Tropospheric and Stratospheric Precursors to the January 2013 Sudden Stratospheric Warming

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ABSTRACT

This paper investigates the tropospheric and stratospheric precursors to a major sudden stratospheric warming (SSW) that began on 6 January 2013. Using the Climate Forecast System Reanalysis dataset, the analysis identified two distinct decelerations of the 10-hPa zonal mean zonal wind at 65°N in December in addition to the major SSW, which occurred on 6 January 2013 when the 10-hPa zonal mean zonal wind at 65°N reversed from westerly to easterly. The analysis shows that the two precursor events preconditioned the stratosphere for the SSW.

Analysis of the tropospheric state in the days leading to the precursor events and the major SSW suggests that high-latitude tropospheric blocks occurred in the days prior to the two December deceleration events, but not in the days prior to the SSW. A detailed wave activity flux (WAF) analysis suggests that the tropospheric blocking prior to the two December deceleration events contributed to an anomalously positive 40-day-average 100-hPa zonal mean meridional eddy heat flux prior to the SSW. Analysis of the stratospheric structure in the days prior to the SSW reveals that the SSW was associated with enhanced WAF in the upper stratosphere, planetary wave breaking, and an upper-stratospheric/lower-mesospheric disturbance. These results suggest that preconditioning of the stratosphere occurred as a result of WAF initiated by tropospheric blocking associated with the two December deceleration events. The two December deceleration events occurred in the 40 days prior to the SSW and led to the amplification of wave activity in the upper stratosphere and wave resonance that caused the January 2013 SSW.

1. Introduction

Sudden stratospheric warming (SSW) events are robust atmospheric phenomena that split or displace the stratospheric polar vortex and can precede extreme weather regimes in the troposphere (e.g., Baldwin and Dunkerton 1999, 2001; Baldwin et al. 2003; Shepherd 2002; Polvani and Waugh 2004). Although a consensus for the exact definition of an SSW is currently an ongoing discussion, major SSWs are typically characterized by a reversal of the 10-hPa zonal mean zonal wind from westerly to easterly poleward of 60° latitude (e.g., McInturff 1978; Butler et al. 2015). This zonal mean zonal wind reversal is accompanied by a poleward increase of the zonal mean temperature at 10 hPa, which is consistent with thermal wind balance. Skillfully predicting the development, evolution, and potential impacts of SSWs is a known forecasting challenge for numerical weather prediction (NWP) models (e.g., O’Neill 2003; Tripathi et al. 2015). One pathway to improving NWP model skill of SSWs is to improve the currently incomplete understanding of the dynamical mechanisms responsible for initiating the breakdown of the polar vortex during a major SSW (e.g., Gerber et al. 2012; Tripathi et al. 2015).

The goal of this study is to provide a better understanding of the precursors to, and potential mechanisms responsible for, major SSWs by investigating the precursors to the major SSW that occurred on 6 January 2013. The January 2013 SSW received wide attention within the forecasting community as major weather disruptions (e.g., extreme snowfall amounts and cold air outbreaks) occurred several weeks after the SSW over large areas of the Northern Hemisphere midlatitudes.
(e.g., Slingo 2013). This analysis begins by presenting background on the dynamics of SSWs in section 2 and describes the methods employed to identify the precursors to the January 2013 SSW in section 3. Two separate events characterized by a rapid deceleration in the 10-hPa zonal mean zonal wind and the major SSW are described in section 4. Section 5 provides a dynamical analysis of the events described in section 4, and section 6 provides a summary of the results and suggestions for future work.

2. Background

The dynamics of SSWs are commonly examined in a zonal mean perspective using the transformed Eulerian mean (TEM) framework (e.g., Dunkerton 1980). The zonal mean zonal momentum equation in the TEM framework states that

\[ \frac{\partial \vec{u}}{\partial t} = f_o \sigma^v + \rho_o^{-1} \nabla \cdot \vec{F} + \vec{X}, \tag{1} \]

where \( \bar{u} \) is the zonal mean zonal wind, \( f_o \) is the Coriolis parameter, \( \sigma^v \) is the residual zonal mean meridional wind, \( \rho_o \) is the base-state density, \( \vec{F} \) is the Eliassen–Palm (EP) flux vector, and \( \vec{X} \) represents the zonal forcing due to small-scale eddies. In the spherical-isobaric coordinate framework presented by Edmon et al. (1980), the EP flux vector can be defined as

\[ \vec{F} = \langle F_\phi, F_p \rangle = \left\langle -a \cos \phi (\bar{u} \bar{v}), a \cos \phi \frac{\partial \bar{v}}{\partial \bar{p}} \right\rangle, \tag{2} \]

where \( a \) is the radius of Earth, \( \phi \) is the latitude, \( u \) is the zonal wind, \( v \) is the meridional wind, \( f \) is the Coriolis parameter, \( \theta \) is the potential temperature, the overbar represents a zonal mean, the prime represents the deviation from the zonal mean, and the \( p \) subscript represents the partial derivative with respect to pressure. It can be seen from Eq. (2) that \( F_\phi \) is proportional to the zonal mean meridional eddy momentum flux and \( F_p \) is proportional to the zonal mean meridional eddy heat flux.

A first-order approximation of Eq. (1) reveals that changes in \( \vec{u} \) with time are directly proportional to the divergence of the EP flux vector. An easterly acceleration (equivalent to the weakening) of the polar vortex, as occurs during the onset of SSWs, is thus associated with EP flux convergence near the zonal mean stratospheric jet. Additionally, the simplified EP theorem (e.g., Andrews et al. 1987, p.131) assumes that dissipation and generation of wave activity flux by frictional and diabatic processes are negligible on the time scales of SSWs, meaning that the EP flux convergence is related to planetary-scale wave transience via

\[ \frac{\partial A}{\partial t} = -\nabla \cdot \vec{F}. \tag{3} \]

In Eq. (3), wave transience is given by the time derivative of \( A \), which represents the wave activity density, defined as

\[ A = \rho_o (\bar{q}^2) \frac{\partial^2}{\partial \bar{y}^2}. \]

where \( q \) is the quasigeostrophic potential vorticity (PV), the overbar represents the zonal mean, and the prime represents the departure from the zonal mean (Andrews et al. 1987, p.133). Through the vertical component of the EP flux vector, upward wave activity flux is directly proportional to the meridional eddy heat flux. As Eq. (1) shows, EP flux convergence is associated with a deceleration in the zonal mean zonal wind with time, meaning that, via Eq. (3), increased wave activity density in the stratosphere must also accompany the dramatic weakening of the polar vortex associated with an SSW. An increase in \( A \) in the stratosphere during an SSW event is primarily described by planetary-scale Rossby wave transience and planetary-scale wave activity flux convergence.

Matsuno (1971) proposed that the planetary-scale waves responsible for increased wave activity flux originate in the troposphere and propagate vertically upward into the stratosphere. In this view, large-amplitude waves propagate from the troposphere to the stratosphere until reaching a critical level where vertical propagation is no longer possible. Once the waves reach such critical level, the waves break and deposit easterly momentum into the stratospheric westerly jet. This deceleration of the polar jet alters the planetary wave propagation characteristics of the stratosphere and allows for continuous wave activity flux convergence and wave breaking along the polar vortex edge, resulting in the reversal of the zonal mean circulation from westerly to easterly. The continuous weakening of the vortex consequently lowers the critical level, above which upward wave propagation is not possible. This descending critical level with time explains why, during major SSWs, the polar vortex breakdown occurs first in the upper stratosphere and then descends to the lower stratosphere.

