Modulating Effects of Planetary Wave 3 on a Stratospheric Sudden Warming Event in 2005

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ABSTRACT

The Eliassen–Palm flux (EPF) and Plumb’s wave activity flux (WAF) were computed, using ERA-Interim data, to analyze the influence of planetary wave 3 on a stratospheric sudden warming event from 17 February to 15 March 2005 (SSW05). It was found that 1) SSW05 consisted of three stages: a prior minor warming (MnW05), a late final warming (FW05), and a warming stagnation between MnW05 and FW05; 2) the wave 3 first decreased total upward EPFs by more than 30% at 100 hPa, resulting in the warming stagnation, and then increased upward EPFs by greater than 50%, leading to FW05; and 3) the anomalies of wave-3 activity fluxes were associated with the pattern of Atlantic blocking high in the latter two stages. The interactions between the wave 3 and wave 1 partitioned the zonal upward channel of total wave activity fluxes from one longitudinal region into two longitudinal regions and affected SSW05.

1. Introduction

A stratospheric sudden warming (SSW) is a typical manifestation of troposphere–stratosphere interaction in winter (e.g., Charney and Drazin 1961; Charlton and Polvani 2007). Since the first dynamical model of SSW was established by Matsuno (1971), subsequent studies have suggested that the process of SSW is closely related to the upward propagation of planetary waves (PWs), especially waves 1 and 2, from the troposphere to the stratosphere (e.g., Nishii et al. 2011; Solomon 2014). In boreal winter, owing to the land–sea thermal contrast and large topographic forcing, PWs originating from the troposphere can propagate vertically under conditions of suitable westerly flow (Charney and Drazin 1961; Andrews et al. 1987). According to this theory, PWs, waves 1 and 2, can penetrate into the stratosphere whereas smaller-scale waves are generally trapped near the tropopause. Hence, wave 3 propagating into the lower stratosphere is less common (Lu and Ding 2013).

PWs 1 and 2 can reach a higher level, resulting in vortex-splitting or vortex-displacement features. Usually, vortex-splitting SSWs are dynamically associated with an amplified wave 2, while vortex-displacement SSWs are controlled by wave 1 (Charlton and Polvani 2007). It is rare for there to be a major SSW event with wave 3 as the precursor (Bancalá et al. 2012). Indeed, the literature indicates that wave-1 SSWs (e.g., Mukougawa and Hirooka 2004; Hirooka et al. 2007; Vargin 2015) and wave-2 SSWs (e.g., Palmer 1981; Harada et al. 2010) form the majority of cases, while wave-3 SSWs are rarely recorded.

That said, a number of studies have in fact reported wave-3 signals in the troposphere during SSWs, but without revealing the mechanism of wave-3 bottom-up influence from the troposphere to the stratosphere. For example, Limpasuvan et al. (2004) indicated momentum-flux anomalies associated with wave 3 near the tropopause in some SSWs, while Song and...
Martius et al. (2009; Colucci and Kelleher 2015) found major SSWs (Martius et al. 2009; Bancalá 2015) and usually leads to development of a wave-1 major SSW (Martius et al. 2009; Bancalá et al. 2012; Ayarzagüena et al. 2015). Nonetheless, there is no significant statistical connection between SSWs and BHs (Taguchi 2008; Woollings et al. 2010). Some BHs occur prior to SSWs, while a large number of BHs are followed by SSWs or a stratospheric zonal-mean-flow anomaly. Limpasuvan et al. (2004) suggested the ABH weakens obviously, or even vanishes, during the SSW, if no wave 3 is taken into consideration in the model.

A major SSW occurs when the mean temperature at 10 hPa or below increases poleward from 60°N and the zonal-mean westerly winds become easterly during winter (Andrews et al. 1987). If the westerly winds only decelerate and not reverse to easterly during the significant warming (i.e., at least 25 K in a period of week or less), it is called a minor SSW. Stratospheric final warming (SFW) refers to a special type of major SSW when zonal-mean westerly flow shifts to easterly flow, which then lasts for at least 10 consecutive days (Charlton and Polvani 2007). SFW usually occurs in April. However, during typical major SSW, the westerly flow can recover from the easterly flow in a few days. Nonetheless, the background circulation and stratospheric wave activity are similar during SFW and SSW (Black and Medaniel 2007). During SFW, anomalies of tropospheric circulation shift more poleward, and waves 1 and 2 are dominant (Black and Medaniel 2007; Hu et al. 2014, 2015). Meanwhile, SFWs are mainly dominated by wave 1 in March but by wave 2 in April (Ayarzagüena and Serrano 2009).

