO to vortex variability, and the relative importance of these mechanisms is still unclear (Holton & Tan, 1980; Garfinkel, Shaw, et al., 2012; Watson & Gray, 2014; White et al., 2015; Silverman et al., 2018).

Table 1. Revisiting the QBO-SSW relationship during 1958–2019, based on the dates computed by Charlton and Polvani (2007) for 1958–2001 and by Rao, Ren, et al. (2019) for 2002–2018 with NCEP/NCAR reanalysis data. The first column is the QBO phase, the second column is the corresponding composite size total winter (Nov–Feb mean) size, the third column is the number of SSWs events for that composite size, and the forth column is the SSW frequency (units: events times per year). EQBO=easterly phase of QBO; WQBO=westerly phase of QBO. The unit of QBO50 is m s^{-1}. Reprinted with permission from Rao, Garfinkel, et al. (2019)

QBO-SSW relationship

QBO phase	Winter no.	SSW no.	SSW frequency
EQBO (QBO50≥5)	20	18	0.9
WQBO (QBO50≤-5)	36	18	0.5
Neutral (|QBO50| <5)	6	1	0.17
Total	62	37	0.60

The relationship between the northern winter stratospheric polar vortex and ENSO, including a full discussion of possible mechanisms, has recently been reviewed in this journal (Domeisen et al., 2019). The statistical relationship between ENSO and SSWs in NCEP-NCAR reanalysis data is revisited and shown in Table 2. The likelihood of SSW events increases in both El Niño and La Niña relative to the ENSO neutral state (Butler & Polvani, 2011; Garfinkel, Butler, et al., 2012). However, increases in SSW frequency during La Niña in the observed record are not thought to be forced and, instead, are associated with internal variability or confounding climate forcings (Weinberger et al., 2019; Domeisen et al., 2019), particularly in the case of weak La Niña events (Iza et al., 2016). High-top models show a response to opposite phases of ENSO that, if anything, is generally stronger than that observed (Taguchi & Hartmann, 2006; Garfinkel, Butler, et al., 2012; Garfinkel et al., 2019) and that can be used for improving predictability over Europe (Domeisen et al., 2015).

Table 2. As in Table 1 but for the ENSO-SSW relationship during 1958–2019. The unit of Niño34 is °C. Reprinted with permission from Rao, Garfinkel, et al. (2019)

ENSO-SSW relationship

ENSO phase	Winter no.	SSW no.	SSW frequency
El Ni*ñ*o (Niño34≥0.5)	20	13	0.65
moderate El Niño (0.5≤Niño34≤2)	17	13	0.77
La Niña (Niño34≤-0.5)	23	15	0.65
Neutral (|Niño34| <0.5)	19	9	0.47
Total	62	37	0.60

The solar cycle may affect the stratospheric polar vortex, and earlier work reported that mid-winter SSWs tend to occur during solar minimum QBO easterly phase (i.e. classical Holton-Tan effect) *and* during solar maximum and QBO westerly phase (Labitzke, 1987; Gray et al., 2004; Labitzke et al., 2006; Gray et al., 2010). Updating these relationship for data through 2019, however, suggests that this relationship holds, but is modest. During solar maximum/westerly QBO years, SSW frequency is 0.44/yr (Table 3). During solar minimum/easterly QBO years the frequency of SSW is increased somewhat (0.67/yr). Observations alone are not sufficient to verify that a solar-QBO-SSW relationship is robust. There is a wide spread in the ability of models to simulate an influence of solar variability on the polar stratosphere (Mitchell et al., 2015), partly related to their ability to capture the effects of solar variability on the tropical stratosphere.

Table 3. As in Table 1 but for the solar-SSW relationship during 1958–2019. Max=solar maximum; Min=solar minimum. The number in parentheses is statistics for midwinter (January–February, JF) SSW events. Reprinted with permission from Rao, Garfinkel, et al. (2019)

solar-SSW relationship

solar phase	QBO phase	Winter no.	SSW no. (JF SSW no.)	SSW frequency
Max	EQBO	11	11 (6)	1.0 (0.55)
	WQBO	16	8 (7)	0.5 (0.44)
	Neutral	3	1 (0)	0.33 (0.0)
	SUM	30	20 (13)	0.67 (0.43)
Min	EQBO	9	7 (6)	0.78 (0.67)
	WQBO	20	10 (6)	0.5 (0.3)
	Neutral	3	0.0 (0.0)	0.0 (0.0)
	SUM	32	17 (12)	0.53 (0.38)
Total		62	37(25)	0.60 (0.40)

October Eurasian snow cover has also been linked to subsequent variability of the stratospheric vortex, with more extensive snow leading to a weakened vortex (Cohen et al., 2007; Henderson et al., 2018) via a strengthened Ural ridge and subsequent constructive interference with climatological stationary waves (Garfinkel et al., 2010; Cohen et al., 2014). There is a slight increase in SSW frequency for winters following enhanced snow cover (Table 4), but this effect is not statistically significant. Results are similar if only early winter SSW events are considered (not shown). Free-running models tend to not capture the link between snow cover and a weakened vortex (Furtado et al., 2015), though models forced with idealized snow perturbations do capture this effect to some extent (Henderson et al., 2018).

Table 4. As in Table 1 but for the snow cover-SSW relationship during 1968–2019. Snow cover data is sourced from `https://climate.rutgers.edu/snowcover/table_area.php?ui_set=1`, with "enhanced" and "reduced" defined as snow cover anomalies exceeding 0.5 standard deviations. Note that snow data is missing for October 1969.

snow-SSW relationship

Snow-coverage	Winter no.	SSW no.	SSW frequency
enhanced	14	9	0.64
reduced	17	9	0.53
Neutral	20	12	0.6
No data	1	1	–
Total	52	31	0.59

The Madden Julian Oscillation (MJO) has also been shown to influence the timing of SSW events: of the 23 events considered by Schwartz and Garfinkel (2017) and the two events since, more than half (13 of 25) were preceded by MJO phases with enhanced convection in the tropical West Pacific (6 or 7 as characterized by Wheeler and Hendon (2004)). The climatological occurrence of these phases is $\sim 18\%$ (updated from Schwartz and Garfinkel (2017)), and hence this represents an increased probability of an SSW occurring. The mechanism whereby convection in the West Pacific weakens the vortex is similar to the mechanism for the influence of ENSO and snow cover: the transient extratropical response associated with the MJO constructively interferes with the climatological planetary wave pattern (Garfinkel et al., 2014). Models simulate an effect similar to that observed (Garfinkel et al., 2014; Kang & Tziperman, 2017), and SSW probabilistic predictability is enhanced when the MJO is strong (Garfinkel & Schwartz, 2017).