Several studies have shown that one potential precursor to increased stratospheric wave activity flux convergence is tropospheric blocking (e.g., Quirroz 1986; Andrews et al. 1987; O’Neill 2003; Martius et al. 2009). However, thorough analysis by Taguchi (2008) did not
find a statistically significant association between tropospheric blocking and SSWs. In an effort to clarify these contradictory results, Woolings et al. (2010) showed that the location of tropospheric blocking is correlated to the tendency of the stratospheric polar vortex to weaken or strengthen. The results of Martius et al. (2009) showed that 25 out of 27 SSWs during 1957–2001 were preceded by tropospheric blocking events over the North Atlantic and/or the North Pacific during the 10-day period prior to SSWs. Martius et al. (2009) also showed that planetary-scale waves that accompanied the blocking events had a westward baroclinic tilt with height from 500 to 10 hPa. This baroclinic structure of the planetary-scale waves was associated with persistently large poleward eddy heat flux, specifically above 100 hPa where it indicated large upward EP flux in the lower stratosphere. Similarly, Polvani and Waugh (2004) found that the average 40-day poleward eddy heat flux near the tropopause (100 hPa) preceding periods of weak vortex events was anomalously large and positive. Polvani and Waugh (2004) did not strictly constrain their study to SSWs but used the Baldwin and Dunkerton (2001) definition of weak (strong) vortex events, which considers extreme negative (positive) values of the northern annular mode in the stratosphere. Their results suggest that it is important to consider the upward tropospheric wave activity flux on subseasonal time scales when considering the deceleration of the stratospheric polar vortex.

Although the troposphere is the source region for the waves that amplify and break during SSWs, recent studies suggest that the characteristics of the stratospheric mean flow are also important for triggering SSWs (e.g., Albers and Birner 2014, and references therein). These studies suggest that in certain circumstances the stratospheric flow allows for planetary wave resonance, which can become the primary contribution to the increased wave amplitude leading to SSWs. In an effort to distinguish between the role of direct tropospheric wave forcing and stratospheric wave resonance, Albers and Birner (2014) performed a composite analysis of vortex splitting SSWs, which were defined as a splitting of the high PV reservoir at high latitudes and vortex displacement SSWs, which were defined as the displacement of the high PV reservoir equatorward (e.g., Charlton and Polvani 2007). Their composite analysis revealed that the leading pathway for SSW development differs depending on the type of the event, such that vortex displacement SSWs are characterized by increased wave transience associated with vertically propagating waves and vortex splitting SSWs are characterized by wave resonance. Albers and Birner (2014) attribute these differences to the characteristics of the zonal mean zonal circulation: the stratospheric polar vortex is anomalously strong and latitudinally narrow before splitting SSWs, while the stratospheric polar vortex is anomalously weak and wide before displacement SSWs.

Recent analyses have also highlighted the importance of the structure of the middle atmosphere leading to major SSWs (e.g., Manney et al. 2008; Thayer et al. 2010; Greer et al. 2013). Greer et al. (2013) demonstrated that upper-stratosphere/lower-mesosphere (USLM) disturbances preceded all major SSWs from 1991 to 2012. However, it was noted that not all USLM disturbances were associated with SSWs. Thayer et al. (2010) described USLM disturbances as synoptic-scale features with horizontal length and vertical depth scales on the order of 1000–2000 and 35 km, respectively, that are characterized by strong horizontal and vertical temperature gradients. A composite of USLM disturbances analyzed by Greer et al. (2013) showed that USLM disturbances occur coincident with upper-stratospheric planetary wave breaking. As Greer et al. (2013) described, a breaking wave in the upper stratosphere decelerates the climatological westerly wind and leads to an ageostrophic vertical motion response. The descent associated with this response produces warming at the stratopause, which, in the composite analysis, was manifested as locally high temperatures on the eastern edge of the polar vortex, between 0° and 90°E. This warm anomaly on the eastern edge of the polar vortex locally increases the horizontal temperature gradient and thus locally increases the vertical wind shear, supporting the growth of baroclinic instability. These narrow baroclinic zones can then provide a positive feedback on the large-scale geostrophic flow through the adiabatic ascent and descent that continues to warm the warm side of the baroclinic zone and cool the cold side of the baroclinic zone (e.g., Fairlie et al. 1990; Manney et al. 1994). This baroclinic instability in the upper stratosphere can trigger major SSWs by providing a conducive environment for enhanced persistent poleward heat flux within the USLM.

The goal of this study is thus to diagnose both the tropospheric and stratospheric structures that preceded the January 2013 major SSW. This SSW was previously examined by Vargin and Medvedeva (2015) and Coy and Pawson (2015). While both studies found evidence of increased wave activity flux in the days preceding the SSW, Vargin and Medvedeva (2015) attributed the enhanced wave activity flux to an eastward-propagating wave train in the upper troposphere and Coy and Pawson (2015) highlighted the contributions of a transient tropospheric system in the North Atlantic to the enhanced wave activity flux. Neither of these studies discussed the subseasonal evolution of the 10-hPa zonal
mean zonal wind. This analysis highlights the relevance of two precursor occurrences of rapid decelerations of the 10-hPa zonal mean zonal winds in the midlatitudes (Fig. 1a, to be discussed in section 4) to the SSW. This analysis will compare the SSW precursor flow features to the precursor flow features of the two precursor zonal mean zonal wind rapid decelerations. Such a comparison will explore the role of the two precursor rapid deceleration events in preconditioning the stratosphere for the major SSW. The goal of this analysis is to highlight the set of factors that contributed to the major SSW on synoptic to seasonal time scales.

3. Data and methods

This study employed the National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR) dataset (Saha et al. 2014) to investigate the precursors to the January 2013 SSW. The CFSR consists of an atmosphere–ocean–land–ice coupled data assimilation system with 64 sigma–pressure hybrid levels and horizontal resolution of approximately 100 km. The highest model level is approximately 0.26 hPa, which allows investigating both tropospheric and stratospheric phenomena. The 6-hourly analyses interpolated to 0.5° latitude × 0.5° longitude horizontal resolution were obtained (NCEP 2015) on isobaric surfaces up to 1 hPa. With the aim of investigating the precursors to the SSW, the 6-hourly analyses were evaluated between 0000 UTC 1 November 2012 and 1800 UTC 31 January 2013. Temperature, height, and wind anomalies with respect to climatology were computed relative to a 31-yr climatology (1979–2009) constructed with the consistent CFSR dataset.

The evolution of the stratospheric polar vortex in the CFSR dataset was assessed by means of the 10-hPa mean zonal wind. This analysis highlights the relevance of two precursor occurrences of rapid decelerations of the 10-hPa zonal mean zonal winds in the midlatitudes (Fig. 1a, to be discussed in section 4) to the SSW. This analysis will compare the SSW precursor flow features to the precursor flow features of the two precursor zonal mean zonal wind rapid decelerations. Such a comparison will explore the role of the two precursor rapid deceleration events in preconditioning the stratosphere for the major SSW. The goal of this analysis is to highlight the set of factors that contributed to the major SSW on synoptic to seasonal time scales.