According to the datasets used and the pressure levels selected, SFWs and major SSWs can sometimes be confused. For example, the warming events that took place in March 1988 and February 1989 were mentioned as SFWs in the statistics of North Pole temperature by Labitzke et al. (2002) but reported as major SSWs by Charlton and Polvani (2007), and the SFW that occurred in March 2005 (Singleton et al. 2007; Manson et al. 2008) was classified as an SSW by Harada et al. (2010). These SFWs occurred during early seasonal transition and featured similar background circulation and stratospheric wave activity as major SSW events in winter (Hu et al. 2014, 2015).

The stratospheric warming in March 2005, encompassing two events from 17 February to 15 March 2005 (collectively referred to as SSW05) and dominated by waves 3 and 1 [the only wave-3 event in Table 1 of Harada et al. (2010)], provides us with a precious opportunity to advance our understanding of the influence of wave 3, which is normally trapped near the tropopause, on stratospheric warming. In particular, we focus on the modulation effects of wave 3 on wave-1–2 activity fluxes in the stratosphere during SSW05.

### 2. Data and methods

The data employed in this study were from the ERA-Interim global daily dataset (horizontal resolution: 1.5° × 1.5°) produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) (Dee et al. 2011). The Eliassen–Palm flux (EPF) is a standard diagnostic method for analysis of wave propagation in a two-dimensional meridional plane (Edmon et al. 1980; Andrews et al. 1987). The quasi-geotropic (QG) EPF in vector form can be written as

\[
\begin{align*}
F_{(\phi)} &= -a \cos \phi \overline{u'v'} \\
F_{(\rho)} &= f a \cos \phi \frac{\overline{\theta'v'}}{\overline{\theta'}}
\end{align*}
\]  

The EPF is the equivalent to the wave activity flux and represents the group velocity of linear QG waves in the meridional–pressure plane. The horizontal and vertical components, \(F_{(\phi)}\) and \(F_{(\rho)}\), are proportional to “eddy momentum flux” and “eddy heat flux,” respectively.

The divergence of EPF is defined by

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

The effect of the eddies on the zonal-mean state is expressed by

\[
\frac{\partial \overline{u}}{\partial t} - f \overline{u'} - \varepsilon = (a \cos \phi)^{-1} \nabla \cdot \mathbf{F}.
\]
Plumb (1985) defined the wave activity flux (WAF) to indicate the propagation of PW packets in three-dimensional space:

$$\text{WAF} = \frac{p}{p_0} \cos \phi$$

where the standard reference pressure $p_0 = 1000 \text{ hPa}$, $S = \delta T / \delta z + \kappa T H$ is the static stability and $\kappa \approx 0.286$, $z = -H \ln(p/p_0)$ is a log $p$ vertical coordinate with scale height $H = 8 \text{ km}$, and $u_g$ and $v_g$ in Eq. (4) are geostrophic winds:

$$\begin{align*}
\mathbf{u}_g &= -\frac{1}{fa} \frac{\partial \Phi}{\partial \phi} , \\
\mathbf{v}_g &= -\frac{1}{fa \cos \phi} \frac{\partial \Phi}{\partial \lambda}.
\end{align*}$$

(5)

The regional blocking index (Tibaldi and Molteni 1990), the southern 500-hPa geopotential height gradient (GHGS), and the northern one (GHGN) were calculated at each longitude on each day:

$$\begin{align*}
\text{GHGS} &= \frac{Z(\phi_n) - Z(\phi_s)}{\phi_0 - \phi_s} \quad \text{and} \\
\text{GHGN} &= \frac{Z(\phi_n) - Z(\phi_s)}{\phi_n - \phi_0}.
\end{align*}$$

(6)

where $\phi_n = 80^\circ \text{ N} + \Delta$, $\phi_0 = 60^\circ \text{ N} + \Delta$, $\phi_s = 40^\circ \text{ N} + \Delta$, and $\Delta = -5^\circ, 0^\circ, 5^\circ$. GHGS can indicate the regional blocking strength if both of the following conditions are satisfied for at least one value of $\Delta$:

$\text{GHGS} > 0$ and $\text{GHGN} < -10 \text{ m}(1^\circ \text{ lat})^{-1}$.

