Gravity Wave Variation from the Troposphere to the Lower Thermosphere during a Stratospheric Sudden Warming Event: A Case Study

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Abstract

High resolution Whole Atmosphere Community Climate Model (WACCM) simulations are used to study how gravity waves vary during a stratospheric sudden warming (SSW) event from the source region to the lower thermosphere. The variation of zonal mean momentum flux of resolved gravity waves (with zonal wavelengths less than 1600 km) during SSW is qualitatively consistent with those obtained from parameterized studies, mainly caused by the change of filtering by the mean zonal wind. At high latitude in the winter hemisphere, stratospheric and mesospheric momentum fluxes vary rapidly during SSW, and their magnitudes decrease significantly following SSW, agreeing with satellite observations. Gravity waves are also found to vary as wave sources change. At tropical regions (especially in the summer hemisphere), convectively generated gravity waves increase due to enhanced deep convection following SSW. At higher latitudes, orographic waves vary during SSW as the wind changes extend from the stratosphere down to the troposphere. Gravity waves generated from adjustment of the polar jet also undergo significant changes during SSW. These changes lead to strong longitudinal variation of gravity waves.

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1. Introduction

Stratospheric sudden warming (SSW) provides an exemplary case to study the coupling between the lower atmosphere and the upper atmosphere (mesosphere, thermosphere and ionosphere). The slowdown and reversal of winter stratospheric wind changes the gravity wave (GW) propagation condition, which in turn changes the circulation in the upper atmosphere (e.g. Liu and Roble 2002; Tan et al. 2012; Yiğit and Medvedev 2012; Yuan et al. 2012; Miyoshi et al. 2015). The SSW impact can also extend into the summer mesosphere and mesopause region, as evidenced in observations and numerical simulations, through inter-hemispheric coupling (Becker and Fritts 2006; Karlsson et al. 2007, 2009a, b; Körnich and Becker 2010; Tan et al. 2012; Miyoshi et al. 2015). Numerical simulations by Liu et al. (2013, 2014a) and Miyoshi et al. (2015) suggest that the SSW impact may extend into the upper thermosphere, though observations of thermospheric responses to SSW are less conclusive (Liu et al. 2011; Fuller-Rowell et al. 2011).

These studies demonstrate GW propagation and breaking play a key role in determining the changes in the upper atmosphere. It is less clear, however, if the GW sources change during SSW, since in most of these studies (with the exceptions of Becker and Fritts (2006); Miyoshi et al. (2015)) GW sources, propagation and their effect on the large-scale flow are parameterized. There have been previous studies suggesting processes that are closely associated with GW excitation, such as tropical convection, vary during SSW (Kodera and Yamada 2004; Kodera 2006; Gómez-Escolar et al. 2014; Evan et al. 2015), though the direct impact on the GWs has yet to be explored. Since GWs can be excited by imbalance of jet flow (O’Sullivan and Dunkerton 1995; Zhang 2004), the rather rapid change of winter jet system is expected to be a source of GW variability during SSW. Analysis of ECMWF-T799 (with horizontal resolution of ~0.75°) simulation by Yamashita et al. (2010) indeed reveals large GW variability in the stratosphere closely associated with the jet and the modulation by planetary waves around the time of the 2009 SSW. It is challenging, however, to unambiguously determine if the variability is related to excitation by jet imbalance or by changes in the propagation and filtering of GWs. Furthermore, it is questionable how well GW parameterization schemes, which usually ignore GW lateral propagation and may over-simplify the breaking process, represent the GW forcing during SSW.

In this study, we explore these questions by analyzing the GW momentum flux, its divergence, and vertical energy flux during a SSW event simulated by a high-resolution version of the NCAR Whole Atmosphere Community Climate Model (WACCM). With a quasi-uniform horizontal resolution of ~25 km and vertical resolution of 0.1 scale-height, the model can resolve waves with horizontal wavelengths down to ~200 km without excessive numerical damping (Liu et al. 2014b). WACCM model and simulations used for this study and the analysis method are described in Section 2. Analysis results are presented in Section 3, and summary in Section 4.

