Opposite Contributions of Stationary and Traveling Planetary Waves in the Northern Hemisphere Winter Middle Atmosphere

Koki Iwao¹ and Toshihiko Hirooka²

¹National Institute of Technology, Kumamoto College, Yatsushiro, Japan, ²Department of Earth and Planetary Sciences, Kyushu University, Fukuoka, Japan

Abstract This study investigates contributions of planetary waves (PWs) to the mean flow in the middle atmosphere during the Northern Hemisphere winter when PW activities are strong, by analyzing satellite data from Sounding of the Atmosphere using Broadband Emission Radiometry (SABER). It is suggested that there would be opposite contributions between easterly acceleration by monthly mean stationary planetary waves (STPWs) and westerly acceleration by remaining traveling planetary waves (TRPWs) in the high-latitude lower mesosphere. When easterly winds appear in the mesosphere, STPWs propagating from the troposphere dissipate in the lower mesosphere to accelerate easterly winds because STPWs could not propagate in easterly wind regions. On the other hand, TRPWs develop there due to the zonal flow instability to accelerate westerly winds. The easterly wind in the mesosphere is frequently triggered by dissipation of westward TRPWs developed in the stratosphere, and would be maintained by these opposite contributions for more than 10 days. Further analyses suggest that the TRPW developing in the high-latitude lower mesosphere would be an eastward TRPW of zonal wavenumber 1 and a period of 10–40 days. This TRPW would develop and propagate downward to dismiss the baroclinic instability by eliminating warm temperature anomalies formed below the easterly wind region due to the PW dissipation in the mesosphere.

Plain Language Summary The importance of traveling planetary waves (TRPWs) in the stratosphere and mesosphere has been underscored in recent studies to bring about significant impacts on the mean circulation through their generation and dissipation. However, details of the contributions are still unclear. This study reveals the importance of TRPWs on the mean circulation during the Northern Hemisphere winter by analyzing satellite data. It is suggested that there would be opposite contributions between easterly acceleration by monthly mean stationary planetary waves (STPWs) and westerly acceleration by remaining TRPWs in the high-latitude lower mesosphere. When easterly winds appear in the mesosphere, STPWs propagating from the troposphere dissipate in the lower mesosphere to accelerate easterly winds. On the other hand, TRPWs develop there due to the instability to accelerate westerly winds. The easterly wind in the mesosphere is frequently triggered by dissipation of westward TRPWs developed in the stratosphere, and would be maintained by these opposite contributions for more than 10 days. Moreover, the TRPW developing in the high-latitude lower mesosphere would be an eastward TRPW of zonal wavenumber 1 and a period of 10–40 days, which propagates downward.

1. Introduction

The meridional circulation in the stratosphere is mainly driven by quasi-stationary planetary waves (STPWs) especially in the winter hemisphere, while in the mesosphere the circulation is mainly driven by gravity waves (GWs) in both hemispheres (e.g., Plumb, 2002). STPWs in the stratosphere are basically originated in the troposphere and propagate upward, dissipating mainly in the stratosphere to decelerate westerly winds, and drive the meridional circulation. Especially, enhanced STPWs frequently cause stratospheric sudden warmings (SSWs) in the Northern Hemisphere (NH), during which the westerly polar vortex is rapidly decelerated or even reversed in a few days with an abrupt rise of stratospheric polar temperature by several tens of Kelvin (see, e.g., Andrews et al., 1987).

On the other hand, traveling planetary waves (TRPWs) are also known to exist in the middle atmosphere. For example, quasi-2-day waves have been frequently observed in the upper stratosphere and mesosphere...
of the subtropics during summer (e.g., Muller & Nelson, 1978), and 4-day waves have been observed in the polar upper stratosphere and mesosphere during winter (e.g., Venne & Stanford, 1979). It is repeatedly underscored that these TRPWs could be caused by barotropic and/or baroclinic (BT/BC) instability (e.g., Lu et al., 2013, 2017, 2019; Manney et al., 1998, Manney & Randel, 1993; Plumb, 1983; Randel, 1994; Watanabe et al., 2009) in which the potential vorticity shows anomalous latitudinal gradients to fulfill the Rayleigh necessary condition for the BT/BC instability.

