Secondary planetary waves in the middle and upper atmosphere following the stratospheric sudden warming event of January 2012

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1The role of planetary waves in causing stratospheric sudden warmings (SSWs) is well understood and quantified. However, recent studies have indicated that secondary planetary waves are excited in the mesosphere and lower thermosphere following SSWs. We use a version of the Whole Atmosphere Community Climate Model constrained by reanalysis data below 50 km to simulate the SSW of January 2012, a minor warming followed by the formation of an elevated stratopause. We document the occurrence of enhanced Eliassen-Palm flux divergence in the mesosphere and lower thermosphere associated with faster, secondary westward-propagating planetary waves of wave number 1 and period <10 days. We confirm the presence of these secondary planetary waves using observations made by the Sounding of the Atmosphere using the Broadband Emission Radiometry instrument onboard NASA’s Thermosphere-Ionosphere-Mesosphere Energetics and Dynamics satellite. Citation: Chandran, A., R. R. Garcia, R. L. Collins, and L. C. Chang (2013), Secondary planetary waves in the middle and upper atmosphere following the stratospheric sudden warming event of January 2012, Geophys. Res. Lett., 40, 1861–1867, doi:10.1002/grl.50373.

1. Introduction

[2] The wintertime Arctic middle atmosphere is characterized by the presence of large amplitude planetary Rossby waves that often interact with the zonal-mean flow and trigger stratospheric sudden warming (SSW) events. SSWs are characterized by the reversal of the zonal-mean winds at middle to high latitudes and the development of a positive poleward temperature gradient over the polar cap in the stratosphere. The planetary waves in the wintertime middle atmosphere are primarily tropospheric in origin and propagate upward. These waves are mostly quasi-stationary Rossby waves and traveling waves, including those with periods around 10 and 16 days, which correspond to natural modes of variability of the Earth’s atmosphere [see, for example, Madden, 2007]. Pancheva et al. [2008] noted the presence of a westward propagating ~16 day wave of wave number 1 in the stratosphere and mesosphere before the 2003/2004 SSW.

[3] While the role of planetary waves in the generation of SSWs is well understood and quantified, the generation and propagation of secondary planetary waves after SSW events is not. We use the term “secondary waves” to denote perturbations that are not present before the SSW and are not associated with the quasi-stationary or traveling wave field in the stratosphere; these waves often appear in the mesosphere and lower thermosphere (MLT) following a SSW. Amplification of secondary planetary waves of wave number 1 was noted during the 2006 SSW [Siskind et al., 2007, 2010; Manney et al., 2008]. While Manney et al. [2008] suggested that the enhanced wave 1 might have been forced in situ by zonally asymmetric gravity wave breaking, Siskind et al. [2010] provided evidence contrary to this hypothesis; they argued that the wave enhancement was likely due to a combination of forcing from below and in situ instability near the stratopause. Modeling studies of SSW events almost always show an enhancement in Eliassen-Palm (EP) flux divergence in the mesosphere immediately following the SSW [Limpasuvan et al., 2012; Chandran et al., 2011, 2013; Tomikawa et al., 2012]. Chandran et al. [2013] also show that the enhancement of EP flux divergence in the MLT following SSWs is a robust climatological feature in composites [their Figures 5 and 6] of elevated stratopause (ES) and SSW events simulated by the Whole Atmosphere Community Climate Model (WACCM). In this study we document the presence of secondary enhanced EP flux divergence in the MLT following the January 2012 SSW, simulated using a “specified dynamics” version of WACCM (described in section 2, termed Specified Dynamics (SD)-WACCM). We confirm the presence of these secondary planetary waves in the MLT with observations from the Sounding of the Atmosphere using the Broadband Emission Radiometry (SABER) instrument on board the NASA Thermosphere-Ionosphere-Mesosphere Energetics and Dynamics (TIMED) spacecraft.