2. Numerical model and analysis method

WACCM used in the current study is the same as that described in Liu et al. (2014b): it employs the spectral element (SE) dynamical core based on a cubed-sphere with a quasi-uniform horizontal resolution of ~25 km (NE120NP4). The vertical resolution is 0.1 scale-height above 40hPa and the model has 209 levels (L209), extending to 5.9 × 10^6 hPa (~145 km). GW parameterization has been turned off in the simulation to better evaluate the resolved GWs. Subgrid-scale deep convection is parameterized according to Zhang and McFarlane (1995). 18 model-months of WACCM simulations have been performed. Two major SSW cases are identified in the simulations, one in February and one in November-December, and the former is analyzed in this study. The GW potential energy density and absolute momentum flux deduced from the simulation are in general agreement with satellite observations (Liu et al. 2014b; Liu 2016), though the zonal wind reversal produced by the resolved GW occurs at altitudes higher than climatology.

One dimensional wavelet method is applied in the zonal direction to quantify GWs and their spatial variation, and five wavelet bands are used: 0–100 km, 100–200 km, 200–400 km, 400–800 km, and 800–1600 km. The vertical fluxes of the zonal momentum are calculated by averaging the wavelet cospectra of horizontal and vertical winds over the respective zonal wavelengths (approximately 100 km, 200 km, 400 km, 800 km and 1600 km). The vertical energy flux due to pressure work for each band is calculated in a similar way by averaging the cospectra of pressure and vertical wind over the corresponding zonal wavelength.

The total fluxes are obtained by summing over all the bands. This study mainly concerns the total fluxes, but we note that the fluxes in 0–100 km and 100–200 km are generally small since they are underestimated due to limited spatial resolution.
3. Analysis

A major SSW event is identified in February, with the zonal mean zonal wind at 60°N and 10 hPa reversing to westward on 17 February (Fig. 1a). This is a vortex displacement warming event. The zonal wind above 1 hPa returns eastward after 5–10 days, while it remains westward for much longer in most of the stratosphere and around the tropopause. It is worth noting that the zonal wind in the troposphere slows down considerably in the second half of February, following the SSW. The main features of the zonal mean zonal wind and temperature change during SSW (Fig. 1), including the development of the elevated stratopause, changes in the mesosphere and lower thermosphere and inter-hemispheric coupling patterns (downward shift of eastward acceleration near 80 km at mid to high southern latitudes, cooling at similar location while warming above), are similar to those obtained in previous studies (mostly using parameterized GW). As discussed in Liu et al. (2014b), the resolved gravity waves in the model are not sufficiently strong and their breaking altitudes are too high. Therefore, the mesopause and stratopause heights are higher than climatology.

Zonal mean forcing by large-scale waves (zonal wavenumber 1–6) and GW forcing (GWF) (zonal wavelength scales between 0–1600 km obtained from wavelet analysis) are shown in Fig. 2. Strong westward forcing by large-scale waves slows down the jet and leads to its reversal, and the change develops poleward and downward. GWF above ~70 km weakens and then changes to eastward, along with the reversal of the stratospheric jet. The GWF is weak when the mesospheric jet recovers while the stratospheric wind is still westward (e.g. on 22 February). These GW changes are qualitatively consistent with the wind filtering scenario, as obtained from parameterization studies.

As reported by Yoshida and Yamazaki (2011), both the stratosphere and tropopause (between 150–100 hPa) in the tropics undergo cooling around the time of 2009 SSW, but the two are not clearly connected. Evan et al. (2015) also found that the cold point tropopause temperature drops by 2–3 K around the time of the 2013 SSW. Similar changes in the stratosphere and near the tropopause are seen in the current simulation (Fig. 3a). Following Kodera (2006) the date when the zonal mean zonal wind at high latitude and 10 hPa displays the largest deceleration rate is used as a reference date (18 February), and it is seen that the tropical stratosphere shows a clear cooling that progresses down toward the lower stratosphere. A strong cooling of ~3 K reaches down between 60–70 hPa about 2 weeks later. The temporal/spatial variation and the magnitude of this stratospheric cooling are similar to those found during 2009 SSW (Yoshida and Yamazaki 2011). Between 150–100 hPa, a cooling occurs between 18 February and 6 March, with the strongest cooling of 3 K found on 24 February. These features are again similar to the tropical tropospheric changes during 2009 SSW. These two cooling regions are separated by an apparent “gap” between 100–80 hPa, which shows much weaker temperature change. Water vapor at the tropical tropopause decreases as a result of the cooling, as reported by Evan et al. (2015), and this is seen in the simulation (Fig. 3b). The zonal mean water vapor at the tropical tropopause, averaged over 15°S–15°N, drops to a minimum of 2.2 ppmv on 6 March. This temporal variation seems to correspond closely to the temperature change at 70 hPa (Fig. 3a).