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The evolution of the stratospheric polar vortex in the CFSR dataset was assessed by means of the 10-hPa
zonal mean zonal wind at 65°N. While the definition of an SSW is currently the center of an ongoing discussion in the stratosphere community (Butler et al., 2015), for this study an SSW is defined at 65°N, which is the latitude band employed by the World Meteorological Organization (WMO) to describe SSWs (e.g., Limpasuvan et al. 2004; Coughlin and Gray 2009). In addition to identifying the major SSW, the time series of the 10-hPa zonal mean zonal wind at 65°N also revealed two precursor decelerations of the zonal mean stratospheric circulation in December 2012 (Fig. 1a, to be discussed in section 4). A synoptic analysis of the troposphere and the stratosphere prior to the precursor decelerations and the major SSW was performed with the purpose of identifying the differences in the flow features during the periods leading to each deceleration of the zonal mean stratospheric circulation to elucidate the impact of the synoptic-scale structure on triggering the major SSW.

The analysis of the troposphere was aimed at identifying potential sources of upward propagating planetary-scale waves originating in the mid- to upper troposphere. Motivated by previous findings that have suggested a relationship between tropospheric blocking and upward-propagating waves prior to SSWs (e.g., Quiroz 1986; Andrews et al. 1987; O’Neill 2003; Martius et al. 2009), tropospheric blocking was quantified by adapting the method outlined by Schwierz et al. (2004) with some important differences. Blocking events were identified through Ertel’s potential vorticity (EPV) anomalies with respect to a 1979–2009 climatology. Instead of using a monthly EPV climatology as done by Schwierz et al. (2004), this study employed a midwinter (December and January) EPV climatology1 to allow for a direct comparison of the state of the upper troposphere in December 2012 and January 2013. A 2-day running mean was applied to the EPV anomalies to remove daily variability but to retain synoptic structures. Blocking events in the upper troposphere were defined by negative closed contours of EPV anomalies in the 310–330-K isentropic layer during a 5-day period prior to each instance of decelerating 10-hPa zonal mean zonal wind at 65°N. This 5-day-averaged blocking analysis was compared against analyses of the 300-hPa isobaric level 24-h prior to each instance to identify the main tropospheric features contributing to the block. Different averaging periods were explored, but the 5-day average revealed the most distinct synoptic and dynamic structures between the major SSW and the other decelerations of the zonal mean stratospheric circulation. As SSWs are, by definition sudden (i.e., occur on time scales of less than a week), this combined approach of averaging the 5 days prior to each stratosphere event and also analyzing the synoptic structure 24-h prior to each event allows for an accurate understanding of the dominate synoptic structure in the week leading to each event and the instantaneous state of the upper troposphere just prior to each event.

Along with the synoptic analysis, an analysis of planetary-scale wave activity flux was also performed and averaged over the 5-day period prior to each event. To determine the structure of planetary-scale waves, a Fourier decomposition was performed on the geopotential height. Only wavenumbers 1 and 2 from the Fourier decomposition are presented here in the interest of describing only the dominant components of the planetary-scale wave structure. Results from the Fourier decomposition were compared with an analysis of both the EP flux vectors and divergence to assess the impact of the wave activity flux on the zonal mean flow. Following Edmon et al. (1980), the EP flux vectors were calculated from Eq. (2) and the EP flux divergence in the spherical-isobaric coordinate framework was calculated as

\[ \nabla \cdot \mathbf{F} = \frac{1}{a \cos \phi} \frac{\partial (F_\phi \cos \phi)}{\partial \phi} + \frac{\partial (F_p)}{\partial p}, \]

To relate the EP flux divergence to the temporal change of the zonal mean zonal wind, Eq. (4) was scaled by factor of \((a \cos \phi)^{-1}\) to yield units of acceleration (m s\(^{-2}\)).

Last, the upper-stratospheric and lower-mesospheric precursors to the major SSW were also analyzed for the 2012/13 boreal winter. The analysis of the stratospheric state focused on the 30- and 10-hPa isobaric surfaces prior to the major SSW and the precursor deceleration events. This analysis was employed to investigate the evolution of the stratospheric polar vortex, as well as to explore the spatial variations of temperature, height, and wind in the weeks prior to the major SSW. To analyze upper-stratospheric wave breaking and the importance of upper-stratospheric processes highlighted by Greer et al. (2013), the CFSR dataset was interpolated from isobaric to isentropic surfaces extending from 330 to 2000 K. Using this interpolated dataset, EPV was analyzed on the 1500-K isentropic surface (near 2 hPa) to determine if upper-stratospheric wave breaking occurred in association with a USLM disturbance. The 1500-K isentropic surface was employed as it is an isentropic surface near 2 hPa and, as Thayer et al. (2010) stated, 2 hPa can be the lower boundary of an USLM disturbance. Employing an isentropic surface

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1 The climatology was also constructed using a monthly mean as in Schwierz et al. (2004) without a large change in the results (not shown).
near 2 hPa is also consistent with the Greer et al. (2013) USLM climatology, which defined USLM disturbances at an isentropic level near 2 hPa.

4. Event overview

a. The boreal winter of 2012/13

For this analysis, the 10-hPa zonal mean zonal wind at 65°N and the 10-hPa polar cap temperature, defined as the 60°–90°N area-weighted average temperature, will be used as the metrics to determine the date and characteristics of the January 2013 major SSW, hereafter referred to as “the SSW.” A 6-hourly time series of the 10-hPa zonal mean zonal wind at 65°N from 0000 UTC 1 November 2012 to 1800 UTC 31 January 2013 is shown in Fig. 1a. The SSW occurred at 1800 UTC 6 January 2013 when the zonal mean zonal wind reversed from westerly to easterly. The reversal of the winds during the SSW was associated with an instantaneous deceleration of the zonal mean zonal wind of \(-0.78 \text{ m s}^{-1} \text{ h}^{-1}\) (Fig. 1b) and was accompanied by a rapid 10-hPa polar cap temperature increase of 28 K in 7 days (Fig. 1c). By 0000 UTC 8 January 2013, 30 h after the 10-hPa 65°N zonal mean zonal wind reversed to easterly, the 10-hPa polar vortex split into two pieces, classifying this major SSW as a splitting event (Coy and Pawson 2015).

In both the days and months leading to the SSW the 100-hPa midlatitude zonal mean eddy heat flux anomaly was primarily large and positive. Figure 2 shows a time series of the 100-hPa zonal mean eddy heat flux anomaly averaged from 45° to 75°N. Following the methodology of Polvani and Waugh (2004), the zonal mean eddy heat flux anomaly was calculated with respect to climatology. After 28 November 2012, there were several episodes when the heat flux rapidly increased and reached magnitudes of around 40 K m s\(^{-1}\). For a comparison with the Polvani and Waugh (2004) study, a 40-day average of the eddy heat flux anomaly prior to the SSW was calculated, yielding a value of 11.7 K m s\(^{-1}\). This zonal mean eddy heat flux anomaly would be considered one of the largest 40-day-averaged heat flux anomalies prior to the weak vortex events had it been analyzed by Polvani and Waugh (2004). This large positive 40-day-average heat flux anomaly serves as motivation to focus on the analysis of the 40-day period prior to the SSW to elucidate what contributed to such a large 100-hPa-average heat flux anomaly. The analysis further aims to determine the role of the features that contributed to this 40-day-average heat flux anomaly in creating an environment conducive to the breakdown of the stratospheric polar vortex.