In Eqs. (1)–(6), $u$ and $v$ are zonal wind and meridional wind; $\Omega$, $a$, $f$, $\Phi$, $Z$, $\phi$, and $\lambda$ represent Earth’s rotation rate, Earth’s radius, the Coriolis parameter, the geopotential, the geopotential height, the latitude, and the longitude, respectively; $\theta_p$ is the partial derivative of potential temperature $\theta$ with respect to $p$; and $u^\phi$ and $v^\phi$ are the meridional velocity of “residual meridional circulation” and the Eulerian-mean friction, respectively. Overbars and primes denote zonal means and deviations, respectively. Hereafter, the vertical components of EPF and WAF are abbreviated as EPFz and WAFz, respectively.

3. Results

a. Structural evolution of temperature and zonal wind in SSW05

A special SSW event (SSW05) occurred near the North Pole during 17 February–15 March 2005, in which there were a minor warming stage (MnW05) during 17–24 February 2005, a stratospheric final warming stage (FW05) during 7–15 March 2005 [according to the definition for SSW (Andrews et al. 1987; Charlton and Polvani 2007)], and a warming stagnation stage between the two warming stages (Figs. 1a and 1b). The warming area in the latter FW05 episode was relatively wider and extended more equatorward from the North Pole (to near $55^\circ \text{ N}$) than the first stage (MnW05). The rapid decrease of the stratospheric zonal-mean zonal wind in the high latitudes occurred after 17 February. However, the zonal winds sped up from 1 March. After 7 March, the zonal wind again decreased rapidly in high latitudes. After 12 March, the westerly flow reversed to an easterly flow, spreading from the North Pole to $50^\circ \text{ N}$.

As shown in Fig. 1b, during MnW05, temperature in the upper stratosphere increased rapidly, and 220-K contours extended as low as 20 hPa. Between MnW05 and FW05, 220-K contours hovered around 20 hPa and the westerly recovered somewhat in the middle stratosphere, resulting in a stage of warming stagnation. During FW05, an abrupt temperature increase occurred in the middle stratosphere, the 220-K contours spread to the lower stratosphere, and, correspondingly, the zonal wind shifted rapidly from westerly to easterly throughout the stratosphere.

b. Roles of waves 1–3 and the ABH in SSW05

Wave 3 played an important role in SSW05, which represents a departure from SSWs traditionally being dominated by waves 1 and 2 [Table 1 in Harada et al. (2010)]. The $50^\circ$–$70^\circ \text{ N}$ mean EPFz at 100 hPa is a typical choice for analyzing the relationship between PWs and SSWs (e.g., Polvani and Waugh 2004; Kodera et al. 2013). The extreme value of wave-3 EPFz was closely related to the ABH (Fig. 1d). Wave-3 EPFz reached its minimum in the mature stage of the strong ABH [blocking strength larger than $10 \text{ m}(1^\circ \text{ lat})^{-1}$] located near the $0^\circ$ line of longitude. When the ABH again reached the moderate strength of $5 \text{ m}(1^\circ \text{ lat})^{-1}$ and
FIG. 1. Event evolution from 1 Feb to 15 Apr 2005: (a) zonal-mean temperature (shading; K) and zonal wind (contours; m s\(^{-1}\)) at 10 hPa; (b) zonal-mean temperature (shading; K) and zonal wind (contours; m s\(^{-1}\)) averaged over 60°–90°N; (c) EPFz of waves 1–3 (curves; 10\(^6\) m\(^2\) s\(^{-2}\)) and the wave-3 contribution ratio (bars; %) averaged over 50°–70°N at 100 hPa; (d) regional blocking index (GHGS) at 500 hPa. The wave-3 contribution ratio represents its portion of the sum of each absolute EPFz. EPFz is normalized by 1000 hPa. The vertical strips mark roughly the beginning of the three stages, based on a combination of temperature variation and wind variation from (a) and (b) and wave-3 contribution from (c).
moved westward, wave-3 EPFz reached its maximum. Between the two extreme values, the ABH was relatively weak [blocking strength mostly less than 5 m (1° lat)^{-1}]. Usually, the Europe trough–Atlantic ridge system is the weakest among the three trough–ridge patterns in boreal winter troposphere, implying that wave 3 is weak. Hence ABH is a manifestation of wave 3 in general. The variation of 100-hPa wave-3 EPFz was not in strict accordance with that of the 500-hPa ABH index (Figs. 1c and 1d) because of the different pressure levels selected in the baroclinic middle latitudes. The wave-3 EPFz contribution ratio, the percentage of wave-3 EPF in the sum of each absolute EPFz, was as high as 35%–55% (reaching its extreme value) at the beginning of the latter two stages of SSW05 (Fig. 1c).