Recently, the importance of slower TRPWs has been reported in the onset and/or recovery of SSWs in the NH. Westward TRPWs of wavenumbers 1 and 2 with periods from 5 to 20 days were found to be enhanced in the mesosphere and lower thermosphere after the SSW onsets (Chandran, Garcia, et al., 2013; Pancheva et al., 2008; Stray et al., 2015). These westward TRPWs are considered to be generated in the stratosphere and mesosphere due to the BT/BC instability brought by the wind reversal in SSWs, and contribute to form the elevated stratopause by dissipating in the upper mesosphere and lower thermosphere (Chandran, Collins, et al., 2013; Limpasuvan et al., 2012, 2016; Siskind et al., 2010; Tomikawa et al., 2012). Regarding eastward TRPWs during SSW events in the NH (e.g., Pancheva et al., 2008), Iida et al. (2014) analyzed satellite data for the major SSW event in January 2009, and suggested that easterly winds appearing in the mesosphere just before the SSW might be caused by the dissipation of wavenumber 2 eastward TRPWs formed in the lower mesosphere due to the BT/BC instability. Although these results indicate that TRPWs may have substantial contributions to the general circulation change in the course of the SSW through its generation and dissipation in the stratosphere and mesosphere, details of the contributions are still unclear especially in long-time-mean fields such as monthly means or climatologies.

Hence, the focus of this study is to present contributions of TRPWs to mean fields in the winter middle atmosphere of NH where planetary wave (PW) activities are strong. Moreover, we reveal that mean fields are formed through interactions with STPWs and TRPWs. Here, we used satellite data from the Sounding of the Atmosphere using Broadband Emission Radiometry (SABER), because this instrument can observe the entire middle atmosphere from the lower stratosphere to the lower thermosphere. However, analyses using only the SABER data are not reliable enough in terms of the data quality and derivation methods, so that we supplementally use reanalysis data of Modern-Era Retrospective analysis for Research and Applications, version 2 (MERRA-2) to validate our analyses, although the MERRA-2 data are available only up to the middle mesosphere.

This paper is outlined as follows: Section 2 describes data and analysis methods. Section 3 presents contributions of STPWs and TRPWs in the climatology of the NH winter middle atmosphere, and daily variations are examined in the winter of 2015 as a typical case, and in other three winters as similar cases. Conclusions are given in the final section.

2. Data and Methods

2.1. SABER

The SABER instrument is onboard Thermosphere Ionosphere Mesosphere Energetics Dynamics (TIMED) satellite and measures CO$_2$ infrared limb radiance from approximately 15–120 km altitude along 15 orbits around the earth per day. Temperature profiles are retrieved from the radiance data (Remsberg et al., 2008), and are integrated vertically to determine geopotentials. In this study, we use Version 2.0 level 2A pressure-level data for years from 2002 to 2018. Due to the yaw cycle of the SABER instrument, the latitudinal coverage shifts from 53°S-83°N to 83°S-53°N every 60 days, so that data are unavailable every 60 days in regions poleward of 53°.

For the analysis, SABER data along satellite orbits are averaged every day in bins of 24° in longitude, 5° in latitude, and 1 km in log-pressure vertical coordinates defined as $z = -H \ln(p/p_0)$, where $H = 7$ km is the approximate scale height, $p$ is pressure, and $p_0 = 1000$ hPa is the reference pressure. Hereafter, $z$ is simply referred to as “altitude”. The local time of SABER observations shifts from day-to-day and covers 24 h local time for 60 days, so that local-time variations are aliased in daily variations in these data. In the mesosphere and lower thermosphere, SABER data strongly suffer from contaminations due to local-time variations (e.g., Pancheva et al., 2008), so that we must exclude such components for our dynamical analyses. To exclude
these local-time variation components including diurnal and semi-diurnal tides, we apply following data processing before analyses. First, we construct climatological local-time variations for each grid point by averaging from 2002 to 2018 for four seasons. Next, the climatological local-time variation components derived from the local time and date of the observation are subtracted from each observation. Then the daily 3-dimensional grid data are created by interpolation and averaging on the above mentioned spatial bins.

Figure 1 shows spectra of zonal-mean geopotential heights at 80 km altitude calculated for the SABER data without subtracting local-time variations (Figure 1a) and the data from which local-time variations are subtracted (Figure 1b). These spectra are calculated for each year and then averaged over the analysis period from 2002 to 2018, excluding years containing missing data. The strong spectral peak around 60 days and the weak peak around 30 days seen in Figure 1a almost vanish in Figure 1b; this result implies that these
peaks would be fakes caused by daily shifts in local time of the SABER observation. Daily 3-dimensional grid data without local-time variations are used throughout this study.

2.2. MERRA-2

Because analyses using only the SABER data are unreliable, we supplementally use MERRA-2 data (Bosilovich et al., 2015) to verify fundamental results based on the SABER data. The horizontal resolution of MERRA-2 data are 0.635° and 0.5° in longitude and latitude. The data set we use in this study has 72 model levels from the ground to about 0.01 hPa level (~80 km altitude) which is higher than that in other reanalysis data sets. The temporal resolution of the data is 6 h which is sufficient to exclude the contaminations from tidal components. We make daily averaged pressure-level data by interpolating model-level data into 31 pressure levels.