While the SSW was associated with the largest deceleration of the 10-hPa zonal mean zonal wind at 65°N during the 2012/13 boreal winter (Fig. 1b), there were also two separate occurrences of the zonal mean zonal wind rapidly decreasing to near 0 m s\(^{-1}\) (Fig. 1a). These two occurrences of rapid zonal mean zonal wind deceleration did not qualify as minor SSWs according to the McInturff (1978) definition, which states that there must be a 25-K temperature increase at any level in the stratosphere during a period of a week or less. However, these rapid decelerations of the 10-hPa zonal mean zonal wind at 65°N played an important role in preconditioning the stratosphere for the SSW and thus will be discussed in detail throughout the rest of this section and section 5a. The first of these rapid decelerations, hereafter referred to as “the first event,” occurred at
0600 UTC 5 December 2012 when the zonal mean zonal wind decreased to 8 m s$^{-1}$ (Fig. 1a). This zonal mean zonal wind minimum corresponded to an average decrease of $-0.16$ m s$^{-1}$ h$^{-1}$ from 1800 UTC 25 November to 0000 UTC 5 December 2012 (Fig. 1b). Concurrent with this rapid decrease in the 10-hPa 65°N zonal mean zonal wind was a maximum in the 100-hPa zonal mean eddy heat flux anomaly of 30.3 K m s$^{-1}$ at 0000 UTC 1 December 2012 (Fig. 2). Although the first event did not fulfill the requirements of an SSW as defined in this study, the polar vortex did not recover to the antecedent zonal mean zonal wind maximum of 47 m s$^{-1}$ obtained at 1200 UTC 15 November 2012. After the first event, the zonal mean zonal wind strengthened to a maximum of 36 m s$^{-1}$ at 0600 UTC 16 December 2012. The first event was also associated with a transient maximum in the 10-hPa polar cap temperature of 206 K on 6 December 2012 (Fig. 1c).

A second rapid deceleration of the 10-hPa zonal mean zonal wind at 65°N, hereafter referred to as “the second event,” occurred at 0600 UTC 26 December 2012 with a minimum zonal mean zonal wind of 5 m s$^{-1}$ (Fig. 1a). Associated with this decrease in the zonal mean zonal wind, the 10-hPa polar cap temperature gradually increased to 202 K on 28 December 2012 before rapidly increasing to 233 K on 7 January 2013 in association with the SSW (Fig. 1c). At 1200 UTC 23 December 2012, just before the maximum deceleration in the zonal mean zonal wind associated with the second event (Fig. 1b), there was a maximum in the 100-hPa zonal mean eddy heat flux anomaly of 42.4 K m s$^{-1}$ (Fig. 2). By 1800 UTC 2 January 2013, the zonal mean zonal wind recovered to the mid-December magnitude of 36 m s$^{-1}$, remaining 11 m s$^{-1}$ less than the mid-November maximum of 47 m s$^{-1}$ (Fig. 1a).

Both of the events were associated with positive tropospheric and stratospheric geopotential height anomalies (Fig. 3). Between 26 November and 1 December 2012, prior to the first event, there were high-latitude height anomalies exceeding $+0.75\sigma$ in the middle and upper troposphere (Fig. 3). These positive tropospheric anomalies preceded lower- and middle-stratospheric height anomalies of a similar sign and magnitude by approximately six days, suggesting that the positive height anomalies originated in the troposphere and progressed upward prior to and during the first event. A peak in the 100-hPa eddy heat flux anomaly around 1 December 2012 accompanied this upward progression of geopotential height anomalies (Fig. 2). Similarly to the first event, positive height anomalies ($\sim +0.5\sigma$) characterized the troposphere prior to the second event. These positive tropospheric height anomalies prior to the second event, however, were weaker than the positive tropospheric height anomalies prior to the first event (Fig. 3). Again, the positive height anomalies appeared to progress upward from the troposphere to the stratosphere over a 6-day period and, for this event, amplify with height. This upward progression of geopotential height anomalies was again accompanied by a peak in the near tropopause-level heat flux anomaly (Fig. 2).

In contrast to the precursor events, prior to the SSW there was no notable upward progression of positive height anomalies from the troposphere to the stratosphere in the week immediately preceding the SSW. Rather, in the period between the second event and the SSW, the largest positive height anomalies were in the stratosphere. On 8 January 2013, the positive height anomalies in the middle stratosphere exceeded $+1.5\sigma$. These positive stratospheric height anomalies then progressed downward into the troposphere into late
January, as suggested by the maximum in the geo-potential height anomaly appearing at lower levels with increasing time in Fig. 3. Although Fig. 3 shows standardized zonal mean geopotential height anomalies from climatology, this apparent descent of positive height anomalies from the upper stratosphere into the lower stratosphere is similar to Matsuno (1971), who analyzed anomalies from the zonal mean and showed that once waves hit a critical level in the upper stratosphere and break, the critical level becomes progressively lower and subsequent waves break at lower altitudes.

To analyze the vertical and meridional structure of the zonal mean zonal wind with height, a 5-day average of the zonal mean zonal wind and associated anomalies with respect to climatology prior to each of the events and the SSW is shown in Fig. 4. Prior to the first event the zonal mean zonal wind was anomalously weak through the troposphere and into the stratosphere poleward of 60°N. The strongest stratospheric zonal mean zonal winds exceeded 25 m s⁻¹ and were located at 10 hPa centered near 50°N (Fig. 4a). The strongest negative zonal mean zonal wind anomalies exceeded 20 m s⁻¹ and were centered near 75°N at 10 hPa. In the days prior to the second event, the negative zonal mean zonal wind anomalies shifted equatorward with the strongest negative zonal mean zonal wind anomalies near 62°N and 10 hPa (Fig. 4b). The strongest stratospheric zonal mean zonal winds remained centered at ~50°N at 10 hPa, but were stronger than prior to the first event, exceeding 30 m s⁻¹.