Figure 1c shows the time dependence of the 100-hPa EPFz for waves 1, 2, and 3 during 1 February–13 April 2005. During the whole process of SSW05, the wave-1 EPFz was relatively important and positive (i.e., upward). However, after 24 February, the wave-3 EPFz decreased rapidly to be negative (indicating downward wave activity flux) and had reduced to $-1.4 \times 10^6$ m$^2$s$^{-2}$ on 28 February. From 26 February to 1 March, wave-3 flux was negative, and the wave-3 EPFz contribution ratio generally exceeded 30%. EPFz (or its proxy, eddy heat flux) at 100 hPa in general reaches its maximum (or minimum) value prior to the stratospheric response [e.g., Fig. 4 in Polvani and Waugh (2004) and Fig. 9a in Kodera et al. (2016)]. The anomalous downward wave-3 EPFz acted to reduce the total upward EPFz at 100 hPa, and thus the stratospheric warming stagnated, as shown in Fig. 1b. Consequently, this warming only reached minor warming strength. After that point, the wave-3 EPFz shifted to be positive (i.e., upward). Around 12 March, the contribution ratio of the wave-3 EPFz exceeded 50% (2.2 $\times 10^6$ m$^2$s$^{-2}$) and surpassed that of the wave-1 EPFz, which led to a dramatic increase in the total upward wave activity flux and triggered the onset of FW05. As a result, the stratospheric zonal wind reversed to an easterly (Figs. 1a,b). After FW05, the wave-3 EPFz decreased rapidly prior to wave-1 EPFz declining. Therefore, SSW05 was dominated by wave 3 and wave 1. Wave 3 first suppressed the development of the early stage (MnW05) and then enhanced that of FW05.

The vertical propagation of the wave 3 can also be revealed by contrasting the EPFz of waves 1–2 only with that of waves 1–3 (Fig. 2). Wave 3 played a small role in the stratosphere at the beginning of MnW05 but had an effect on the lower stratosphere in the latter two stages. Early on in MnW05 (17–19 February), the EPFs of waves 1 and 2 and those of waves 1–3 were similar (Figs. 2a and 2d), indicating that wave 3 could be ignored in the stratosphere (Fig. 2g; also as in Fig. 1c). The MnW05 EPFs and EPF convergence in the upper stratosphere were stronger than those in the latter two stages because of waves propagating upward along the favorable westerly jet axis (strong westerly winds but weaker than Rossby critical velocity: about 28 m s$^{-1}$ for wave 1, 16 m s$^{-1}$ for wave 2, and 9 m s$^{-1}$ for wave 3 at 60°N; Andrews et al. 1987). In the early warming stagnation (26–28 February), the westerly winds decelerated in the extratropical stratosphere and the upward fluxes of waves 1–3 were weaker than those of waves 1 and 2 in the lower stratosphere. Referring to the regions covered by $-5$ m s$^{-1}$ day$^{-1}$ contours, the EPF convergence regions extended to middle latitudes (near 45°N) from the polar region in the middle and lower stratosphere (10–70 hPa; Fig. 2b), while they occupied only part of the polar region in the middle and lower stratosphere (10–50 hPa; Fig. 2e). The downward wave-3 flux and its divergence in the extratropical lower stratosphere, shown in Fig. 2h, reduced the total upward wave activity fluxes (Fig. 2e) and suppressed the deceleration of the polar night jet [contributing about 10 m s$^{-1}$ day$^{-1}$ to increasing the zonal winds, according to Eq. (3)], which hindered the development of MnW05. However, during 11–13 March, upward fluxes of waves 1–3 and their convergence was stronger than those of waves 1 and 2 in the high-latitude lower stratosphere (20–70 hPa; Figs. 2c and 2f). There was a window (zonal-mean zonal wind less than the Rossby critical velocity for wave 3) in which the upward wave-3 flux and its convergence occupied the extratropical lower stratosphere (20–70 hPa; Fig. 2i). Wave 3 strengthened the total wave activity fluxes and decelerated the polar night jet (by about $-5$ m s$^{-1}$ day$^{-1}$), opposite to the situation during 26–28 February. Thus, the warming developed downward (the 220-K contours spread from the middle stratosphere to the lower stratosphere; Fig. 1b).