2.3. Horizontal Winds and E-P Fluxes

Horizontal winds of SABER data are derived from the geopotential heights in the same method as Iida et al. (2014). Zonal-mean zonal winds \( \overline{u} \) are calculated on the basis of the gradient wind approximation:

\[
\frac{\overline{u}^2 \tan \phi + f \overline{\Phi}}{a} + \frac{1}{a} \frac{\partial \overline{\Phi}}{\partial \psi} = 0,
\]

where overbars represent zonal averages, \( a \) is the earth’s radius, \( \phi \) is latitude, \( f \) is the Coriolis parameter, and \( \Phi \) is the geopotential. Randel (1987) showed that the gradient wind approximation gives a better estimate for \( \overline{u} \) than the geostrophic wind approximation in the stratosphere. Eddy components of zonal and meridional winds, \( u' \) and \( v' \), are calculated on the basis of the geostrophic approximation:

\[
u' = \frac{1}{fa} \frac{\partial \Phi'}{\partial \psi}, \quad v' = \frac{1}{fa \cos \phi} \frac{\partial \Phi'}{\partial \lambda},
\]

where primes represent zonal deviations from the zonal-mean, and \( \lambda \) is longitude. As for horizontal winds in the MERRA-2 data, zonal-mean and eddy components are simply derived as zonal averages and zonal deviations of model winds.

In this study, we investigate PW behaviors on the basis of the Transformed Eulerian Mean (TEM) equations proposed by Andrews and McIntyre (1976). The Eliassen-Palm (E-P) fluxes are calculated as follows:

\[
F = \left( 0, F_z, F_x \right) = \rho_0 a \cos \phi \left( 0, -u' \psi', f \frac{\psi'}{\theta} \right),
\]

where \( \theta \) is potential temperature, and \( \rho_0 = \rho_s\exp(-z/H) \) is the standard density of the atmosphere with \( \rho_s \) the density at \( z = 0 \). The magnitude of E-P fluxes reduces rapidly with height so that we depict E-P fluxes divided by \( \rho_0 \) in figures to be recognized easily. The wave forcing to the zonal-mean zonal wind is expressed as E-P flux divergence (Andrews et al., 1987),

\[
DF = \left( \rho_0 a \cos \phi \right)^{-1} \nabla \cdot F
\]

Note that, E-P fluxes in the SABER data are calculated from geostrophic winds, whereas those in the MERRA-2 data are from model winds. In section 3.1, we will compare the zonal-mean zonal winds \( \overline{u} \) and E-P fluxes \( F \) derived from the SABER and the MERRA-2 data.

2.4. The Meridional Gradient of the Potential Vorticity

According to Andrews et al. (1987), the square of the refractive index \( n_2^2 \) is valid in zonal-mean basic fields for describing PW behaviors. PWs tend to be refracted toward maximum regions of \( n_2^2 \) and to avoid regions...
where $n_k^2 < 0$. Distributions of $n_k^2$ are primarily determined by the latitudinal gradient of zonal-mean potential vorticity $q_\zeta$, so that $q_\zeta$ distributions can describe PW behaviors as well. On the other hand, it is known that the existence of negative $q_\zeta$ regions is the necessary condition for the BT/BC instability, that is, the Rayleigh necessary condition (Andrews et al., 1987). In this study, we evaluate the $q_\zeta$ to investigate characteristics of PW behaviors: A maximum region of $q_\zeta$ would be a waveguide of PWs, and a negative region of $q_\zeta$ fulfills the necessary condition of the zonal flow instability.

Regarding to the potential vorticity $q$, we actually use the modified potential vorticity (MPV) defined as the Ertel’s potential vorticity weighted by $\theta^2$ (Lait, 1994):

$$\text{MPV} = -g \left(f + \zeta \frac{\partial}{\partial p} \left( \frac{\theta}{\theta_0} \right)^2 \right).$$

where $g$ is the magnitude of gravitational acceleration, $\theta_0$ is the reference potential temperature, and $\zeta$ is the relative vorticity derived from geostrophic winds in the SABER data and from model winds in the MERRA-2 data. Unlike the Ertel’s potential vorticity, the MPV shows small vertical dependence, and hence the vertical structure of its meridional gradient (MPVy) is easy to capture (Sato & Nomoto, 2015). In this study, the zonal-mean MPVy is used mainly as a measure of the zonal flow instability instead of $q_\zeta$.