2. Specified Dynamics Whole Atmosphere Community Climate Model

[4] The WACCM is a fully coupled chemistry climate model derived from the Community Atmosphere Model [see Garcia et al., 2007, and references therein] that extends from the Earth’s surface to the lower thermosphere (~145 km or 4.5 × 105 hPa). Although WACCM is usually operated as a free-running climate model, as SD-WACCM it is operated using reanalysis data to constrain the dynamics and temperature in the troposphere and stratosphere. In SD-WACCM, the horizontal winds and the temperature are relaxed to GEOS-5.2 reanalysis data [Rienecker et al., 2008] from the surface to 40 km. The strength of the relaxation is linearly reduced between 40 and 50 km, such that the model...
becomes free running above 50 km. In this mode, it is possible to use WACCM to simulate and study a particular meteorological event. For example, Marsh [2011] used SD-WACCM to study the major SSW of 2006. SD-WACCM has 88 vertical levels, with a resolution of approximately two grid-points per scale height in the mesosphere and lower thermosphere. The horizontal resolution is $1.9^\circ \times 2.5^\circ$ (latitude \times longitude). The simulations used in this study have a time step of 1800 s for the physical parameterizations, with model output at midnight UT every day. SD-WACCM resolves large-scale waves, including planetary waves, but the effect of mesoscale orographic and nonorographic small-scale gravity waves is parameterized in the model [Garcia et al., 2007; Richter et al., 2008, 2010]. We use established analysis techniques to determine the EP flux divergence [Andrews et al., 1987] due to planetary-scale Rossby waves from this daily output.

3. Sounding of the Atmosphere using the Broadband Emission Radiometry

[5] Sounding of the Atmosphere using the Broadband Emission Radiometry is a limb scanning, 10 channel broadband infrared radiometer, which measures temperature, ozone, water vapor, carbon dioxide, nitric oxide, and airglow emissions from the stratosphere to the thermosphere [Russell et al., 1999; Mlynczak 1997; Remsberg et al., 2003, 2008]. For this study we have used version 1.07, level 2A retrievals of temperature from January to April 2012, spanning the range of altitude ~15 to 110 km. To analyze the SABER data, we have employed the fast Fourier synoptic mapping technique developed by Salby [1982a, 1982b] that allows us to obtain synoptic spectra from asymptotic measurements. This method has previously been used for planetary wave analysis of SABER temperature measurements [Garcia et al., 2005]. We use the SABER measurements at 52°N to characterize the planetary wave activity as this is the northernmost latitude that is continuously sampled by SABER. Higher latitudes are sampled only on successive 60 day yaw cycles [Remsberg et al., 2008].

4. Analysis of SD-WACCM Simulation of the January 2012 SSW

[6] We first describe the SD-WACCM simulation of the January 2012 SSW event. Figure 1 shows (a) the evolution of EP flux divergence, (b) geopotential height perturbation due to planetary wave 1, and (c) zonal-mean zonal wind, all averaged over midlatitudes (55°–70°N), and (d) zonal-mean temperature averaged over the polar cap (75°–90°N). The EP flux divergence (Figure 1a) and wave 1 geopotential height perturbation (Figure 1b) show an enhancement in westward (negative) planetary wave forcing and wave 1 growth in the stratosphere (between 40 and 60 km) between days −8 and 2. The westward EP flux divergence due to these waves weakens the eastward zonal-mean zonal wind (but does not cause a reversal in the stratosphere). The enhanced wave forcing does lead to a slight warming of the stratosphere and a clear downward displacement of the stratopause at the very beginning of 2012 (day 0), as can be seen from Figures 1c and 1d, respectively.