These temperature changes are also similar to those found by Kodera (2006) through statistical analysis of SSW events. By contrasting the temperature and zonal wind changes from ~1 week prior to the reference date to 1 week after (Figs. 3c and 3d)
we obtain patterns similar to those found in Kodera (2006): for temperature, warming at high latitudes in the stratosphere and troposphere and cooling at low to middle latitudes in the stratosphere down to the tropopause, and eastward changes at low latitude in the stratosphere. It is worth noting that this simulated warming is particularly strong and both the warming and westward wind reversal reach down deep into the troposphere at high winter latitudes (Fig. 3d). Figure 3d shows that the tropical zonal wind has a strong westward change between 200−100 hPa. Although this is not seen in the statistical analysis by Kodera (2006), it is consistent with the westward forcing by the equatorward-propagating waves that dissipate around tropical tropopause and drive the upwelling (Yoshida and Yamazaki 2011).

As suggested by Thuburn and Craig (2000) and by Kodera (2006), changes of stratospheric meridional circulation can lead to changes of the depth of convective heating in the tropics. During SSW, the cooling of the lower stratosphere/tropopause leads to the increase of convective heating depth. A case study by Kodera and Yamada (2004) and the statistical analysis by Kodera (2006) further revealed that the tropical convection increases in the SH and decrease in the NH, and the most significant increases in SH condensational heating and vertical wind occur 5−10 days after the aforementioned reference date. Upward movement of tropical tropopause layer is also noted in the case study by Evan et al. (2015), with peak upward motion found about two weeks after the maximum warming. Similar changes in condensational heating and vertical motion are seen in this simulation (Fig. 4a): Both the magnitude and depth of zonal mean condensational heating rate averaged over 0°S−25°S reach maximum values (~2 K d−1 between 600−300 hPa) about a week after 18 February. The vertical motion changes accordingly (Fig. 4b): The vertical wind peaks on 25 February with a maximum value of 7 mm s−1 near 300 hPa. These changes are also consistent with tropical tropopause cooling shown in Fig. 3a. In spite of the large variability of these troposphere/tropopause quantities, the strengths of their variation and their consistency with previous case studies and statistical analysis suggest the variation is likely related to the SSW.