PWs we will analyze here are defined as perturbations with zonal wave numbers smaller than or equal to 3. Moreover, we divide PWs into STPWs and TRPWs to evaluate contributions separately. STPWs are defined as monthly means from January 16 to February 15 because of the SABER data limitation due to the yaw cycle, and TRPWs are defined as daily anomalies from STPWs. In the analysis of daily variations in Section 3.2, STPWs are defined as 31-days running means and TRPWs are remaining components.

3. Results

3.1. Climatology and Interannual Variation

Figure 2 shows climatological fields of zonal-mean temperatures and zonal-mean zonal winds $\bar{u}$ in the NH averaged for winters from January 16 to February 15 for years from 2002 to 2018 on the basis of the SABER data.
(Figure 2a) and the MERRA-2 data (Figure 2b). It is found that both data sets show a good coincidence of dynamical structures in the winter NH, for example, the stratopause around 53 km altitude in polar regions, the polar night jet in high latitudes, and the separated subtropical jet in the mesosphere, though the MERRA-2 data covers only up to around 75 km altitude. These structures are also similar to those shown in other observational data such as the 1986 Committee on Space Research (COSPAR) International Reference Atmosphere models (CIRA-86) (Fleming et al., 1990).

However, detailed comparison reveals some differences between these two data sets especially in $\vec{u}$ fields (Figure 5), whereas temperature differences are relatively small within 4 K (not shown). Figure 5 shows that the $\vec{u}$ field in the SABER data overestimates the polar night jet over 5 m/s and underestimates weak westerlies in the high-latitude lower mesosphere by more than 10 m/s. These features are known as the limitation of the gradient wind approximation which neglects the nonlinear effects (Randel, 1987). We need to be careful in evaluating results from the SABER data. Moreover, there is another westerly wind overestimate over 10 m/s in the subtropical lower mesosphere. Regarding this overestimate, Smith and Garcia (2017) warned that the gradient wind approximation might cause erroneous results in the tropical mesosphere by large eddy fluxes due to migrating tides, however, the current study focuses on phenomena in extratropical regions in the NH.

Figure 3 also shows climatological fields related to the PW propagation during the NH winter calculated from the SABER data. E-P fluxes by PWs (Figure 3a) are seen to propagate upward and southward and to converge in the two separated regions around the stratopause (55°N, 50 km) and in the polar mesosphere (65°N, 80 km). A closer observation shows that there could be two main propagation paths: one propagates upward from the high-latitude lower stratosphere and curves southward to converge around the stratopause, the other is from the polar region in the lower mesosphere and turns upward to converge in the polar mesosphere. The amplitude of PWs in geopotential heights (red contours) shows large values in the high-latitude upper stratosphere, and the large amplitude region is extended toward these two convergence peaks.

These characteristics of PW behaviors could be interpreted by illustrating the MPVy distribution (Figure 3b). It can be seen that MPVy maxima are distributed along westerly wind maxima, such as the polar night jet and relatively strong westerlies in the polar upper mesosphere. E-P fluxes are seen to be refracted toward the MPVy maximum regions and converge there around the stratopause and in the polar mesosphere as mentioned above. Moreover, there is another MPVy maximum in the core of the poleward extended subtropical jet in the mesosphere (∼35°N), although the convergence is relatively weak. On the other hand, we can see small or negative MPVy values in the polar region from the upper stratosphere to the mesosphere, meaning that these regions frequently fulfill the necessary condition for the BT/BC instability during the winter, as shown later.

E-P fluxes and their divergences by STPWs and TRPWs are shown in Figures 3c and 3d, respectively. It is found that E-P fluxes by STPWs propagate upward largely in the stratosphere to converge in the widely extended region with a center in the polar lower mesosphere (Figure 3c). However, E-P fluxes by TRPWs, which are large in the mesosphere in addition to the stratosphere, diverge in the high-latitude lower mesosphere and converge in the above mentioned regions (Figure 3d). This divergence region corresponds to the small or negative MPVy region, suggesting that TRPWs could be developed in the high-latitude lower mesosphere due to the BT/BC instability (e.g., Lu et al., 2019). The E-P flux divergence due to the TRPW development has opposite contribution to the convergence by STPWs propagating upward from lower levels. On the other hand, another E-P flux divergence due to TRPWs can be seen in the high-latitude lower stratosphere as well, which, however, does not correspond to the region with small MPVy values. We will discuss this issue in the following subsection.

Moreover, the maximum of the TRPW amplitude is located at higher altitude (∼50 km) than that of STPWs (∼40 km), and the large amplitude region of TRPWs penetrates upward and southward as seen in Figure 3a. These features indicate that TRPWs developing in the stratosphere and mesosphere contribute to form the convergence peaks described above.