6 How well can SSWs be forecast?

Typically, individual SSW events are well forecast out to approximately one–two weeks. As reviewed by Tripathi et al. (2015), models are typically able to capture the onset of SSW events at least five days before the event and sometimes on much longer, sub-seasonal timescales (two weeks to two months) (Rao, Garfinkel, et al., 2019). There is, however, significant event to event variability in predictability for the same modeling systems as demonstrated for ECMWF forecasts by Karpechko (2018). Much of this variation in predictive skill is likely linked to the limitations in predictive skill of key tropospheric drivers of the SSW process. An interesting recent example of this is the limited skill that models had in forecasting the February 2018 SSW which has been linked to the inability of some models to capture high pressure over the Urals (Karpechko et al., 2018) and related anticyclonic wave breaking in the North Atlantic sector (Lee et al., 2019).

There can also be substantial variation in forecasting skill for different modeling systems, both in forecasting individual SSW events (Tripathi et al., 2016; Taguchi, 2018; Rao, Garfinkel, et al., 2019; Taguchi, 2020) and in the mean aggregate skill (Domeisen, Butler, et al., 2020a). High-top models are generally able to predict SSWs at least five days in advance, while this skill decreases to less than 50% of ensemble members predicting the SSW date at lead times of two weeks (Domeisen, Butler, et al., 2020a), though individual events can exhibit longer predictability. The impact of long standing stratospheric biases and how these influence the skill of different modeling systems, for example cold biases in the middle world stratosphere, remains an area of active research interest. As noted by Noguchi et al. (2016) predictions of SSW events are also sensitive to the background stratospheric state prior to the SSW.

Nonetheless, our ability to predict SSW events into the medium-range (lead times of three to ten days) and sub-seasonal timescales and to capture changes to the seasonal likelihood of SSW events has increased substantially in the past decade (e.g. Marshall and Scaife (2010)) as forecasting systems have increased their model top, stratospheric vertical resolution and increased the sophistication of key stratospheric physical processes like gravity wave drag. Remaining challenges include resolving the difference in forecast skill between vortex displacement and vortex splitting SSWs (e.g. Taguchi, 2016; Domeisen, Butler, et al., 2020a).

7 Effects on weather and climate

7.1 Dynamical theories for downward influence

There are several theoretical reasons to expect that SSWs (and stratospheric variability in general) should affect surface weather. The main categories of mechanisms are:

1. The remote effects of wave driving (EP flux divergence) in the stratosphere (Song & Robinson, 2004; Thompson et al., 2006). The downward effect through the induced meridional circulation has been termed "downward control" (Haynes et al., 1991).
2. Planetary wave absorption and reflection (Perlwitz & Harnik, 2003; Shaw et al., 2010; Kodera et al., 2016)
3. Direct effects on baroclinicity and baroclinic eddies (Smy & Scott, 2009).
4. The remote effects of stratospheric PV anomalies (Hartley et al., 1998; Black, 2002; Ambaum & Hoskins, 2002). This category includes studies such as White et al. (2020), in which deep polar temperature anomalies are prescribed, because they are equivalent to PV anomalies (Baldwin et al., 2020).

All of these mechanisms may contribute in some way to tropospheric effects from SSWs. If we are trying to explain the surface pressure anomalies following SSWs (Figure 3), or shifts in the NAM index (e.g. Baldwin & Dunkerton, 2001), it is clear that the main observed feature is that surface effects are roughly proportional to the anomalous strength of the polar vortex in the lower stratosphere (as measured by temperature, wind, or the NAM index). In a model study, White et al. (2020) found a robust linear relationship between the strength of the lower-stratospheric warming and the tropospheric response, with the linearity also extending to sudden stratospheric cooling events. A second observation is that surface pressure anomalies are largest near the North Pole. A mechanism based on EP flux divergence cannot explain the timing of the tropospheric response, since anomalous EP flux divergence changes sign as the SSW develops. Also, Thompson et al. (2006) found that surface effects were too small, and there was no indication of a NAM-like pressure pattern. Planetary wave absorption and reflection primarily affects tropospheric wave fields, and is not generally proportional to the anomalous strength of the stratospheric polar vortex. Direct effects on baroclinic eddies would be proportional to the anomalous strength of the stratospheric polar vortex, but the effects should be felt mainly in mid-latitudes.

The remote effects of stratospheric PV anomalies would be expected to look similar to the NAM pressure pattern, and the effects are proportional to the anomalous strength of the stratospheric polar vortex (Black, 2002). However, as pointed out by Ambaum and Hoskins (2002), the remote effects of stratospheric PV anomalies are expected theoretically to decrease through the troposphere with an e-folding depth of $\sim$5 km. PV theory explains very well the atmospheric response down to the tropopause, but it does not explain the enhanced surface pressure response in Figure 3b. Surface pressure anomalies should be only $\sim$20% of those at the tropopause. The surface pressure response is an order of magnitude larger than PV theory indicates. This "surface amplification" is well reproduced in prediction models (Domeisen, Butler, et al., 2020b).

The remote effects of stratospheric PV anomalies, combined with a mechanism to amplify the surface pressure signal, could explain the main observed SSW effects. It is clear from the observations that following an SSW, tropospheric processes act to move mass into the polar cap, raising Arctic surface pressure. The low-level build-up of mass over the polar cap cannot come from the stratosphere because the surface pressure anomalies are larger than seen at any stratospheric level. The mechanisms for this movement of mass have not been fully explained. Both synoptic-scale and planetary-scale waves are found to contribute to the tropospheric response following SSW events (Simpson et al., 2009; Domeisen et al., 2013; Garfinkel et al., 2013; Hitchcock & Simpson, 2014, 2016; K. L. Smith & Scott, 2016). Baldwin et al. (2020) hypothesized that the low-level polar cap temperature anomalies (as seen in Figures 3 and 8) are responsible for the movement of mass through the mechanism of radiative cooling-induced anticyclogenesis ((Wexler, 1937; Curry, 1987), also see modeling results in Hoskins et al. (1985)). If the Arctic lower troposphere cools, the air mass contracts and pulls in additional mass from lower latitudes, raising the average surface pressure over the Arctic, as is observed.