c. Modulating effect of wave 3 on waves 1 and 2

According to the essential difference between EP Flux and Plumb WAF, although the 2D EPF can calculate individual wave activities, which can then be accumulated, it is in a zonal-mean sense and does not involve the different wave interactions locally (in the longitudinal direction). However, the 3D WAF involves the different wave interactions whose wave fluxes are complex and can no longer be calculated separately and then summed. Waves 1–3 are usually dominant in the stratosphere, and the additional waves contributed little in terms of 3D WAF in this case. Hence, the modulating effect of wave 3 can be investigated by contrasting waves 1 and 2 only with waves-1–3 WAF in 50°–70°N, the primary latitudinal band for the upward propagation of PWs (Harada et al. 2010; Kodera et al. 2013). The
FIG. 2. Latitude–pressure cross sections of 3-day-averaged zonal-mean zonal winds (contours; m s$^{-1}$), EPF [vectors of Eq. (2); >9 m s$^{-2}$ only], and EPF divergence [right-hand side of Eq. (3); shading; m s$^{-1}$ day$^{-1}$] of waves 1 and 2 in (a) early MnW05, (b) the early warming stagnation period, and (c) early FW05. (d)–(f) As in (a)–(c), respectively, but for waves 1–3. (g)–(i) As in (a)–(c), respectively, but for wave 3 only. The EPF components are normalized by $\pi \times 6378 \text{ km}$ horizontally and by 1000 hPa vertically.
geopotential height anomaly in the stratosphere in Fig. 3 and the EPFz in Fig. 1c indicated that wave 1 was dominant most of the time, compared to wave 2. The upward wave-1 and -2 fluxes were basically channeled over one location (left column in Fig. 3), first in the Eastern Hemisphere (Figs. 3a,c) and later in the Western Hemisphere (Figs. 3e,g). Wave 3 tended to modulate the upward channel of total wave activity fluxes from one longitudinal region into two longitudinal regions (right column in Fig. 3). From the late MnW05 to the early warming stagnation period (23–28 February; Figs. 3b,d), the ABH (solid contours near 30°W) developed well (blocking strength larger than 10 m (1°lat)^{-1}; Fig. 1d) and the WAF structures changed significantly with wave 3 added. The upward wave-1 and -2 fluxes occupied a wide area mainly from Eurasia to...
FIG. 4. Longitude–pressure cross sections of the 3-day 50°–70°N mean zonal anomaly of geopotential height (contours; m) and temperature (shading; K) during (a) 23–25 Feb, (b) 26–28 Feb, (c) 8–10 Mar, and (d) 11–13 Mar.
the western Pacific (Figs. 3a,c). However, the upward wave-1–3 fluxes were mainly focused within two narrow channels located in western Europe (east of the ABH) and the central North Pacific, respectively (Figs. 3b,d). Meanwhile, a downward channel (blue shading) existed in the west of the ABH (around 60°W), consistent with the wave-3 downward EPFz in the lower stratosphere (Fig. 2h) and causing the warming stagnation. From the early FW05 to its mature period (8–13 March; Figs. 3e and 3g), the ABH reached a moderate strength of 5 m (1°lat)−1 and moved westward (also in Fig. 1d), which resulted in a westward-tilting ridge and the upward wave-1 and -2 fluxes mainly gathered over the Labrador Peninsula of the Atlantic west coast. With wave 3 included, the upward channel moved eastward over the Atlantic slightly, and a new upward channel established over the East Asia–Pacific area (90°E–150°W; Figs. 3f, h), increasing the upward EPFz (Fig. 2i) and promoting the occurrence of FW05. According to the complex wave fluxes by wave interactions in 3D WAF, interactions between the dominant wave 3 and wave 1 were responsible for the two upward channels of wave activity in these two stages.