These changes in deep convection can cause GW variability. Figures 5a, 5b, 5c, 5d, 5e, 5f, and 5h show the zonally averaged vertical fluxes of zonal momentum of GWs with zonal scales between 0−1600 km from 20−90 km. Up to 60 km, the net momentum flux over ~20°S is positive (upward fluxes of eastward momentum), due to a westward background wind (as can be seen in Fig. 2). Its temporal variation corresponds quite well with the convection change, with a decrease prior to the major warming and an enhancement afterward (maximum on 26 February immediately after the peak condensational heating). Although convectively generated GW can increase in both eastward and westward directions at source level as a result of enhanced condensational heating, it is unclear why there is an enhancement at source level as a result of enhanced condensational heating.
heating, the eastward increase becomes more apparent probably because of preferential filtering of the westward components. At 70 km and above, the peak GW activities in SH are found at higher latitudes, and decrease quite markedly around 26 February. They appear less correlated with the changes in the subtropical region at lower altitudes, and probably result from the interaction with the wind: The general shift of GW activities to higher latitudes with altitudes in the SH is congruent with the upward/poleward tilt of the mesospheric wind reversal in the summer hemisphere. The decrease of the GW eastward momentum flux at higher altitude is probably related to the eastward shift of the mesospheric wind in the summer hemisphere from the inter-hemispheric coupling during SSW (as seen in Fig. 2. See also Körnich and Becker 2010). The variation of the GW momentum flux leads to variation of GWF (Figs. 5i, 5j, 5k, and 5l in the mesosphere where the forcing is most prominent), with the eastward GWF over the subtropics in the SH peaking around the time of SSW at 60 and 70 km. The eastward GWF at higher altitudes and higher latitudes also decreases significantly around 26 February, like the momentum flux. It is noted that the wind changes in the southern subtropical mesosphere during the SSW can be westward (Fig. 1b), signifying the complex dependence of GW momentum flux and forcing on the large-scale flow as well as the wave sources, and the challenge to unambiguously isolate the various processes. At high latitudes in the winter hemisphere, the zonal mean momentum fluxes between 30−80 km are mostly westward prior to SSW and change to eastward during SSW. Following SSW, the eastward momentum flux persists longer in the stratosphere than in the mesosphere, consistent with the zonal wind variation (Fig. 1a). The momentum fluxes variability can be very large around the peak warming time at some altitudes (40−70 km), and can be either strongly westward or eastward. It is also noted that the magnitudes of the momentum flux are generally lower following the SSW than that before the warming, regardless of the sign. For the mesosphere, this is probably because the persisting westward wind in the stratosphere after the SSW filters out a portion of the westward GWs. These results agree with observations of gravity wave momentum flux change during SSW in the stratosphere and mesosphere (Wright et al. 2010; France et al. 2012; Thurairajah et al. 2014; Ern et al. 2016). The GWF varies similarly during SSW: the predominant westward forcing becomes eastward at mid to high latitudes in the winter hemisphere. The westward forcing poleward of 50°N at 80 and 90 km after SSW drives poleward/downward circulation at lower altitudes, which in turn leads to the formation of the elevated stratopause.

The distribution of the GW momentum fluxes at 30 km on 4 and 22 February are shown in Figs. 6a and 6b. 4 February is chosen because this is before the deceleration and poleward shift
of the winter stratospheric jet (as seen from Fig. 2), and is valuable for studying GWs associated with the jet. Clusters of eastward momentum fluxes are found in the SH centered around 20°S, presumably due to tropical convection. The momentum fluxes over Southern Atlantic Ocean-Africa-Indian Ocean and over South America are larger on 22 February, while those over central and eastern Pacific Ocean extend westward. These changes contribute to the increase of the zonal mean momentum fluxes in the southern tropical region (Fig. 5), but it is not clear which are associated with SSW.

The changes at middle and high latitudes in the NH, on the other hand, are clearly caused by SSW. Prior to SSW (4 February), large westward momentum fluxes are found extending from the Atlantic Ocean eastward and northward to Bering Sea. On 22 February, the momentum fluxes to the east of 100°E change sign and become weakly positive. At the same time, strong eastward momentum fluxes appear at high latitude between 150°W and 30°E. These changes are likely caused by the changes of tropospheric and stratospheric winds: The zonal wind at 170°W (where the wave momentum flux is strongly westward before SSW) is predominantly eastward in the troposphere and stratosphere on 4 February (Fig. 6c), and it is thus conducive for the upward propagation of GWs. On 22 February, the wind reversal extends down from stratosphere to the troposphere and the zonal wind at this longitude is either 0 or weakly westward, and the zero wind region is quite broad (Fig. 6d). These are unfavorable for either the excitation or the propagation of GWs. Over the Canadian Arctic Archipelago and Greenland, on the other hand, the zonal wind becomes more favorable for wave excitation and propagation after SSW: it becomes strongly westward from the surface to the stratosphere between 65°N-85°N after SSW, while before there is a 0 wind line over most of that latitude range.

Evidence of GW generation from the jets is also found in the simulations. In Fig. 6a, it is seen that on 4 February, the momentum flux has positive values between 60°N−70°N and 100°E−160°E, and negative values to the east (to ~140°W) at the same latitudes. By comparing the momentum flux with the pressure perturbation, it is found that both areas with large positive and negative momentum flux values are located in the jet exit region, with Aleutian high to the east and a low around Novaya Zemlya to the west. Given the wind direction, the positive values of momentum flux likely correspond to downward wave propagation and the negative values to upward propagation. This is confirmed by
forcing in the Southern tropical/subtropical region up to the mesosphere is consistent with the increase of the enhanced convection.

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