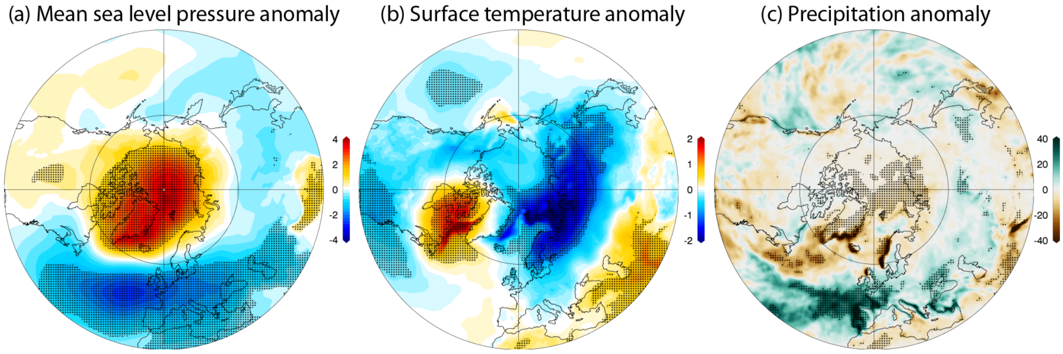


Figure 8. Composites of the 60 days following historical SSWs in the JRA-55 reanalysis for (a) mean sea level pressure anomalies (hPa), (b) surface temperature anomalies (K), and (c) precipitation anomalies (mm). Stippling indicates regions significantly different from climatology at the 95% level. [Figure from Butler et al. (2017), ©Copernicus. Used with permission..]

7.2 Observed and modeled downward impact for both hemispheres

Both hemispheres show a significant tropospheric effects following stratospheric extreme events. In particular, SSW events tend to be followed by a negative signature of the NAM in the NH and the SAM in the SH (Baldwin & Dunkerton, 1999, 2001). In the NH, the strongest response to SSW events is observed in the North Atlantic basin (Figure 8), where the response to SSW events often projects onto the negative phase of the North Atlantic Oscillation (NAO) (Charlton-Perez et al., 2018; Domeisen, 2019). The negative phase of the NAO is associated with cold air outbreaks (Kolstad et al., 2010; Lehtonen & Karpechko, 2016; King et al., 2019) over Northern Eurasia and the eastern United States, and warm and wet anomalies over Southern Europe (Ayarzagüena et al., 2018) due to the southward shift of the storm track. There are also anomalously warm temperatures over Greenland and eastern Canada, and subtropical Africa and the Middle East. Anomalous tropospheric blocking is often observed after SSW events (Labitzke, 1965; Vial et al., 2013).

In the SH, the winter stratospheric variability is weaker compared to the NH due to less wave driving (Plumb, 1989), meaning far fewer SSWs have been observed [Section 3]. Nonetheless anomalous weakenings of the SH polar vortex, tied to shifts in the seasonal evolution of the vortex, are associated with a negative SAM pattern and significant surface impacts over Antarctica, Australia, New Zealand, and South America (Lim et al., 2018, 2019). Following the only major SSW that occurred in September 2002, the SAM stayed persistently negative from September to November (Thompson et al., 2005), with warmer and drier conditions over southeast Australia, and colder and wetter conditions over New Zealand and southern Chile (Gillett et al., 2006). Similar impacts were seen following the extreme polar vortex weakening in 2019 (Hendon et al., 2019).

Furthermore, there are significant effects of SSW events in the tropics, which contribute to a downward pathway to the troposphere through tropical convective activity. In particular, the induced meridional circulation associated with the anomalous wave driving leads to anomalous tropical upwelling and anomalous cooling in the tropical tropopause region (visible in Figure 2), modulating tropical convection (Kodera, 2006). The anomalous tropical upwelling may also lead to drying of the tropical tropopause layer (Eguchi & Kodera, 2010; Evan et al., 2015).

The downward response to SSWs tends to be well reproduced in model studies. Models ranging from idealized dynamical cores to complex coupled model systems show a tropospheric response, though its persistence is often overestimated, especially in simplified models (Gerber, Polvani, & Ancukiewicz, 2008; Gerber, Voronin, & Polvani, 2008). Additionally, idealized model experiments confirm the direction of causality, i.e., stratospheric anomalies have a downward impact even if the troposphere is perturbed and does not retain memory from potential tropospheric precursors (Gerber et al., 2009). This stratospheric downward effect is known to contribute to surface predictability (Sigmond et al., 2013; Scaife et al., 2016; Domeisen, Butler, et al., 2020b).

While on average the “downward impact” of SSWs is robust, not all SSWs appear to couple down to the surface. Most studies agree that about two thirds (Charlton-Perez et al., 2018; Domeisen, 2019; White et al., 2019) of SSW events are characterized as having a visible downward impact (e.g., persistent negative phase of the NAM or NAO in the lower troposphere and/or the lower stratosphere, (e.g Karpechko et al., 2017; Domeisen, 2019)). One factor affecting the appearance of downward impact is the tropospheric NAM index prior to and at the time of the SSW. If the NAM is already negative, there will be a vertical connection to the negative stratospheric NAM. On the other hand, if the tropospheric NAM is strongly positive prior to the SSW, the appearance of vertical coupling is less likely, at least initially. The same is true for the NAO: if a negative NAO is present at the time of the SSW, the downward coupling is instantaneous but short-lived, while otherwise the negative NAO often appears after the SSW event (Domeisen, Grams, & Papritz, 2020). Because the stratosphere is one of several factors influencing the NAM, the important thing is that the effect is seen in composites of many SSWs; it cannot be expected to be seen during every SSW. The concept of surface amplification of the polar pressure signal (Figure 3) is not well understood, so it is not understood when and if the surface pressure signal will be amplified. It is also unclear whether the stratosphere always has an effect compared to what would have happened without stratospheric influence.