While the vertical and meridional structure of the zonal mean zonal wind and anomalies in the stratosphere was similar prior to the precursor events, in the days prior to the SSW, the zonal mean zonal wind structure was quite different. Prior to the SSW, the zonal mean zonal wind anomalies poleward of 75°N were negative below 100 hPa and positive above 100 hPa (Fig. 4c). The stratospheric negative zonal mean zonal wind anomalies were meridionally confined between ~55° and 75°N. This dipole structure of negative zonal mean zonal wind anomalies equatorward of positive zonal mean zonal wind anomalies suggests that the fastest winds associated with the stratospheric polar vortex were shifted anomalously poleward, which can be seen in Fig. 4c as the 10-hPa jet core was now centered at ~75°N. This anomalously poleward jet core allowed for wave activity to be directed into the polar region (e.g., Labitzke 1981; Butchart et al. 1982; McIntyre 1982; Albers and Birner 2014). This anomalous poleward shift of the strongest zonal mean zonal winds is consistent with the results of Albers and Birner (2014) who showed a similar pattern in the 5–10 days prior to splitting SSW events. The role of the precursor events in altering the vertical and meridional structure of the stratospheric zonal mean zonal winds and preconditioning the stratosphere in such a way to favor the development of the SSW will be explored in the following sections.

b. Tropospheric evolution

Previous studies have shown that high-latitude tropospheric blocks can act as precursors to SSWs because the associated amplified tropospheric flow pattern produces anomalous upward wave activity flux (e.g.,

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Figure 4. Latitude–height cross section of the 5-day-averaged zonal mean zonal wind [solid (dashed) black every 5 (~5) m s⁻¹] and anomalies with respect to climatology (shaded, m s⁻¹) prior to (a) the first minor event (0600 UTC 30 Nov–0600 UTC 4 Dec 2012), (b) the second minor event (0600 UTC 21 Dec–0600 25 Dec 2012), and (c) the SSW (1800 UTC 1 Jan–1800 UTC 5 Jan 2013).
The location of tropospheric blocks prior to the precursor events and the SSW is diagnosed in the blocking analysis shown in Fig. 5. The blocking analysis shows that the positive height anomalies in the troposphere in the days prior to the first event were associated with two mid- and high-latitude tropospheric blocks (Fig. 5a). These blocks were both represented by EPV anomalies of $-3 \text{ PVU}$ (1 PVU = $10^{-6} \text{Kkg}^{-1} \text{m}^2 \text{s}^{-1}$) and were located north of Greenland, centered on $90^\circ \text{W}$, and in the northern Pacific Ocean centered on $50^\circ \text{N}, 180^\circ$. The blocking pattern prior to the first and second event were similar (Figs. 5a,b). Prior to the second event there was a block over northeastern Canada, represented by a $-2 \text{ PVU}$ anomaly, centered near $60^\circ \text{N}, 60^\circ \text{W}$ and a block in the southern Arctic Ocean, also represented by a $-2 \text{ PVU}$ anomaly, centered at $165^\circ \text{W}$ (Fig. 5b). In contrast to the precursor events, no blocks were identified in the polar high latitudes (i.e., $\geq 65^\circ \text{N}$) prior to the SSW (Fig. 5c). The absence of tropospheric blocks at high latitudes was consistent with the small high-latitude positive tropospheric height anomalies prior to the SSW (Fig. 3) and suggests that tropospheric blocking in the days immediately preceding the SSW was not the primary forcing mechanism for the SSW. This result is similar to that of Albers and Birner (2014), who showed that in the case of the 2009 major splitting SSW, wave resonance in the stratosphere was the primary trigger for the SSW, while forcing from the troposphere was comparably weaker immediately preceding the SSW.

The following 300-hPa analysis of geopotential height and standardized anomalies 24-h prior to the two precursor events and the SSW highlights the variability in the precursor upper-tropospheric state prior to the perturbations in the stratospheric flow. Although the blocking analysis showed the location of blocks with respect to a midwinter climatology, the 300-hPa analyses provide a representative snapshot of the upper-tropospheric structure within the 5-day periods leading to the precursor events and the SSW and the associated standardized height anomalies with respect to the 31-yr climatology. At 0600 UTC 4 December 2012, 24 h prior to the first event, there was a zonally narrow, but meridionally stretched, $+1\sigma$ height anomaly that extended across the North Pole from $45^\circ \text{N}$, $160^\circ \text{W}$ to $75^\circ \text{N}$, $22^\circ \text{W}$ with two local $+2\sigma$ 300-hPa height anomalies located near $60^\circ \text{N}$, $180^\circ$ and over the pole (Fig. 6a). This broad positive height anomaly was associated with the high-latitude phasing of two synoptic-scale 300-hPa ridges: one located in the northern Pacific and the other located over northern Greenland. The locations of these two 300-hPa ridges...
were consistent with the locations of blocks identified by the blocking analysis (Fig. 5a).

At 0600 UTC 25 December 2012, 24 h preceding the second event, there was a $+3\sigma$ 300-hPa height anomaly located over northeastern Canada at 75°N, 90°W (Fig. 6b), which was again consistent with the location of the block identified in the blocking analysis (Fig. 5b). The 300-hPa height field highlights the differences between the first and second events in the structure and location of the synoptic-scale upper-tropospheric ridges identified by the blocking analysis. Prior to the first event, the synoptic-scale ridge was in the Pacific, whereas prior the second event the synoptic-scale ridge was over northern Canada.

At 1800 UTC 5 January 2013, 24 h prior to the SSW, a $+1\sigma$ anomalous ridge at 0° longitude extended from 45° to 75°N (Fig. 6c). This ridge was not identified as a block in the blocking analysis (Fig. 5c), as it was a short-lived feature at mid- and high latitudes. In contrast to the blocks over the northern Pacific and northeastern Canada seen in the 24-h period prior to the two precursor events, in the 24-h period prior to the SSW there was no ridging in the northern Pacific and northern Canada was characterized by a $-1\sigma$ anomalous trough (Fig. 6c). There was, however, a $+1\sigma$ anomalous ridge over the North Pole, but this ridge was not characterized as a block by the blocking analysis (Fig. 5c).

c. Stratospheric evolution

The following 30- and 10-hPa analyses compare the stratospheric structure prior to the precursor events and the SSW. Prior to the first event, at 0600 UTC 4 December 2012, the 300-hPa northern Pacific ridge corresponded with a $+3\sigma$ 30-hPa Aleutian ridge (Fig. 7a). This amplified Aleutian ridge coincided with the 30-hPa polar vortex becoming displaced off the pole and split with one center of circulation between 45° and 90°W and the other center of circulation between 45° and 90°E. The strongest negative height anomalies that accompanied this displaced polar vortex were in the Western Hemisphere at 112°W (Fig. 7a). At 0600 UTC 25 December 2012, 24 h prior to the second event, there was a $+2\sigma$ anomalous 30-hPa ridge (Fig. 7b) to the west of the 300-hPa anomalous ridge (Fig. 6b), with the maximum 30-hPa positive height anomaly over the Arctic Ocean’s Beaufort Sea. The anomalous 30-hPa ridge displaced the polar vortex toward central Siberia (Fig. 7b). At 1800 UTC 5 January 2013, 24 h prior to the SSW, both Alaska and the central Atlantic were characterized by $+1\sigma$ 30-hPa height anomalies that together produced an elongated and displaced polar vortex in the Eastern Hemisphere (Fig. 7c).

The 10-hPa geopotential height and temperature anomalies reveal noticeable differences between the
precursor events and the SSW. At 0600 UTC 4 December 2012, 24 h prior to the first event, the 10-hPa vortex was zonally elongated, as represented by low geopotential heights that extended between 90°W and 90°E at 65°N (Fig. 8a). Associated with this displaced vortex were locations with \(-2\sigma\) 10-hPa temperature anomalies between 90° and 135°W (Fig. 8a). At 0600 UTC 25 December 2012, 24 h prior to the second event, the polar vortex was displaced toward the Eastern Hemisphere (Fig. 8b). There were two regions of \(-1\sigma\) temperature anomalies at 10 hPa associated with this displaced vortex. The first negative temperature anomaly was on the western flank of the vortex, located between 90°W and 25°E, and the second negative temperature anomaly was on the eastern flank of the vortex, between 135°E and 135°W (Fig. 8b). At 1800 UTC 5 January 2013, 24 h prior to the SSW, the 10-hPa amplified Aleutian ridge displaced the elongated polar vortex toward 22°W (Fig. 8c). Associated with this amplified Aleutian ridge was a strong height gradient between the Aleutian ridge and the polar vortex and thus strong cross-polar winds and a \(+3\sigma\) temperature anomaly that extended from 45° to 75°N over Siberia (Fig. 8c). From mid-December to the beginning of January, the 10-hPa Aleutian ridge amplified, increasing the aspect ratio of the polar vortex, resulting in a strong height gradient and strong winds across the pole (Figs. 8b,c).