In order to validate these results, the same quantities as in Figure 3 but for the MERRA-2 data are shown in Figure 4. All the features seen in Figure 3 below the altitude of 75 km can be identified in this figure, that
is, the peak of E-P flux convergence around the stratopause, MPVy maxima along westerly jets and small MPVy values in the polar region, and the E-P flux convergence by STPWs and the divergence by TRPWs in the high-latitude lower mesosphere, although these features are slightly obscured compared with those in Figure 3.

Figure 5 shows differences of E-P fluxes and MPVy values based on the SABER data from those based on the MERRA-2 data. As mentioned in Section 2, these variables are derived from geostrophic winds in the SABER data, while from model winds in the MERRA-2 data. It is found that the SABER data show equatorward E-P flux anomalies at high latitudes and the poleward anomalies at mid-latitudes, resulting
in their convergences at mid-to high latitudes and divergences in the subtropics and the polar region (Figure 5a). Thus, E-P fluxes in the SABER data overestimate the convergences at mid-to high latitudes and underestimate them in the subtropics and the polar region. These DF anomalies in PWs are considerably large compared to DFs in the MERRA-2 data (up to 80%) and are also seen in DFs by STPWs and TRPWs as (not shown). These anomalies are consistent with Randel (1987) which indicated anomalies in E-P fluxes derived from geostrophic winds in the winter stratosphere.

Moreover, MPVy values in the SABER data show positive anomalies along positive gradient wind anomalies around 50°N–60°N and negative anomalies in the north and south of this region. Randel (1987) also indicated that the geostrophic wind overestimates the westerly polar night jet more than the gradient wind due to a neglect of local curvature effects, which should lead to the MPVy anomalies through the change of the latitudinal curvature of zonal winds.

Figure 4. As in Figure 3, but for the MERRA-2 data. MERRA-2, Modern-Era Retrospective analysis for Research and Applications, version 2.
On the other hand, it is recently reported that discrepancies among reanalysis data increase with height depending on the model configuration used in each reanalysis (Kawatani et al., 2020). Hence, the above anomalies might be caused by plausible deficiencies in the MERRA-2 data. Although there are large differences between the SABER and the MERRA-2 data, investigating PW behaviors by the use of these data sets will be worthwhile, because PW features targeted in this study were identified in both data sets from the stratosphere to the lower mesosphere (Figures 3 and 4).

In Figures 3 and 4, opposite contributions by STPWs and TRPWs to zonal-mean zonal winds \( \vec{u} \) were revealed in the high-latitude lower mesosphere in the winter climatology. We will investigate these contributions in the monthly mean fields for years from 2002 to 2018. Figure 6 shows interannual variations of \( \vec{u} \) and DFs caused by PWs, STPWs, and TRPWs in the high-latitude lower mesosphere (55°–70°N, 55–70 km) during the winter from January 16 to February 15, calculated from the SABER (Figure 6a) and the MERRA-2 data (Figure 6b). Overall, these interannual variations are quite similar between two data sets despite large differences in the climatological means (Figure 5), indicating that these interannual variations are large compared to the climatological means and consistent with each other. In each data, the zonal-mean zonal wind \( \vec{u} \) is positively correlated with the DF by PWs, with correlation coefficients of 0.55 in the SABER and 0.42 in the MERRA-2 data, suggesting the large contribution of PWs to \( \vec{u} \) even in the lower mesosphere. Moreover, DFs by STPWs and TRPWs are negatively correlated each other, with correlation coefficients \(-0.62\) in the SABER and \(-0.65\) in the MERRA-2 data. The large E-P flux convergence by STPWs and the large divergence by TRPWs are simultaneously observed in this region, which is especially clear in 2002, 2007, 2008, 2015, and 2016 (shown by red triangles). Hereafter, the simultaneous occurrence of E-P flux convergence by STPW and the divergence by TRPW in the high-latitude lower mesosphere is referred to as “PW cancellation”.

Here, we examine a plausible relationship to major SSWs. When we detect major SSWs on the basis of the commonly used definition by Charlton and Polvani (2007), that is, \( \vec{u} \) at 60°N and 10 hPa becomes easterly, seven major warming events are found during the period from January 16 to February 15 in the SABER data (shown by black triangles in this figure). It is found that the PW cancellation hardly occurs in winters with major SSWs. This is because easterly winds stay in the stratosphere for a long time after the onsets, and STPWs cannot propagate to the mesosphere and form the instability condition.
Next, dynamical properties of TRPWs in the winter NH are investigated on the basis of a spatiotemporal Fourier analysis (Hayashi, 1971). Since the 60 days window in which the SABER can observe the high-latitude region in the NH gradually shifts year by year, overlapping 40 days from January 16 to February 24 are used for the spectrum analysis and for the following PW separation.