However, it is still not possible to predict which SSW events will have a downward impact. Knowing in advance or at the time of its occurrence if a stratospheric event will have a downward impact could have a significant benefit for medium-range to sub-seasonal predictions. Several studies have investigated possible stratospheric causes for the different surface impacts of SSW events:

1. The type of wave propagation during SSW events has been characterized as either absorbing or reflecting (Kodera et al., 2016) based on wave propagation during the recovery phase of the polar vortex, leading to different surface impacts. Absorbing type events are found to induce the canonical negative NAO response, while reflecting events are associated with wave reflection and blocking in the Pacific basin.
2. The type of SSW in terms of split or displacement had been suggested to produce different surface responses (Mitchell et al., 2013), however no significant difference in the annular mode response can be identified in long model simulations (Maycock & Hitchcock, 2015; White et al., 2019).
3. The duration and strength of the signal in the lower stratosphere has been suggested to contribute to the duration and strength of the surface impact (Karpechko et al., 2017; Runde et al., 2016; Rao et al., 2020). In particular, weak vortex events that are classified as PJO events have a stronger and more persistent coupling to the troposphere than those events that lack PJO characteristics (Hitchcock et al., 2013).

Further studies have investigated tropospheric sources for different responses to stratospheric forcing, in terms of jet stream location (Garfinkel et al., 2013; Chan & Plumb, 2009), North Atlantic weather regimes (Domeisen, Grams, & Papritz, 2020), Eastern Pacific precursors (Afargan Gerstman & Domeisen, 2020), and the characteristics of tropospheric precursors to SSW events, in particular Ural blocking (White et al., 2019). The response is also likely dependent on concurrent tropospheric climate patterns such as ENSO (Polvani et al., 2017; Oehrlein et al., 2019) and the MJO (Schwartz & Garfinkel, 2017; Green & Furtado, 2019).

8 Effects on the atmosphere above the stratosphere

The effects of SSW events are now recognized to extend well above the stratosphere, and can significantly alter the chemistry and dynamics of the mesosphere, thermosphere, and ionosphere. They are thus a significant component of the short-term variability in the upper atmosphere. This section briefly reviews the major impacts of SSWs on the upper stratosphere-mesosphere, thermosphere, and ionosphere. More detailed reviews focused solely on the upper atmosphere can be found in Chandran et al. (2014) and Chau et al. (2012).

8.1 Impacts on the Upper Stratosphere-Mesosphere

The stratopause often reforms at high altitudes ($\sim$70-80 km) after SSW events, followed by a gradual descent to its climatological altitude of $\sim$50–55 km over the ensuing 2-3 weeks (Manney et al., 2008; Siskind et al., 2010). Such elevated stratopause events occur in roughly one-third of Northern Hemisphere winters (Chandran et al., 2013, 2014). Numerical simulations successfully reproduce elevated stratopause events, providing insight into the formation mechanisms. The elevated stratopause forms due to enhanced westward gravity wave forcing following SSW events, which leads to downwelling and adiabatic heating at high altitudes where the stratopause reforms (Chandran et al., 2013; Limpasuvan et al., 2016).

SSW events lead to dramatic changes in the mesosphere. This includes high latitude cooling, as well as a reversal of the zonal mean zonal winds from westward to eastward (the opposite as in the stratosphere) (Labitzke, 1982; H.-L. Liu & Roble, 2002; Hoffmann et al., 2007; Siskind et al., 2010; Limpasuvan et al., 2016). The mesospheric changes during SSWs are primarily due to changes in gravity wave drag. The weakening, and potential reversal, of the eastward stratospheric winds leads to more eastward propagating gravity waves reaching the mesosphere, where, upon breaking, they increase the eastward forcing at mesospheric altitudes. The enhanced eastward forcing leads to the reversal of the mesospheric winds, and also changes the residual circulation in the high latitude mesosphere from downward to upward, resulting in adiabatic cooling of the mesosphere (H.-L. Liu & Roble, 2002; Siskind et al., 2010; Limpasuvan et al., 2016). The altered stratosphere-mesosphere residual circulation during SSW events may also lead to a warming of the summer hemisphere mesosphere, and a decrease in the occurrence of polar mesospheric clouds (Karlsson et al., 2007, 2009; Körnich & Becker, 2010). Though Körnich and Becker (2010) originally explained the coupling between wintertime SSWs and mesospheric warmings in the summer hemisphere as due to altered wave forcing in the summer hemisphere, A. K. Smith et al. (2020) recently proposed that the inter-hemispheric coupling is due to changes in the stratosphere-mesosphere circulation, and not due to modified wave forcing in the summer hemisphere mesosphere. The mesospheric changes that occur during SSWs are only weakly correlated with the changes that occur in the stratosphere (e.g. A. K. Smith et al., 2020), and there is significant event-to-event variability (Zülicke & Becker, 2013; Zülicke et al., 2018). The lack of a direct linear correspondence between the stratosphere and mesosphere illustrates the complexity of the coupling processes.

The circulation changes in the upper stratosphere and mesosphere that are discussed above lead to notable changes in chemical transport, altering the distribution of chemical species in the stratosphere and mesosphere. Changes in chemistry are particularly

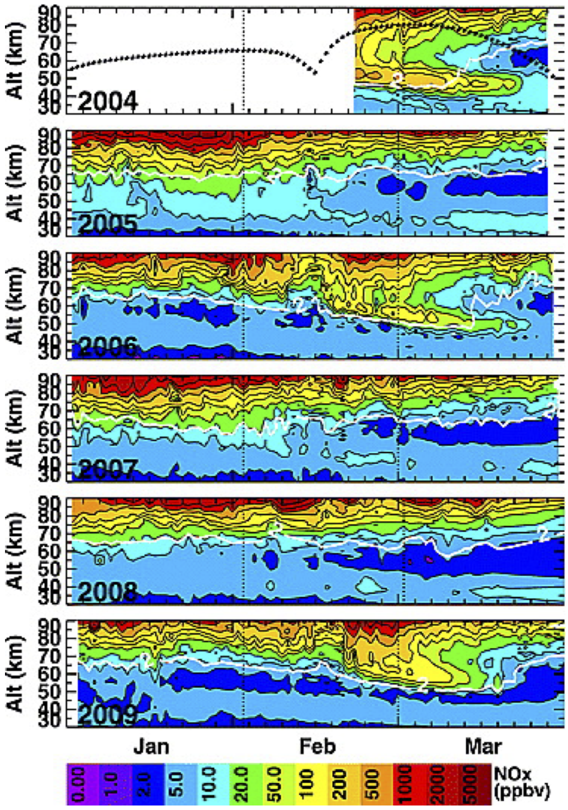


Figure 9. Zonal average ACE-FTS NOx (color) in the NH from 1 January through 31 March of 2004-2009. The white contour indicates CO=2.0 ppmv. Measurement latitudes are shown in the top panel as black dots. From Randall et al. (2009)