5. Dynamical analysis

a. Fourier decomposition and EP flux diagnostics

The tropospheric and stratospheric analyses presented in sections 4b and 4c describe the synoptic state of the atmosphere throughout the period leading to the SSW, while the following dynamical analysis investigates the association between the previously discussed anomalous flow structures and the breakdown of the stratospheric polar vortex. A longitude–height plot of the 5-day-averaged Fourier decomposition into wave-number 1 (WN1) and wavenumber 2 (WN2) of the height field along 65°N is shown in Fig. 9 for the 5-day period preceding each of the three events. The analysis of the vertical structure of the WN1 and WN2 patterns are used to examine the barotropic or baroclinic nature of the waves, where a westward tilt with height implies a baroclinic wave structure, the presence of poleward eddy heat fluxes, and upward wave activity flux.

Prior to the first event, the WN1 magnitude was maximized at \(\sim 3\) hPa with an amplitude of 1000 m while the WN2 magnitude was maximized at \(\sim 7\) hPa with an amplitude of 500 m. Both the WN1 and WN2 structures had a minimal vertical tilt with height above 50 hPa, suggesting the planetary waves were barotropic in the
stratosphere (Fig. 9a). This barotropic structure suggests there was limited poleward heat flux by planetary waves and thus a small vertical component to the wave activity flux in the stratosphere. Consistent with this barotropic structure in the stratosphere, the $45^\circ$–$75^\circ$N zonal mean eddy heat flux anomaly peaked at 30 hPa, decreasing in magnitude by 10 hPa (Fig. 2). Prior to the second event, the WN1 amplitude of 1400 m was maximized at $7\ hPa$ and far exceeded the WN2 amplitude, which peaked at 100 m at $50\ hPa$. Again, both the WN1 and WN2 structures were vertically stacked above 50 hPa, indicative of a barotropic wave structure in the stratosphere (Fig. 9a).
stratosphere (Fig. 9b), which was consistent with the 45°–75°N zonal mean eddy heat flux anomaly that peaked in the lower stratosphere at 30 hPa (Fig. 2). This analysis suggests that although there was high-amplitude tropospheric flow prior to the two precursor events, the precursor structure of the stratosphere, in which the polar night jet was located at midlatitudes (Fig. 4a), was not favorable for vertical propagation into the polar vortex (e.g., Limpasuvan et al. 2004), thus the upward wave activity flux was limited to the troposphere and lower stratosphere.

To compare the upward wave activity flux prior to the precursor events and the SSW, Fig. 10 shows the EP flux vectors and their convergence averaged over the 5-day periods prior to each of the events. All analyzed EP flux convergence prior to the precursor events and the SSW was dominated by the planetary-scale WN1 and WN2 (not shown). Prior to the first event, although the EP flux was large in the troposphere, it was relatively small in the stratosphere (Fig. 10a). There was only a small region of EP flux convergence centered at ~7 hPa and 75°N. As explained in Eqs. (1) and (3), the location of EP flux convergence corresponds to an increase in wave activity density and a decrease in the zonal mean zonal wind. Recall that in Eq. (1) EP flux divergence is scaled by the inverse of density, thus, when equal EP flux divergence exists in a vertical column, the impact of that divergence on the tendency of $\pi$ increases in magnitude with height. Therefore, this limited area of EP flux convergence in the lower-stratospheric high latitudes, which suggests a weak increase in the wave activity density, corresponds to a decrease in the zonal mean zonal wind. Although there was no analyzed EP flux convergence at 100 hPa, suggesting that the EP flux convergence was very small, the large vertical component of EP flux from ~55° to ~60°N suggests that there was upward wave activity flux near the tropopause, which was consistent with the large positive 100-hPa zonal mean eddy heat flux anomalies prior to the first event (Fig. 2). These large values of upward wave activity flux near the tropopause, however, appeared to be limited to the lower stratosphere as the vertical component of EP flux was small above 20 hPa (Fig. 10a) where the WN1 and WN2 geopotential height structures had limited westward tilt with height (Fig. 9a). In the stratosphere, the EP flux vector was dominated by the meridional component, which was directed equatorward, suggesting that, upon reaching the midlatitude stratosphere, the waves propagated equatorward and out of the midlatitudes (Fig. 10a).

Prior to the second event there was an increase in the stratospheric EP flux convergence suggesting there was also an increase in the wave activity density in the stratosphere (Fig. 10b). Above 50 hPa, there was a meridional pattern of convergence between ~40° and ~55°N, divergence between ~58° and ~70°N, and convergence poleward of ~75°N (Fig. 10b). This meridional pattern of EP flux divergence was dominated by the meridional component of EP flux and resulted in EP flux divergence at 65°N and 10 hPa, which was inconsistent with the decrease in the 10-hPa zonal mean zonal wind at 65°N that occurred prior to the second event (Fig. 1). Although the total EP flux was divergent in the stratosphere at 65°N and was dominated by the meridional component of EP flux, the weak vertical component of EP flux was convergent (not shown). Similarly to the first event, 100 hPa was again characterized by a large vertical component of EP flux from ~55° to ~60°N, which was consistent with the large 100-hPa zonal mean eddy heat flux anomalies prior to the second event (Fig. 2). Associated with the barotropic structure of the planetary waves in the stratosphere (Fig. 9b) and the meridional structure of the stratospheric zonal mean zonal winds (Fig. 4b), the vertical component of wave activity flux was again limited to the lower stratosphere. The decrease in the heat flux anomaly above 30 hPa (Fig. 2) and the EP flux convergence near 40°N (Fig. 10b) suggests the upward-propagating waves may have dissipated in the midlatitude stratosphere and/or propagated equatorward out of the region.

In the days prior to the SSW, both the WN1 and WN2 geopotential height structures were tilted westward with height through the troposphere, becoming even more tilted in the stratosphere (Fig. 9c). The WN1 geopotential height amplitude peaked at 2000 m near 1 hPa, while the WN2 geopotential height amplitude peaked at 300 m near 10 hPa. The baroclinic WN1 and WN2 geopotential height structures were consistent with the appearance of positive (negative) 10-hPa temperature anomalies collocated with anomalous southerly (northerly) flow supporting the presence of poleward heat flux (Fig. 8c). As implied by the WN1 and WN2 geopotential height structures, the stratosphere was characterized by a baroclinic planetary wave structure in the days leading to the SSW, suggesting upward wave activity flux in the middle and upper stratosphere that was not present prior to the two precursor events. This shift to a baroclinic structure prior to the SSW was a major difference between the precursor events and the SSW. The meridional structure of the zonal mean zonal winds prior to the SSW, in which the strongest winds in the stratosphere were confined to the polar region (Fig. 4c) allowed for wave activity flux to converge into the polar region, where wave transience led to the breakdown of the zonal mean zonal winds and the SSW.