Figure 7 shows amplitudes of zonal wavenumber 1 TRPWs calculated from geopotential heights at 50 km altitude for the 40 days on the basis of the SABER data. Figure 7a is the average for all years (2002–2018) except for years including missing data (2002, 2003, 2005, and 2017), and Figure 7b is the average for 4 years when the PW cancellation was clear in Figure 6 (2007, 2008, 2015, and 2016). Amplitudes of TRPWs in the climatology show a nearly monotonic decrease with increasing frequency, and slightly larger values in westward TRPWs are observed than those in eastward TRPWs (Figure 7a). In years with the PW cancellation, however, the amplitudes of eastward TRPWs with a period of around 20 days are larger around 65°N (Figure 7b). These eastward TRPWs would be generated in the high-latitudes lower mesosphere in years with the PW cancellation. The typical eastward phase speed of these TRPWs is estimated to be ∼10 m/s, which is smaller than the climatological zonal-mean zonal wind $\bar{u}$ in this region (20–30 m/s) shown in Figures 3 and 4. In years with major SSWs (not shown), amplitudes were smaller than the climatology probably
due to the long-lasting easterly winds in the stratosphere after major SSWs as mentioned above. Note that amplitudes of TRPWs of zonal wave numbers 2 and 3 (not shown) are much smaller than those of zonal wavenumber 1.

In order to investigate characteristics of these TRPWs, we separate westward and eastward TRPW components of zonal wavenumber 1 and a period from 10 to 40 days, which are called TRPW1W and TRPW1E, respectively. The 40 days period components seem to be too slow for TRPWs but are included in TRPW1Ws and TRPW1Es, because we would like to investigate slowly developing features in the 40 days limited data span.

**Figure 7.** Period-latitude sections of zonal wavenumber 1 TRPW amplitudes calculated from geopotential heights at 50 km altitude for 40 days from January 16 to February 24 with 50 m contour intervals on the basis of the SABER data. Spectra are averaged (a) from 2002 to 2018 except for years including missing data, and (b) for 4 years when the PW cancellation (see the text for details) was clear (2007, 2008, 2015, and 2016). PW, planetary waves; SABER, Sounding of the Atmosphere using Broadband Emission Radiometry; TRPW, traveling planetary waves.
Figure 8 shows E-P fluxes and DFs calculated from the TRPW1W and TRPW1E components and averaged for years from 2002 to 2018 except for years including missing data. In this figure, E-P flux vectors, DF colors, and wave amplitude contours are enhanced than those in Figures 3 and 4. It is found that E-P fluxes by TRPW1Ws show upward and southward components which converge around the stratopause and in the polar mesosphere (Figure 8a), while E-P fluxes by TRPW1Es show downward and southward components at high latitudes which strongly diverge in the lower mesosphere and weakly converge at mid-latitudes around the stratopause (Figure 8b). These results suggest that the two convergence peaks and the divergence peak by TRPWs shown in Figure 3d are contributed by upward propagating TRPW1Ws and downward propagating TRPW1Es, while the divergence in the stratosphere is due to total TRPWs of TRPW1Ws and TRPW1Es.

3.2. Case Study in 2015 Winter

From winters when the PW cancellation was clear in Figure 6, we select the 2015 winter as a typical case and examine the monthly mean field and the time evolution. Figure 9 shows the same variables as in Figure 3 and Figure 8 but for the monthly mean field for the period from January 16 to February 15 in 2015 calculated from the SABER data. In this winter, easterly winds stay in the polar mesosphere, and in association with that, MPVy values are negative in the polar lower mesosphere (Figure 9b). E-P fluxes by STPWs converge in the front side of this easterly wind region (Figure 9c) because STPWs could not propagate into the easterly wind region, while E-P fluxes by TRPWs diverge in the same region (Figure 9d) probably due to the BT/BC instability. Both wave forcings are related to the easterly winds in the mesosphere, and they are clearer than those seen in the winter climatology (Figure 3). Given no E-P flux divergence by TRPWs, the convergence by STPWs would accelerate easterlies, and easterly winds in the polar mesosphere would descend downward as in cases of SSWs. Thus, it can be considered that the divergence by TRPWs is important for maintaining easterlies in the polar mesosphere.

E-P fluxes and DFs by TRPW1Ws (Figure 9e) and TRPW1Es (Figure 9d) are also shown in this figure. Climatological features of E-P fluxes by TRPW1Ws and TRPW1Es seen in Figure 8 can be identified in this year.
as well, that is, upward E-P fluxes by TRPW1Ws converge in the two convergence regions, downward E-P fluxes by TRPW1Es diverge in the high-latitude lower mesosphere and converge around the stratopause, and both waves contribute to form E-P flux divergences in the stratosphere.
Figure 10 shows daily variations of these variables on the basis of the SABER data. In this figure, STPWs are defined as 31 days running means and TRPWs are daily anomalies from STPWs. Moreover, zonal-mean zonal wind $u$ is averaged for 65°N–75°N, and MPVy is averaged for 65°N–70°N in this figure to show the easterly wind and negative MPVy values clearly, though other variables are averaged for 55°N–70°N.