notable following elevated stratopause events, when there is significantly enhanced downward transport in the lower mesosphere and upper stratosphere (e.g., Siskind et al., 2015). The enhanced downward transport leads to enhancements in NOx and CO in the stratosphere (Manney, Harwood, et al., 2009; Randall et al., 2006, 2009). Observations of NOx during the winters of 2004-2009 are shown in Figure 9, clearly illustrating the enhanced downward transport of NOx during the winters of 2004, 2007, and 2009 during which major SSWs occurred. An increase in NOx is particularly relevant as it can lead to the loss of stratospheric ozone. Though enhanced descent of trace species is well observed, the descent in numerical models is typically too weak, leading to simulations with a deficit in NOx and CO in the stratosphere following SSW events (Funke et al., 2017). This is partly due to inadequate representation of the mesospheric dynamics (Meraner et al., 2016; Pedatella et al., 2018), though may also be due to insufficient source parameterizations (Randall et al., 2015; Pettit et al., 2019).

The changes in stratosphere-mesosphere chemistry and zonal winds during SSWs influence solar and lunar atmospheric tides, which, in-turn, play a key role in coupling SSWs to variability in the ionosphere and thermosphere. The most notable changes are an enhancement in the migrating semidiurnal solar and lunar tides. The migrating semidiurnal solar tide is primarily generated by stratospheric ozone, and Goncharenko et al. (2012) proposed that it is enhanced during SSWs due to changes in stratospheric ozone. However, recent numerical experiments by Siddiqui et al. (2019) demonstrate that the migrating semidiurnal solar tide in the lower thermosphere is primarily enhanced due to altered wave propagation, with ozone only being a minor (∼20-30%) contributor to the maximum enhancement. Though generally small, the migrating semidiurnal lunar tide is greatly enhanced during SSWs, and can obtain amplitudes equal to or larger than the migrating semidiurnal solar tide (e.g., Chau et al., 2015). The enhanced lunar tide is attributed to changes in the background zonal mean zonal winds, which shifts the Pekeris resonance mode of the atmosphere close to the frequency of the migrating semidiurnal lunar tide (Forbes & Zhang, 2012). The magnitude and timing of the semidiurnal lunar tide enhancements appear to be correlated with the stratospheric variability (Zhang & Forbes, 2014; Chau et al., 2015), though, as discussed in Chau et al. (2015), there are events that do not follow the linear relationship.

8.2 Impacts on the ionosphere

The influence of SSWs on the ionosphere was first hypothesized several decades ago by Stening (1977) and Stening et al. (1996). However, it was not until Goncharenko and Zhang (2008) and Chau et al. (2009) that the impact of SSWs on the ionosphere was unequivocally demonstrated. Since these studies there has been considerable research into the role of SSWs on generating variability in the low-latitude and mid-latitude ionosphere.

Observations have revealed that the low-latitude ionosphere exhibits a consistent response to SSWs, with an increase in vertical plasma drifts and electron densities in the morning and a decrease in the afternoon (Figure 10). The morning enhancement and afternoon depletion gradually, over the course of several days, shifts towards later local times (e.g., Chau et al., 2009; Goncharenko, Chau, et al., 2010; Goncharenko, Coster, et al., 2010; Fejer et al., 2011). This behavior is primarily attributed to the enhancement of the solar and lunar migrating semidiurnal tides during SSWs, which influence the generation of electric fields via the E-region dynamo mechanism (Fang et al., 2012; Pedatella & Liu, 2013). The migrating semidiurnal lunar tide is thought to be especially important in producing the gradual shift of the ionospheric perturbations towards later local times. As demonstrated by Siddiqui et al. (2015), there is a linear relationship between the strength of the stratospheric disturbance and the magnitude of the semidiurnal lunar tide in the equatorial electrojet. Numerical modeling studies indicate that the influence of SSWs on the low-latitude ionosphere should be larger during solar minimum compared to solar maximum (Fang et al., 2014; Pedatella et al., 2012). Observations have, however, revealed that equally large responses can occur during solar maximum (Goncharenko et al., 2013), indicating that factors in the lower-middle atmosphere, such as the SSW strength and lifetime, may be equally as important as solar activity.

A number of studies have investigated the impact of SSWs on the low-latitude ionosphere in different longitudes. They have found that the characteristic features of the ionosphere variability during SSWs is broadly similar across longitudes (Anderson & Araujo-Pradere, 2010; Fejer et al., 2010; Siddiqui et al., 2017). There are, however, differences in the response at different longitudes. In particular, the response is strongest, and tends to occur earliest, over South America. The longitudinal differences are related to the effects of nonmigrating semidiurnal tides and the influence of Earth's geomagnetic main field (Maute et al., 2015).

One of the reasons that the ionosphere variability during SSWs has attracted attention is that it potentially enables improved forecasting of ionosphere variability. Due to being primarily an externally forced system, the ionosphere and thermosphere are less sensitive to initial conditions compared to the troposphere-stratosphere (Siscoe & Solomon, 2006). This leads to skillful forecasts of the ionosphere being typically less than 24 h (Jee et al., 2007). If, however, the external drivers of ionosphere variability can be well-forecast, then the length of skillful ionosphere forecasts can be extended. The two external drivers of the ionosphere are solar activity, and effects from the lower atmosphere. The relatively good predictability of SSWs means that they could enable enhanced ionosphere forecast skill by improved forecasting of the lower atmospheric driver of ionosphere variability.