Associated with this baroclinic planetary wave structure throughout the depth of the stratosphere, was a large vertical component of the EP flux above 10 hPa at
FIG. 10. Latitude–height cross sections from 1000 to 1 hPa of the Eliassen–Palm (EP) flux
vectors and the EP flux divergence for the same dates as in Fig. 5. The horizontal arrow scale
for the meridional component of EP flux is shown in the bottom right is $10 \times 10^8 \, \text{m} \, \text{s}^{-2}$. A
vertical arrow of the same length represents the vertical component of EP flux of $10 \times
10^6 \, \text{m}^2 \, \text{Pa} \, \text{s}^{-1}$. No additional scaling has been applied to the EP flux vectors in the
stratosphere. The EP flux divergence (convergence) is scaled by $1/\cos \phi$ in dashed
(solid) black, starting at 20 (−20) m s$^{-1}$ day$^{-1}$ every 20 (−20) m s$^{-1}$ day$^{-1}$. Pressure
levels below 700 hPa and latitudes greater than 80°N are excluded.
~65°N (Fig. 10c). Collocated with the large vertical component of EP flux in the stratosphere was EP flux convergence that extended vertically from 7 to 1 hPa and meridionally from 40° to 78°N with the maximum EP flux convergence centered at ~60°N and ~2 hPa. This broad area of EP flux convergence suggests there was an increase in wave activity density in the upper stratosphere and a deceleration in the zonal mean zonal winds prior to the SSW. The upward component of EP flux, consistent with the positive 100-hPa zonal mean meridional eddy heat flux anomaly shown in Fig. 2 prior to the SSW, suggests upward wave activity flux near the tropopause. However, the vertical component of EP flux was much larger in the midlatitude upper stratosphere (10 hPa and above) than at the tropopause, suggesting the largest wave forcing was in the upper stratosphere (Fig. 10c). This result is consistent with the 45°–75°N zonal mean eddy heat flux anomaly analysis, which revealed that the magnitude of the zonal mean eddy heat flux anomaly from 100 to 10 hPa prior to the SSW simultaneously increased at all levels (Fig. 2). This concurrent appearance of a large positive vertical component of EP flux throughout the depth of the stratosphere, which contrasted the vertical structure of the heat flux anomaly prior the precursor events, suggests the occurrence of wave resonance, meaning the amplifying planetary wave magnitude in the stratosphere was due to nonlinear effects (Albers and Birner 2014). The vertical variation in heat flux anomaly prior to the two precursor events revealed the stratospheric maximum lagged the 100-hPa maximum, such a structure suggests the presence of vertically propagating waves (Albers and Birner 2014). The presence of the large upward component of EP flux and EP flux convergence at 10 hPa and 65°N prior to the SSW and a smaller upward component of the EP flux near the tropopause level also contrasts the two precursor events, which were characterized by a small upward component of EP flux in the stratosphere and a large upward component of EP flux at the tropopause (Fig. 10). This result is consistent with the view that prior to major SSWs, wave breaking occurs at higher altitudes (closer to the upper stratosphere) than wave breaking prior to minor SSWs, or in this case the precursor events, which typically occurs lower in the stratosphere.

b. Upper-stratosphere/lower-mesosphere analysis

While the vertical structure of the heat flux anomaly in the days prior to the two precursor events and major SSW was consistent with the associated warming type, the precursor blocking patterns prior to the two precursor events and major SSW contrasted the studies that showed that blocking often precedes SSWs (e.g., Quiroz 1986; Andrews et al. 1987; O’Neill 2003; Martius et al. 2009). The following analysis of the USLM conditions prior to the major SSW will attempt to elucidate the upper-stratospheric dynamical structures that distinguish the major SSW from the two precursor events.

The two precursor events altered the stratospheric thermal and momentum field prior to the SSW through 1) corresponding changes in the location of the EP flux convergence toward the upper stratosphere, 2) the enhancement of the baroclinic westward tilt in the WN1 and WN2 patterns, 3) the change in the meridional and vertical structure of the zonal mean zonal wind, and 4) the development of positive 10-hPa temperature anomalies at ~90°E, on the eastern flank of the polar vortex. Greer et al. (2013) noted that a trend such as this toward the enhancement of a baroclinic structure in the upper stratosphere and anomalously high temperatures over Siberia (on the eastern flank of the polar vortex) suggests the presence of a USLM disturbance. Figure 11 shows a longitude–pressure cross section of geopotential height and temperature anomalies with respect to the zonal mean at 65°N at 1200 UTC 1 January 2013. There was a +60-K temperature anomaly centered at ~100°E and 5 hPa to the east of negative zonal height anomalies collocated with the strongest zonal geopotential height anomaly gradient.

Figure 12a shows EPV and temperature on the 1500-K isentropic surface (near 2 hPa) at 1200 UTC 1 January 2013 to determine if there was breaking planetary wave in the upper stratosphere. Climatologically, breaking planetary waves are common and often occur at the edge of the polar vortex (e.g., Greer et al. 2013). The comma shape of the largest EPV magnitudes and the reversal of the EPV gradient near 160°E suggests wave breaking occurred and that portions of the high EPV reservoir inside the polar vortex were being irreversibly mixed within the midlatitude surf zone (e.g., McIntyre...
Fig. 12. Analysis of the Northern Hemisphere stratospheric flow at 1200 UTC 1 Jan 2013 of (a) the EPV on the 1500-K (−2 hPa) surface (black, every 800 PVU starting at 2000 PVU) and temperature (shaded according to the color bar) from 5° to 90°N and (b) the 2-hPa geopotential height (black, every 400 m), vertical motion [descent (ascent) solid (dashed) magenta, every 1 (−1) × 10⁻³ hPa s⁻¹ starting at 2 (−2) × 10⁻³ hPa s⁻¹ (the zero line is omitted)], and temperature (shaded according to the color bar) from 20° to 90°N. A nine-point filter was applied to the EPV and vertical motion field 25 times.
and Palmer 1983). As described in Greer et al. (2013), a USLM disturbance is often associated with a breaking planetary wave in the upper stratosphere although not every upper-stratospheric breaking planetary wave is associated with a USLM disturbance. Greer et al. (2013) also stated that USLM disturbances are associated with breaking planetary waves that have a large spatial scale of the detrained EPV (i.e., the tail portion of the comma structure of the EPV reaches low latitudes), similar to the breaking planetary wave seen in Fig. 12a.