At first, easterly winds can be seen in the mesosphere for about 20 days after January 27. When easterly winds appear in the mesosphere, negative MPVy values become noticeable around the easterly wind region (Figure 10b), and E-P fluxes by TRPWs diverge in the lower part of the easterly wind region (Figure 10d), while E-P fluxes by STPWs converge in the stratosphere and mesosphere (Figure 10c). These features support the speculation that the westerly wind acceleration by TRPW developments due to the BT/BC instability and the easterly wind acceleration by STPW dissipation cancel out each other, and this cancellation could keep the easterly winds staying in the mesosphere for a long time.

In order to clarify what formed the easterly winds in the mesosphere, we focus on January 27 when the easterly winds appeared. On this day, a minor SSW is seen to occur in the stratosphere. This SSW event is caused by the upward propagation of PWs and the following E-P flux convergence in the stratosphere a few days before the SSW occurrence (Figure 10a). When the convergence decelerates westerlies, MPVy values become negative (Figure 10b), and E-P fluxes diverge in the stratosphere (Figure 10a). This would correspond to a development of TRPWs after SSWs (Chandran, Garcia, et al., 2013; Tomikawa et al., 2012) mentioned in Introduction. In response to this divergence, E-P flux convergences appear in the mesosphere (Figures 10a and 10d), suggesting that TRPWs developing in the stratosphere could propagate upward into the mesosphere. On the other hand, E-P fluxes by STPWs also converge around this level (Figure 10c). The dissipation of TRPWs in addition to STPWs are considered to form the easterly winds in the mesosphere. The easterly winds formed in the mesosphere would be maintained by the PW cancellation as described above.
Similar situation can be seen around January 15 as well (Figure 10a). Easterly winds are formed in the mesosphere probably in association with the occurrence of another minor SSW event on this day, though the easterly wind was maintained only for about 5 days.

As seen in Figure 10d, E-P flux divergences by TRPWs in the stratosphere around 18 and 29 January are observed after negative MPVỹ values, indicating the development of TRPWs due likely to the BT/BC instability. However, on the contrary, large positive MPVỹ values in the lower stratosphere do not correspond to large E-P flux convergences by TRPWs, probably because the upward propagation of TRPWs from the troposphere is small. When these variables are averaged over certain period as in Figures 3 and 9, E-P fluxes by TRPWs might seemingly show divergence in the stratosphere in spite of positive MPVỹ values.

Moreover, contributions by TRPW1Ws and TRPW1Es are shown in Figures 10e and 10f. DFs by TRPWs (Figure 10d) will be attributed to these waves. Note that the calculation is made by components with a period of 10–40 days for 40 days from January 16 to February 24, so that DFs by TRPW1Ws and TRPW1Es appear to be smoothly prolonged. It is found that the E-P flux divergences in the stratosphere seen after the minor SSW around 27 January are attributed to both TRPW1Ws and TRPW1Es, but the corresponding convergences in the upper mesosphere are due to TRPW1Ws. Furthermore, the divergences in the lower mesosphere during easterly winds in the mesosphere are clearly due to TRPW1Es.

Figure 11 shows daily variations of longitudinal distributions of TRPWs, TRPW1Ws, and TRPW1Es in geopotential heights at 50 km altitude and 65°N for the winter in 2015. From Figure 11a, we can see westward TRPWs of zonal wavenumber 1 with a period of about 25 days in the second half of January when easterly winds are formed in the mesosphere, while a clear appearance of eastward TRPWs of zonal wavenumber 1 with a period of about 20 days in the first half of February when the easterly winds are maintained in the mesosphere. Features seen from the filtered components of TRPW1Ws and TRPW1Es in Figures 11b and 11c basically show similar amplified periods. The behavior of TRPW1Ws and TRPW1Es well captures the time evolution of DFs shown in Figures 10e and 10f, that is, the convergence in the second half of January and the divergence in the first half of February around 50 km. The characteristics of the eastward TRPWs are also consistent with the result obtained from the spectral analysis (Figure 7b), while those of the westward TRPWs are weakly detected in Figure 7b.