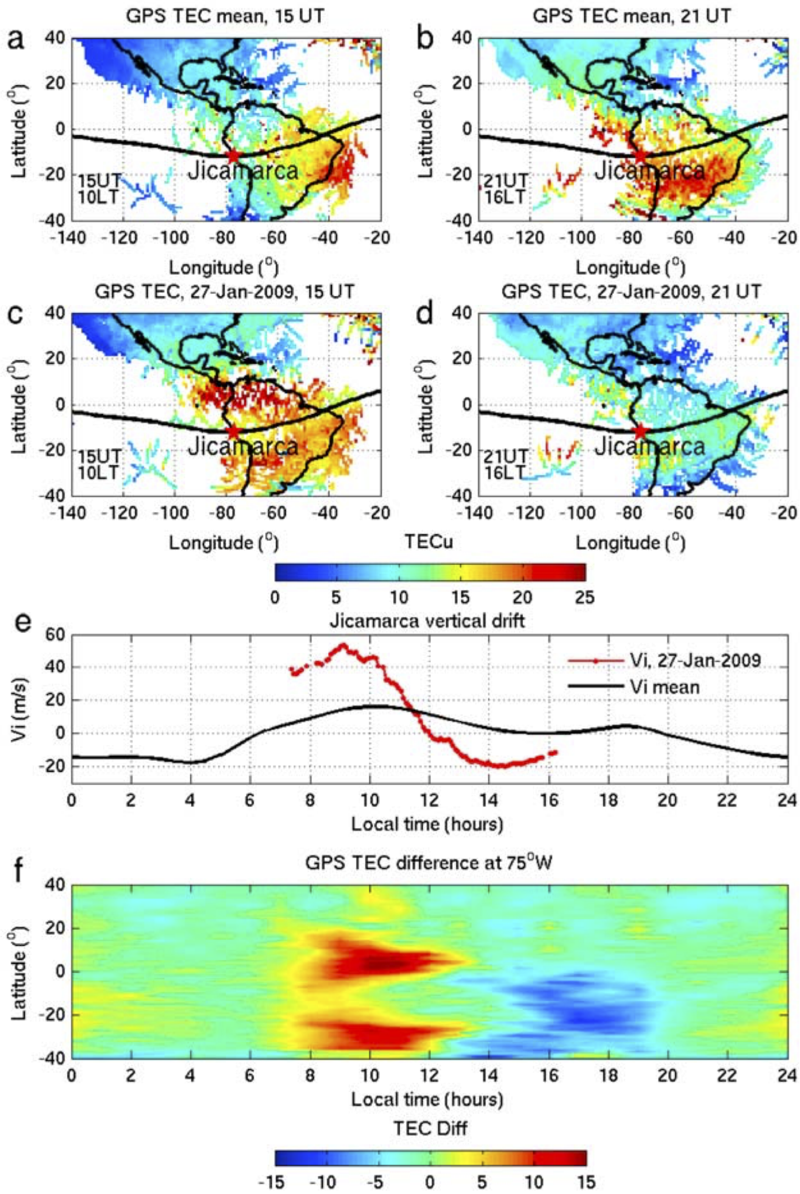


Figure 10. Observations of ionospheric behavior during the 2009 SSW event. (a) Mean total electron content (TEC) at 15 UT (morning sector, 10LT at 75). (b) Same as Figure 10a, except for at 21 UT (afternoon sector, 16 LT at 75). (c) TEC in the morning sector (15 UT) on January 27, 2009, during the SSW. (d) TEC in the afternoon sector (21 UT) on January 27, 2009. (e) Vertical drift observations by the Jicamarca incoherent scatter radar (12, 75) at 200-500 km altitude. The red line indicates observations on January 27, 2009, and the black line indicates the average behavior for winter and low solar activity. (f) Change in TEC at 75during the SSW as a function of local time and latitude. From Goncharenko, Chau, et al. (2010)

The ability to forecast the low-latitude ionosphere during the 2009 SSW was investigated by Wang et al. (2014) and Pedatella et al. (2018). Both studies found that the ionosphere variability could be forecast $\sim$10 days in advance of the SSW, which is consistent with the ability to predict the occurrence of SSWs. SSWs may thus provide a pathway for improving forecasts of the ionosphere.

The effects of SSWs on the ionosphere extend to middle latitudes, and are, perhaps surprisingly, stronger in the SH. Fagundes et al. (2015) and Goncharenko et al. (2018) both observed notable daytime enhancements in the SH middle latitude ionosphere. Goncharenko et al. (2018) also observed large decreases in nighttime ionosphere electron densities at middle latitudes. The mechanism generating variability in the middle latitude ionosphere is thought to be changes in the thermosphere neutral winds, and the greater response in the SH has been interpreted as being due to a larger amplitude semidiurnal lunar tide in the SH which propagates upwards into the thermosphere where it modulates the neutral winds (Pedatella & Maute, 2015).

Understanding the formation of small-scale irregularities in the ionosphere, often referred to as spread-F, equatorial plasma bubbles, or scintillation, is important owing to the disruptive impact of small-scale irregularities on communications and navigation (e.g., GPS) signals. Determining the role of SSWs on the formation of ionosphere irregularities is thus of considerable interest. Current observational evidence of the impact of SSWs on ionosphere irregularities is inconclusive, with some studies suggesting a suppression of irregularities (de Paula et al., 2015; Patra et al., 2014), and others an enhancement of irregularities (Stoneback et al., 2011). This is therefore an area that requires considerably more research.

8.3 Impacts on the thermosphere

The impact of SSWs on the thermosphere has received considerably less attention compared to the ionosphere. This is primarily due to the limited number of direct observations as well as generally smaller impacts of SSWs on the thermosphere. Nonetheless, investigations have revealed that there are clear impacts on the thermosphere temperature, density, and composition.

Numerical simulations by H.-L. Liu and Roble (2002) first revealed that the effects of SSWs can extend into the lower thermosphere. They found that the lower thermosphere ($\sim$110-170 km) in the NH warms by *ssh* $\sim$20-30 K during a SSW. Warming of the Northern Hemisphere lower thermosphere was confirmed observationally by Funke et al. (2010). Subsequent simulations by H. Liu et al. (2013) using the GAIA whole atmosphere model revealed that the zonal mean temperature changes globally, and throughout the thermosphere. In particular the GAIA simulations revealed upper thermosphere cooling in the tropics and Southern Hemisphere, and a global average cooling of $\sim$10 K during the 2009 SSW. The global cooling of the thermosphere is largely attributed to the dissipation of enhanced semidiurnal solar and lunar tides during the SSW, which significantly alters the circulation of the lower thermosphere (H. Liu et al., 2014). The cooling of the thermosphere leads to a contraction of the thermosphere, and a reduction in the neutral density at a fixed altitude. Based on satellite orbital drag derived thermosphere densities, Yamazaki et al. (2015) investigated the thermosphere density response to SSW events. They found a 3-7% decrease in global mean thermosphere density at altitudes of 250-575 km.