The upper-stratospheric analysis presented here thus far suggests that the components necessary for the development of a USLM disturbance existed in the days prior to the January SSW, such that an application of the results of Greer et al. (2013) to this case is as follows. Planetary waves from the troposphere progressed vertically into the stratosphere in association with the two precursor events. Between the second event and the SSW nonlinear processes likely amplified the planetary wave magnitudes in the stratosphere. The large amplitude waves then broke in the upper stratosphere, deposing easterly momentum to the mean flow and decelerating the westerlies. In accordance with the geostrophic adjustment process, a thermally indirect ageostrophic circulation developed on the edge of the stratospheric polar vortex coincident with the deceleration of winds. This ageostrophic circulation was characterized by descent in the upper stratosphere and associated adiabatic warming. Figure 12b shows areas of 2-hPa descent poleward of 45°N between ~35° and ~55°E that exceeded 2 × 10⁻¹ hPa s⁻¹ collocated with locally high upper-stratospheric temperatures. The ascending branch of the ageostrophic circulation was located poleward of 60°N along ~150°E and was collocated with locally lower upper-stratospheric temperatures. The adiabatic warming associated with the descending branch of the circulation was then manifest as a positive temperature anomaly on the eastern flank of the polar vortex, which can be seen in the positive zonal temperature anomaly at 1200 UTC 1 January 2013 in Fig. 11 and at 10 hPa 24 h prior to the SSW at 1800 UTC 5 January 2013 in Fig. 8c. As Greer et al. (2013) suggest, this indirect ageostrophic circulation can move an environment into a more baroclinic state that, in this case, was manifest as an increase in the westward tilt in the WN1 and WN2 pattern above 10 hPa. This westward tilt in the WN1 and WN2 patterns then favored the continued upward progression of planetary waves in the stratosphere (e.g., Greer et al. 2013), which were focused toward the polar region due to the preconditioning of the stratosphere by the precursor events (Fig. 4c).

During the 5-day period prior to the SSW, the Fourier decomposition showed a westward tilt in the WN1 and WN2 height field through the depth of the troposphere and stratosphere (Fig. 9c). This stratospheric westward tilt of the WN1 and WN2 pattern was in contrast to the barotropic vertical structure of the WN1 and WN2 magnitudes in the stratosphere prior to the precursor events (Figs. 9a,b). Therefore, it is suggested that the USLM disturbance developed around 1 January and matured by 4 January 2013, increasing the stratospheric baroclinicity through the associated ageostrophic indirect circulation. The stratospheric baroclinic structure of the WN1 and WN2 patterns created a favorable environment for the upward progression of planetary waves in the stratosphere and the rapid deceleration of the zonal mean zonal wind associated with the SSW.

6. Summary and future work

Three distinct decelerations of the 65°N 10-hPa zonal mean zonal wind during the 2012/13 boreal winter were examined using the National Centers for Environmental Prediction Climate Forecast System Reanalysis dataset. The first deceleration resulted in a minimum zonal mean zonal wind of 8 m s⁻¹ on 5 December 2012 and was termed the first event. The second deceleration resulted in a minimum zonal mean zonal wind of 5 m s⁻¹ on 26 December 2012 and was termed the second event. These events did not meet the definitions of a “major” or “minor” sudden stratospheric warming (SSW). The final rapid deceleration occurred on 6 January 2013 when the 10-hPa zonal mean zonal wind at 65°N reversed from westerly to easterly, resulting in a major SSW, termed the SSW. The state of the atmosphere prior to each of the three events was analyzed to highlight the characteristic differences between the deceleration events in terms of troposphere and stratosphere precursor conditions and to assess the role of the precursor events in preconditioning the stratosphere for the SSW. The blocking analysis averaged over the 5-day period prior to the two precursor events and the SSW showed that the two precursor events were preceded by high-latitude tropospheric blocks, while the SSW was not immediately preceded by any high-latitude blocking. Although there were no identified high-latitude blocks prior to the SSW, the planetary wavenumber-1 and wavenumber-2 structures transitioned from barotropic to baroclinic prior to the SSW.

In association with an enhancement of the baroclinic planetary wave structure, an analysis of the upper stratosphere suggested the development of an upper-stratospheric/lower-mesospheric (USLM) disturbance after the second event and prior to the SSW. The simultaneous occurrence of the USLM disturbance and the westward tilt in the planetary wave structure
throughout the troposphere and stratosphere created an environment favorable for the vertical propagation of planetary waves into the upper stratosphere and the convergence of wave activity flux in the USLM. This increased wave activity density rapidly decreased the zonal mean zonal wind at 65°N and 10 hPa.

The analysis of the two precursor events and the SSW suggests that the upward wave activity flux that preceded the two precursor events contributed to a large 40-day-average 100-hPa 45°–75°N anomalous zonal mean eddy heat flux of 11.7 K m s⁻¹. Polvani and Waugh (2004) showed that weak (strong) vortex events, as defined in Baldwin and Dunkerton (2001) as the negative (positive) northern annular mode, are preceded by anomalous positive (negative) 40-day-average zonal mean eddy heat flux anomaly. Comparing this 40-day-average heat flux anomaly of 11.7 K m s⁻¹ to the Polvani and Waugh (2004) climatology shows that the 40-day-average heat flux anomaly that preceded the January 2013 SSW was one of the largest in the Polvani and Waugh (2004) climatology. This large 40-day-average zonal mean eddy heat flux anomaly suggests there was strong forcing from the troposphere into the stratosphere in the 40 days prior to the January 2013 SSW. The largest 100-hPa daily zonal mean eddy heat flux anomalies within those 40 days preceded the precursor events, suggesting that the blocking events that preceded the events contributed to the anomalously positive 40-day-average heat flux. The precursor events are also suggested to have preconditioned the stratosphere for the SSW by altering the meridional and vertical structure of the zonal mean zonal wind. After the precursor events and before the SSW the region of strongest zonal mean zonal winds in the stratosphere was shifted poleward, allowing for wave activity flux to be directed into the polar region where it converged and further decelerated the zonal mean zonal winds. Finally, the second event generated positive geopotential height anomalies that remained in the stratosphere and amplified associated with nonlinear affects, which were manifest as the simultaneous occurrence of positive heat flux anomalies through the depth of the stratosphere. These large amplitude waves subsequently broke in the upper stratosphere. These results suggest that blocking events are not only important in the days prior to SSWs, but also in the weeks prior to SSWs. The upward wave activity flux into the stratosphere associated with tropospheric blocking can both precondition the stratospheric structure into a favorable configuration for an SSW to occur and act as an influx of wave activity into the stratosphere that can amplify due to wave resonance.

The analysis presented here complemented the recent studies of the January 2013 SSW by Vargin and Medvedeva (2015) and Coy and Pawson (2015) and presented the importance of two precursor events in preconditioning the stratosphere in the 40 days prior to the SSW. The detailed synoptic and dynamic analysis of the precursor events and their associated predecessor tropospheric and stratospheric states in this study provided further analysis of the increased wave activity flux noted by Vargin and Medvedeva (2015) at the end of November and middle of December. Coy and Pawson (2015) also noted the importance of the tropospheric ridging that occurred in mid-December on the stratosphere prior to the SSW and specifically emphasized the importance of accurately forecasting the precursor tropospheric state prior to SSWs in order to accurately forecast SSWs.

Additional analysis is required to elucidate other potential contributions to this SSW. Specifically, would it be possible to assess the contributions of individual synoptic- and planetary-scale features? For example, a strong Madden–Julian oscillation (MJO) event was propagating eastward into the Pacific Ocean just prior to and during the second event. Did the MJO affect the structure of the jet over Eurasia and the Pacific and contribute to the planetary-scale wave activity that led to the second event (e.g., Garfinkel et al. 2012; Liu et al. 2014)? Analyzing these features from both a transformed Eulerian mean and synoptic perspective could possibly lead to a better understanding of both the zonal mean and zonally varying precursors to SSWs.

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