[7] Following the initial period of wave growth and decay, there is a subsequent period of planetary wave growth after day 10 at 30–60 km, during which the peak in westward EP flux divergence occurs around day 15, with values that reach ~40 m s$^{-1}$ d$^{-1}$. The zonal-mean wind weakens significantly after day 10 and starts reversing by day 13 in the stratosphere (indicated by the dashed lines in Figure 1); the reversal extends a little below 40 km on days 17–18, but does not reach the 10 hPa (~32 km) level, which makes this a minor warming. In the stratosphere, there is pronounced warming and further descent of the stratopause, with the zonal-mean temperature over the polar cap (75°–90°N) reaching ~270 K near 45 km on day 18. A double stratopause is present from day 20 until day 28, when a single, ES (~240 K) is formed at 70–75 km. Eastward winds are re-established in the mesosphere by day 20, but the wind remains reversed between 40 and 55 km until day 23. The westward zonal-mean wind reaches peak values of about ~40 ms$^{-1}$ on day 17 between 40 and 60 km. The peak amplitude in geopotential height perturbation in the stratosphere is ~2000 m, reached initially during the first phase of wave growth between day −10 and day 0, and during a second period of wave growth between day 10 and day 12.
It can also be seen from Figures 1a and 1b that there is enhanced westward EP flux divergence between 80 and 100 km between day 12 and day 23, with largest negative values in this altitude range occurring on day 20, which corresponds to the growth in wave 1 geopotential height in the MLT. There are also short periods (4–5 days) during which regions of eastward (positive) EP flux divergence develop at 60–80 km, around day 0 and again around day 18. Wave 1 is the dominant feature in the stratosphere and MLT region; wave 2 amplitudes are much weaker (by more than a factor of 4) and, therefore, they are not shown or discussed in this study. The gravity wave forcing in the MLT (not shown) reverses from westward to eastward during the reversal of the zonal mean wind in the stratosphere (day 3 to day 21) and exhibits a behavior similar to the case study and composite behavior illustrated in Chandran et al. (2011, 2013).

The zonal-mean polar cap temperature structure is similar to that observed in the 2004, 2006, 2009, and 2010 SSW events [Manney et al., 2008, 2009; Thuraiarajah et al., 2010a, 2010b]. These SSW events show a warming and subsequent descent of the stratopause, followed by the formation of an ES in the upper stratosphere and lower mesosphere between 70–80 km. ES are formed primarily by gravity wave driven adiabatic downwelling in the MLT [Chandran et al., 2011; Tomikawa et al., 2012]. Unlike the 2009 SSW event, which was a vortex splitting event, the 2012 SSW event is a vortex displacement event similar to those of 2004 and 2006. Compared to the SSW events of 2004, 2006, and 2009, when the lower stratosphere zonal mean winds remained westward for more than three weeks, the duration of the wind reversal during the 2012 SSW is much shorter, lasting ~10 days at 45 km. The lowest altitude that experiences a wind reversal is ~40 km. Both de la Torre et al. [2012] and Chandran et al. [2013] have previously noted that ES events can occur following minor SSW events. The SSW of January 2012 is one such example.

5. Planetary Waves in the Stratosphere and MLT During the SSW Event

To characterize the propagation and evolution of planetary waves in the stratosphere and mesosphere, we show in Figure 2 Hövsmoller plots of geopotential height perturbation averaged between altitudes of (a) 40–50 km and (b) 80–100 km, and latitudes of 55°–70°N. The planetary wave field in the stratosphere (Figure 2a) shows predominantly a wave number 1, quasi-stationary structure throughout the period of analysis, along with the two periods of growth and decay, initially between day ~10 and day 2, and subsequently between day 5 and day 20, which results in the SSW. The geopotential height perturbation in the MLT, between 80 and 100 km (Figure 2b), shows quasi-stationary wave behavior similar to that seen at 40–50 km until about day 15. However, between approximately days 12 and 30, the geopotential perturbation in the MLT shows clear evidence of westward propagating waves, which are not present in the stratosphere.

Because wave 1 is the dominant planetary wave feature, we process the wave 1 component of the geopotential height with a continuous wavelet procedure [Torrence and Compo, 1998] to determine its dominant periodicities in the stratosphere and MLT. In Figure 2c we show the wavelet power spectrum of wave 1 in geopotential height between 55°N and 70°N averaged over altitudes 40–50 km and, in Figure 2d, over altitudes of 80–100 km. We have used a Morlet wavelet with the wavelet power spectrum normalized by a scale factor to rectify the power bias at lower frequencies [Liu et al., 2007]. The peaks shown in Figures 2c and 2d are all significant at the 95% significance level and are not affected by any edge effects because the WACCM time series used in the wavelet analysis cover a longer period (20 October 2011 to 23 March 2012) centered on the period of interest. (Note, however, that the wavelet procedure does not distinguish between eastward and westward propagating waves.) From Figure 2c, the nonstationary wave 1 field in the stratosphere has a dominant periodicity near 16–17 days. The wavelet analysis shows a broad peak for this wave occurring between day 0 and day 20, which corresponds to the period leading to the SSW. However, between 80 and 100 km, the wave 1 field shows several shorter period waves in addition to the 16–17 day wave. Around day 21, when westward EP flux divergence in the MLT is largest (Figure 1a), there are peaks in the spectrum at periods of 6–7 days, and at 3 days, 11–12 days and 16–17 days. The faster waves (periods of about 10 days or less) attain substantial amplitude from about day 10 to day 30. The fast, westward propagating wave structure seen in Figure 2b between days 12 and 30 reflects the superposition of these waves. It should be noted that the wavelet spectrum does not show the presence of any short period waves in the stratosphere.