The composition of the thermosphere is also impacted by SSWs, with model simulations and observations finding a $\sim$10% reduction in the ratio of atomic oxygen to molecular nitrogen ($[O]/[N_2]$) during SSW events (Pedatella et al., 2016; Oberheide et al., 2020). This reduction arises due to the enhancement of migrating semidiurnal solar and lunar tides during the SSW, and their influence on the mean meridional circulation. In particular, the dissipation of the tides induces a westward momentum forcing in the lower thermosphere, which drives a mean meridional circulation that is upward in the equatorial region, poleward at middle latitudes, and downward at high latitudes. This altered mean meridional circulation leads to an increase of [O] and a decrease of [N_2] in the lower thermosphere that is then communicated to the upper thermosphere via molecular diffusion (e.g., Yamazaki & Richmond, 2013). As thermospheric [O]/[N_2] influences the production and loss of O^+, the [O]/[N_2] reduction during SSWs leads to a decrease in the diurnal and zonal mean ionosphere electron densities, which are approximately equal to O^+ in the F-region ionosphere.

9 Chemical/tracer aspects

The dramatic dynamical perturbations during SSWs are associated with anomalies in the transport circulation, and thus lead to anomalies in stratospheric constituents such as ozone and other trace gases throughout the atmosphere, with the impacts on the upper stratosphere and lower mesosphere discussed in the previous section.

It has been known since the mid-twentith century that the winter is dynamically the most active season in the stratosphere (see Baldwin et al. (2019)), and it was also correctly anticipated that the largest ozone changes would also occur during this season. However, measurements of both total column and profile ozone remained sparse before the 1970s and also exhibited large differences among measurement stations, leading to large uncertainties in deriving knowledge on natural variability in ozone and its drivers. Nevertheless, the influence of SSWs was recognised and could be shown from observations as early as in the late 1950s, with Dütsch (1963) revealing a close spatial correlation between total column ozone and temperatures in the 50-10 hPa layer during the 1957-1958 SSW. Based on averaged total column ozone observations over all available stations north of 40°N, Züllig (1973) further developed the findings by Dütsch to show that the seasonal evolution of ozone was exhibiting a much stronger initial increase during two years with SSW events (1962-1963 and 1967-1968) than during a year without an SSW event (1966-1967).

In the early 1970s, the Backscatter Ultraviolet (BUV) instrument on Nimbus IV provided the first global ozone data from space, with which the findings by Dütsch (1963) and Züllig (1973) from single measurement stations of total column ozone during SSWs could be verified (Heath, 1974). These global satellite measurements have continued to date, with a series of SBUV, TOMS, GOME, and OMI instruments flown on different satellites, providing immediate information on the impact of SSWs on total column ozone distributions in a visual manner. Figure 11 (left column) shows the total column ozone distribution over Antarctic in 2002, before and after the occurrence of the SSW. This event was the first to be observed in the SH and as mentioned above led to an impressive split of the 2002 Antarctic ozone hole (Varotsos, 2002; von Savigny et al., 2005), at least partially cutting short ozone depletion during that year (Weber et al., 2003). A similar event is shown on the right for the winter 1989 over the Arctic. While the overall ozone levels are much higher than in the SH, the split vortex can be clearly identified.

The clear signature of SSWs in total column ozone can be seen in the vertical structure of ozone (e.g Kiesewetter et al., 2010); de la Cámara et al. (2018). With the advent of stratospheric limb sounders in the late 1970s, a wealth of observations had become available to study these features, also in transport trace gases other than ozone such as nitrous oxide, carbon monoxide, and nitrogen oxides (e.g. Manney, Schwartz, et al., 2009; Manney, Harwood, et al., 2009; Tao et al., 2015). As shown by Kiesewetter et al. (2010) or de la Cámara et al. (2018), after onset of an SSW, ozone anomalies become positive above 500 K and negative below. The positive anomalies then slowly descend to lower levels, with the middle stratosphere relaxing back to normal levels the fastest. The enhanced poleward and downward transport during an SSW will lead to an increase in transport of other species such as carbon monoxide as well, with the breakdown of the po-

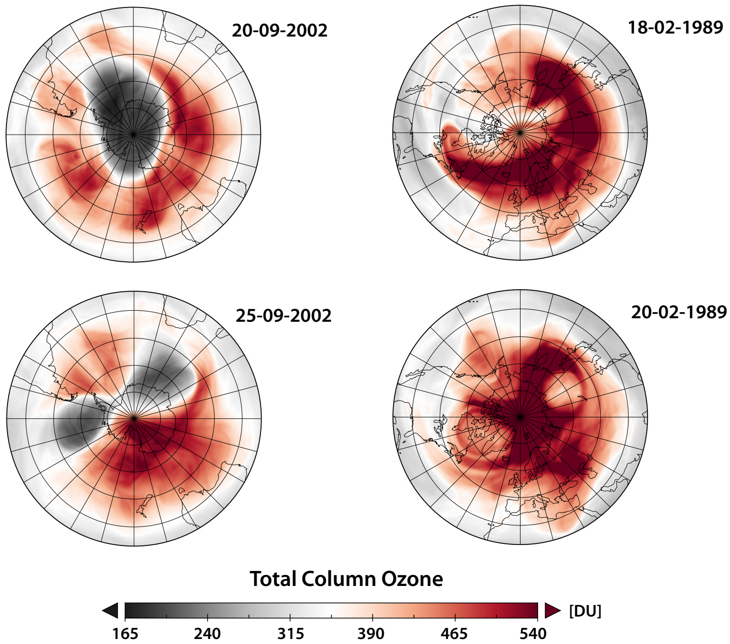


Figure 11. Total column ozone distributions in [DU] as obtained from ERA5 before (upper panels) and after (lower panels) an SSW event in 2002 in the Antarctic (left column) and in 1989 in the Arctic polar region, respectively.

lar vortex leading to enhanced mixing between mid and high latitudes and a flattening of the tracer gradients (Manney, Schwartz, et al., 2009). This will lead to cutting short ozone depletion by halogens in the Arctic polar stratosphere during spring, the opposite as found during the very cold and undisturbed 2013 Arctic winter that featured unprecedented Arctic ozone loss (Manney et al., 2015).