In the above process, wave activity flux near the Atlantic region (60°W–30°E) was sensitive to wave 3, which was closely related to the evolution of the ABH (Fig. 1d) and the nearby structures of the temperature and pressure fields (Fig. 4). SSWs are generally accompanied by BHs (e.g., Martius et al. 2009; Colucci and Kelleher 2015). During 23–28 February (Figs. 4a,b), the anomalous high over the ABH region extended vertically up to 30 hPa, with relatively low and cold air straddling both sides on the top. This structure resulted in westward-tilting troughs of pressure (height) and temperature with altitude over the eastern portion of the ABH, and the temperature trough trailed behind the pressure trough, which contributed to upward wave activity flux (Harada et al. 2010; Kodera et al. 2013). On the contrary, over the west of the ABH, pressure and temperature troughs tilted eastward vertically, with the temperature trough leading. This situation promoted downward wave activity flux. During 8–13 March (Figs. 4c,d), the ABH was at a moderate intensity level and moved westward (also seen in Fig. 1d), merging with the stratospheric Alaska high and forming a ridge tilted westward with altitude. The temperature trough fell behind the pressure trough to the east of the ridge, which benefited upward wave activity flux. The ABH induced the abnormal wave-3 flux that was linked with polar night jet and SSW (Figs. 1d, 1c, and 1b).

Wave 3 primarily modulated the horizontal direction of planetary wave activity during SSW05. The WAFs of waves 1–3 (Figs. 5b,d) were clearly larger than those of waves 1 and 2 (Figs. 5a,c) at 250 hPa near the tropopause. In the early warming stagnation period (26–28 February), the fluxes of waves 1–3 originated from
the west coast of the Atlantic (80°–40°W) and, passing through the Atlantic meridional ridge, shifted from poleward to equatorward direction (Fig. 5b). Compared with waves 1 and 2 (Fig. 5a), wave 3 weakened the poleward wave activity that can enter upward into the stratosphere (Chen and Huang 2002). Subsequently, the warming stagnated. On the contrary, in the stage of FW05 occurrence (11–13 March), the poleward wave-1–3 fluxes formed in North America (120°–90°W) crossed the Atlantic northwest–southeast ridge (Fig. 5d). In contrast with waves 1 and 2 (Fig. 5c), wave 3 enhanced the poleward wave activity, which produced favorable conditions for upward wave fluxes into the stratosphere (Andrews et al. 1987; Kodera et al. 2013). This process was conducive to the warming developing again, prompting the occurrence of FW05.

4. Conclusions and discussion

This study reveals that there were different modulating effects of wave 3 on an SSW in 2005, especially during the warming stagnation period and the latter final warming (FW05) stage. Being modulated by wave 3 in association with the ABH, the total upward wave activity flux was reduced early on by the anomalous downward channel near the western portion of the deep upright structure of the strong ABH, while later on it was enhanced by the anomalous upward channel near the eastern portion of the moderate ABH located west of its previous position. Therefore, the total EPFz decreasing by more than 30% (or increasing by more than 50%) resulted in the warming stagnation (or FW05). This represents a departure from previous reports that the ABH during SSWs mostly strengthens the upward propagation of PWs (e.g., Nishii et al. 2011; Kodera et al. 2013; Ayarzagüena et al. 2015). Nevertheless, BHs might be a necessary but not sufficient condition for SSWs. A large number of BHs are followed and influenced by SSW or stratospheric zonal mean flow (e.g., Taguchi 2008; Woollings et al. 2010).

Wave 3 was found to regulate the atmospheric background fields near the tropopause by weakening or strengthening the poleward wave activity flux, which hampered or enhanced the upward wave activity flux for the stratospheric warming in different stages. Furthermore, the interactions between the dominant wave 3 and wave 1 partitioned the zonal upward total wave fluxes from one channel into two channels. Woollings et al. (2010) mentioned similar wave-3 modulations in the Southern Hemisphere; the wave-3 anomaly vanishes at the onset of blocking and is replaced by an anomalous wave 1 and 2.

Aside from PWs, small-scale transient waves may also be important for SSWs. Coy and Pawson (2015) observed strong transient wave activity resulting in upward WAF in the middle stratosphere before an SSW in January 2013. Therefore, to thoroughly understand the dynamical mechanisms involved in SSW, the interactions between transient waves and PWs also deserve more attention (Kushner and Polvani 2005; Wüst and Bittner 2011).

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