In Figure 2e we show the spectrum of RMS temperature amplitude vs. frequency (in cycles per day, cpd) for zonal wave number 1 at 52°N and 13 scale heights (approximately 91 km) obtained from an analysis of SABER data over the period 27 October 2011 to 25 April 2012, which encompasses the period of interest for this study. Negative and positive frequencies indicate eastward and westward propagating waves, respectively. The spectrum is shown at 52°N, which is on the equatorward side of the latitude band used to obtain the WACCM spectra shown in Figures 2c and 2d. Nonetheless, it can be seen that there is a peak near zero frequency, which corresponds to the quasi-stationary component, and three other peaks at westward frequencies, indicated by the dashed lines at 0.05–0.06 cpd (period of 20–16 days), 0.09–0.1 cpd (11–10 days) and 0.14–0.18 cpd (7–5.5 days). The prominent periodicities seen in the SABER temperature spectrum agree well with those seen in the wavelet frequency spectrum of Figure 2d. The frequency spectrum also shows that a substantial portion of the variance around 15–16 days is westward propagating, as is most of the variance near 10 and 6.5 days. Figure 2f shows a Hövsmoller diagram of wave 1 RMS temperatures observed by SABER at 52°N and 91 km, for the period 1 January to 1 March 2012, band-passed in the frequency range ~0.25 to 0.25 cpd. This is generally consistent with the Hövsmoller plot of geopotential height for the MLT shown in Figure 2b. The fast westward propagating wave can be seen between 10 and 30 January in the SABER temperature amplitudes with a peak perturbation in temperature amplitudes happening around 20 January. We do not expect exact agreement between Figures 2f and 2b, as the former shows temperature at 52°N and 91 km, whereas the latter depicts the geopotential averaged over 55°N–70°N and 80–100 km.

To investigate the contribution of planetary waves with different periods to the net EP flux divergences in SD WACCM, we determine the EP flux divergence for short period (<10 days) waves, medium period (10–20 days) waves, and long period (>20 days) waves. We actually examine the behavior of the quantity \( \cos^2 \theta F \), where \( F \) is...
Figure 2. Höv moller diagram of geopotential height perturbation between 55°N and 70°N at (a) 40–50 km altitude and (b) 80–100 km altitude. Contour levels are shown every 500 m. The period between days 12 and 30 is demarcated by dashed lines in panel (b) to emphasize the presence of fast, westward propagating waves in the MLT at that time. Wavelet power spectra of wave 1 averaged between 55°N and 70°N are shown for (c) 40–50 km and (d) 80–100 km altitudes. (e) SABER wave number 1 frequency spectrum for the period November 2011 to April 2012 at 52°N and ~ 91 km; negative frequencies indicate eastward and positive frequencies indicate westward waves. (f) Höv moller diagram of SABER wave number 1 temperature amplitudes at 52°N and ~ 90 km in the frequency range (~0.25, 0.25 cpd) for the first 60 days of 2012.
The zonal-mean zonal acceleration due to wave growth or dissipation, that is
\[ F = \frac{1}{\rho a \cos \theta} \nabla \cdot \mathbf{F} \]  
(1)
and \( \nabla \cdot \mathbf{F} \) is the divergence of the EP flux in spherical coordinates [Andrews et al., 1987, Ch. 3]
\[ \nabla \cdot \mathbf{F} = \frac{1}{a \cos \theta} \frac{\partial (\cos \theta F_\theta)}{\partial \theta} + \frac{\partial F_z}{\partial z} \]  
(2)