Due to the highly variable character of SSWs, observations fall however short of providing the statistical information needed to fully explain trace gas transport during these events, hence models are used to more closely investigate the drivers behind the transport. Local tracer mixing ratios are the results of a balance between chemical sources or sinks and transport. In the TEM framework, the equation for a tracer mixing ratio X can be written as in equation 3:

$$\frac{\partial \overline{X}}{\partial t} = -\left[\overline{v}^* \frac{\partial}{\partial y} + \overline{w}^* \frac{\partial}{\partial z}\right] \overline{X} + \nabla \cdot M + S \tag{3}$$

The chemical sources and sinks are represented by S, while the first two terms on the right hand side represent transport: The first term describes slow residual advection, with upward transport in the tropics and downward transport in the extratropics (see also Section 4). The second term is the divergence of eddy tracer fluxes (of the form $\left(\overline{v'X'}, \overline{w'X'}\right)$), and thus describes the effect of mixing processes. The latter arises due to stirring of tracer contours and subsequent small-scale diffusion, leading to no net mass transport, but, in the presence of tracer gradients to tracer transport.

As described in Section 4, the strongly enhanced wave forcing prior and during a SSW event drives a strongly enhanced residual circulation. High latitude downwelling is enhanced by up to one standard deviation between about 10 days prior the SSW up to the central date (de la Cámara et al., 2018). After the central date, wave propagation is mostly prohibited and subsequently the lack of wave forcing leads to weakened polar downwelling. The weakening of the residual circulation can persist up to two months after the SSW, in particular for "PJO" events (de la Cámara et al., 2018; Hitchcock et al., 2013). The extended persistence in the lower stratosphere is partly a result of longer radiative timescales in the lower stratosphere (Hitchcock et al., 2013), but it has been shown that also enhanced diffusive PV mixing leads to the prolonged recovery phase of the polar vortex (de la Cámara et al., 2018; Lubis et al., 2018).

Next to the anomalous vertical residual advection in the polar vortex region, tracers are affected by anomalous mixing during SSW events. Mixing, as measured by effective diffusivity or equivalent length [a measure of the disturbances of a tracer contour line relative to a zonally symmetric contour line (see Nakamura (1996)], is enhanced in the aftermath of SSW events: strongest anomalies are found around 10 days after the central date at the vortex edge in the mid-stratosphere, with anomalies propagating poleward and downward in the following weeks to months (de la Cámara et al., 2018; Lubis et al., 2018). Enhanced mixing in the lower stratosphere is found to persist for more than two months for "PJO" events (de la Cámara et al., 2018), largely equivalent to "absorptive" events as classified by Lubis et al. (2018). Note that those prolonged diffusive mixing anomalies of PV delay the vortex recovery (see above); however they are not necessarily associated with enhanced eddy PV fluxes (or negative EP flux divergence), but rather are compensated by wave activity transience, as revealed by an analysis of finite-amplitude wave activity (Lubis et al., 2018). The exact mechanism of the lower stratospheric mixing enhancement remains to be understood.

In summary, prior and during an SSW tracers are affected mostly by enhanced downwelling, while after the SSW, downwelling is reduced and at the same time, enhanced quasi-horizontal mixing sets in. Together with the eroded polar vortex, and thus eroded transport barrier (see, e.g., Tao et al. (2015)), enhanced mixing between mid-latitude and high latitude air will affect tracer concentrations after SSW events.

10 Outlook

Perhaps the most important outstanding question regarding SSWs is if they will be affected by climate change. Will the frequency of SSWs be affected by increasing greenhouse gas concentrations? Despite many efforts in the last 30 years (e.g Rind et al., 1990; Butchart et al., 2000; McLandress & Shepherd, 2009b; Mitchell et al., 2012), the answer remains unclear. Analyses of SSWs in the two most recent multi-model intercomparison projects (CCMI and CMIP6) do not provide a robust answer. Ayarzagüena et al. (2018) shows, *on average* in CCMI models, insignificant future changes in SSWs. Most individual CMIP6 models do project significant changes though, but with no consensus on the sign of the change (Ayarzagüena et al., 2020). The uncertainty in the sign of the response can, in part, be attributed to the opposing climate change effects of enhanced CO_2-cooling of the stratosphere and increased adiabatic warming from a faster Brewer-Dobson circulation leading to a large spread between models (Oberländer et al., 2013). Understanding how models project future changes in SSW frequency may have to do with the representation of the mean stratospheric state and how it reacts to climate change.

There are also outstanding questions over the factors influencing variability/likelihood of SSWs. The underlying observational limitation is that the relatively short observational record (which has large internal variability) must be interpreted with caution (Polvani et al., 2017). For example, SSW occurrence was significantly reduced in the 1990s relative to the 2000s (Domeisen, 2019) and it is not clear if this decadal variability occurred by chance or was perhaps due in part to, say, ocean variability or sea-ice loss (Garfinkel et al., 2017; Hu & Guan, 2018; Sun et al., 2015). Separating the effects of internal variability on the occurrence of SSWs from other influences (e.g. ocean variability, solar cycles, the QBO) is essentially not feasible on a statistical basis alone, due to the short data record and multiple potential factors influencing SSWs. Quantifying these effects will require a combination of theory and modelling, though again confidence in the results is likely to depend on the fidelity of the simulated SSWs (e.g. are the models reproducing the mechanisms correctly). Further, as confidence in theory and modelling improves it is possible that somewhat different answers are obtained than from the limited observational record.

The effects of SSWs (and stratospheric variability in general) on surface weather and climate are well quantified, but not completely understood. In particular, we do not have a good understanding of how the troposphere amplifies the stratospheric signal. Accurately simulating the effects of the stratosphere on surface weather will depend on identifying those aspects of the models which require improvement and is relevant on all timescales from weather forecasts to climate projections. Unlike the surface effects, the upward effects of SSWs above stratosphere are less well quantified and it is yet to be established if these effects are largely limited to SSWs or if the effects are proportional to stratospheric disturbances of either sign (White et al., 2020).

Acknowledgments

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