[14] Multiplying (1) times \( \cos^2 \theta \) and making the convenient (albeit weakly justified) assumption
\[ \frac{1}{\rho} \frac{\partial}{\partial z} \left( \frac{F_z \cos \theta}{a} \right) \approx \frac{\partial}{\partial z} \left( \frac{F_z \cos \theta}{\rho a} \right) \]  
(3)
one can write
\[ \cos^2 \theta F = \frac{\partial F_\theta}{\partial \theta} + \frac{\partial F_z}{\partial z} \]  
(4)
where
\[ F = (F_\theta, F_z) = \left( \frac{F_\theta \cos \theta}{\rho a}, \frac{F_z \cos \theta}{\rho a} \right) \]  
(5)
is a wave activity flux vector, akin to the EP flux. While assumption (3) is not generally valid, it should hold best in regions where the vertical rate of change of \( F_z \) dominates the derivative on the right-hand side of (3); this will be the case wherever wave amplitude changes rapidly with height, as in regions of vigorous growth and dissipation. Insofar as assumption (3) holds, it follows from (4) that contours of \( \cos^2 \theta F \) will be consistent with the apparent convergence/divergence of \( \mathbf{F} \) in the Cartesian coordinate system \((\theta, z)\). We emphasize that, because assumption (3) is not generally justified, \( \mathbf{F} \) must be viewed as a tool to visualize qualitatively the propagation and dissipation of wave activity flux over a very broad range of altitude. On the other hand, the actual EP flux vector, \( \mathbf{F} \), cannot in practice be plotted over a deep layer of the atmosphere because it varies by several orders of magnitude between the stratosphere and the MLT.

[15] Figure 3 shows \( \mathbf{F} \) vectors superimposed on contours of \( \cos^2 \theta F \), both averaged over days 19 to 23 (following the

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**Figure 3.** Latitude-height cross-sections of the quantity \( \cos^2 \theta F \) (color-filled contours) averaged between days 19 and 23 for (a) all waves, (b) short period waves, (c) medium period waves, and (d) long period waves. Superimposed on the contours of \( \cos^2 \theta F \) is the vector field \( \mathbf{F} \), which is a measure of the wave activity flux. \( F \) is the zonal-mean acceleration due to planetary wave EP flux divergence; see text for details. The red contours enclose regions where \( \tilde{q}_y < 0 \), the necessary condition for baroclinic or barotropic instability. The zonal-mean zonal wind distribution (m s\(^{-1}\)) is indicated by the black contours.
SSW, when the peak enhancements in EP flux divergence occur in the MLT). The zonal-mean zonal wind is plotted as black contours (at 10 m s$^{-1}$ intervals); the red contours enclose regions where the latitudinal gradient of zonal-mean quasi-geostrophic potential vorticity [$q$; Matsuno, 1970] reverses sign, that is, where $q_i < 0$. The presence of such regions is a necessary condition for the occurrence of barotropic or baroclinic instability [e.g., Pedlosky, 1979, Ch. 7]. Figure 3a illustrates the EP flux divergences due to all the waves. The enhancement of westward (negative) cos$^2\theta F$ in the MLT between 80–100 km and 20°N–70°N (shaded in blue), is evident, as is a region of eastward (positive) cos$^2\theta F$ between 60–80 km (shaded in red), which lies above a region of $q_i < 0$ where the zonal wind is westward. This region acts as a source for planetary wave activity excited by instabilities of the background, zonal-mean state. The direction of the $\mathbf{F}$ vectors in Figure 3a indicates propagation from the region of positive EP flux divergence, located between 60–80 km, to lower latitudes and higher levels in the MLT. In Figures 3b–3d we show cos$^2\theta F$ individually for waves of short period (<10 days), medium period (10–20 days) and long period (>20 days). From these figures, we conclude that the enhancement of westward cos$^2\theta F$ in the MLT between 80–100 km and 20°N–70°N is primarily due to the short period waves, with a smaller contribution from the medium period waves. In addition, regions of eastward cos$^2\theta F$ are associated mainly with the short period waves. The $\mathbf{F}$ vectors indicate that the short period waves originate between 50–80 km. The medium period waves also undergo some enhancement between 60–80 km and contribute to the enhancement in westward cos$^2\theta F$ between 90–100 km. The long period or quasi-stationary waves are not associated with positive cos$^2\theta F$ and appear to be propagating from lower levels.

6. Discussion and Conclusions

[16] The January 2012 sudden warming is an example of a minor SSW followed by an ES event. Compared to other recent events (2004, 2006, 2009) in the Arctic [Mannay et al., 2008, 2009; Thurairajah et al., 2010a, 2010b], the zonal mean wind in midlatitudes remains reversed for a much shorter time, but the evolution of polar cap temperature closely matches that seen after the stronger events. Before the SSW, quasi-stationary and traveling waves of wave number 1 are present in the stratosphere. Spectral analysis of the latter shows concentrations of westward propagating variance at periods of about 11–12 days and 15–16 days, which correspond approximately to the theoretical periods of the second antisymmetric and symmetric Rossby normal modes of wave number 1. Following the SSW, westward propagating planetary waves of wave number 1 and periods of 3 days and 6–7 days are seen in the MLT for approximately 10–15 days and are distinct from the planetary waves in the stratosphere. SABER temperature observations also indicate the presence of these waves in the MLT. From their absence below 50 km, it can be inferred that their source lies in the mesosphere, and is most intense following the occurrence of the SSW.

[17] The results shown in Figure 3 indicate that the shorter period planetary waves cause the enhancement of wave forcing seen in the MLT after the SSW. The regions of positive wave activity flux divergence and the wave activity flux vectors shown in that figure are consistent with excitation via instability in the mesosphere (between 50–80 km) at latitudes poleward of 40°N. A similar analysis, centered around the time of the weak warming episode that occurs at the very beginning of 2012 (days –2 to +2 in Figure 1a) indicates that the positive wave activity flux divergence seen in the MLT region at that time is also associated with waves of periods less than 10 days. Furthermore, the wave activity flux divergence pattern (not shown) is consistent with wave excitation via instability of the background state. On the other hand, the large negative EP flux divergence that occurs in the MLT in the period leading to the SSW (days 1–15 in Figure 1a) is associated primarily with long-period (>20 days) waves. Examination of wave activity flux plots on days 1–15 (not shown) shows that wave activity propagates mainly from the stratosphere to the MLT. While there is some contribution to westward forcing in the MLT from faster waves, there is no clear indication of a region of wave generation in the high latitude mesosphere such as seen in Figure 3b.

[18] The presence of a 6–7 day wave of wave number 1 between 50–60 km following a SSW was also reported by Tomikawa et al. [2012] in their study of a model-generated ES event. However, their model has a top level of 80 km and the propagation of the wave to the MLT could not be seen as clearly as in the present case study. Meyer and Forbes [1997], in their investigation of an observed 6–7 day wave during August/September 1993, found that the wave was connected to atmospheric instability of the mesosphere. Lieberman et al. [2003] and Liu et al. [2004] have provided evidence for high-latitude instabilities enhancing the 6.5 day wave. Liu et al. also showed the presence of instabilities in the Northern Hemisphere high latitudes in winter following the occurrence of a SSW. They noted that the phase speed of the 6.5 wave is approximately the same as that of the 2 day wave and suggest that both waves are excited by instabilities in the zonal-mean zonal wind distribution. This is consistent with Plumb’s [1983] proposed baroclinic instability mechanism for the origin of the quasi-two day wave in the summer mesosphere. Plumb’s midlatitude beta-plane analysis indicates a phase speed for unstable modes of 50–60 m s$^{-1}$ (his Figure 2); this is comparable to the phase speeds of the 2 day and 6.5 day waves, which are about 50 m s$^{-1}$ at 45° latitude. It thus appears that, after SSW events, the reversal of the stratospheric jet at high latitudes in the winter hemisphere leads to conditions conducive to the generation of short period waves via instability of the background zonal wind field.

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