Figure 12 shows characteristics of TRPWs, TRPW1Ws, and TRPW1Es, on January 27 when the easterly winds appear in the mesosphere (top panels) and on February 2 when the easterly winds are maintained in the mesosphere (bottom panels). On January 27, zonal-mean fields in Figure 12d illustrate that there are easterly winds in the stratosphere, negative and positive northward temperature gradient above and below that, respectively, and in association with that, a negative MPVỹ region around the 40 km altitude, due to the minor SSW event in the stratosphere. TRPWs show upward and downward components of E-P fluxes above and below the 40 km altitude, respectively, with southward components of the fluxes, and significant E-P flux divergences in the stratosphere (Figure 12a). These features of E-P fluxes can work to reduce the negative MPVỹ values, meaning that TRPWs develop to dismiss the instability condition caused by the SSW event. The upward E-P fluxes above 40 km are associated with the convergence in the mesosphere around 80 km altitude (Figure 12a). Contributions by TRPW1Ws and TRPW1Es depicted in Figures 12b and 12c, respectively, show that the upward E-P fluxes and their convergences in the mesosphere are mainly due to TRPW1Ws, and that downward fluxes below 40 km are mainly due to TRPW1Es. Moreover, amplitudes of these waves (red contours) show that the TRPW1W has a tall vertical structure extending from the stratosphere to the mesosphere with the phase tilting westward with height (not shown), while TRPW1Es exist separately in the stratosphere and mesosphere and both have relatively short vertical structures tilting eastward (not shown).

On February 2, zonal-mean fields in Figure 12h shows that there are strong easterlies around the 75 km altitude, negative and positive northward temperature gradient above and below that, respectively, and a negative MPVỹ region in the mesosphere, due to the dissipation of TRPWs (Figures 10d and 12a) and STPWs (Figure 10c). TRPWs show upward and downward components of E-P fluxes above and below the 70 km altitude, respectively, with southward components of the fluxes, and significant E-P flux divergences in the lower mesosphere (Figure 12e). Contributions by TRPW1Ws and TRPW1Es depicted in Figures 12f and 12g, respectively, show that the downward E-P fluxes and their divergences in the lower mesosphere
are obviously due to TRPW1Es. Because downward E-P fluxes mean southward heat fluxes, it can be considered that this downward propagating TRPW1E develops to relax the positive temperature gradient below the easterly wind region to dismiss the BC instability. The double maximum in TRPW1E amplitudes (red contours) on January 27 turns into a large single maximum around the 50 km altitude on February 2 (Figure 12g).

3.3. Features in Other Winters

Analyses performed for the winter of 2015 are applied to three other winters when the PW cancellation was clear in Figure 6. Figure 13 shows daily variations of zonal-mean zonal winds $\vec{u}$ and E-P flux divergences DFs by PWs and TRPW1Es in winters of 2007, 2008, and 2016. It is commonly seen in these winters that easterly winds are formed in the mesosphere around January 28 due to the E-P flux convergences by PWs, which should be associated with the divergences in the stratosphere after minor SSWs (left column). These easterly winds are maintained in the mesosphere for more than 10 days, during which E-P fluxes by TRPW1Es diverge strongly in the lower part of the easterly wind region, although the timing is a little earlier than the easterly wind period in 2007 and 2008 (right column).

Further investigation shows that the TRPW1Es developing in the lower mesosphere propagate downward in northward temperature gradient (not shown) as in the case of 2015. When the easterly winds are formed
in the mesosphere and the polar air is warmed below the easterly wind region, zonal-mean fields fulfill the unstable condition, and downward propagating TRPW1Es develop to reduce the temperature anomaly and to dismiss the unstable condition. The westerly acceleration by the development of TRPW1Es would cancel out the easterly acceleration by the dissipation of STPWs and TRPW1Ws to maintain the easterly winds in the mesosphere.

During the easterly winds in the mesosphere, the vertical components of E-P fluxes (red contours) are enhanced in the beginning of February to converge in the stratosphere in all three winters (left column), which could break the balance of the PW cancellation and pull easterly winds in the mesosphere down into the stratosphere to cause minor SSWs again. In the recovery from the minor SSWs, PWs generated in the stratosphere around February 8, in 2007 and 2008, and February 11 in 2016 propagate upward and dissipate in the mesosphere to form easterly winds again. These processes involving minor SSWs could be repeated as clearly seen in 2008.

4. Conclusions

This study examined contributions of PWs to the general circulation in the winter middle atmosphere of NH where PW activities are strong, using satellite data of TIMED/SABER. We also used reanalysis data of MERRA-2 to verify our results.
It is revealed that there exists a cancellation mechanism in the high-latitude lower mesosphere between easterly acceleration by STPWs and westerly acceleration by TRPWs. When easterly winds appear in the mesosphere, E-P fluxes by STPWs propagating from lower levels converge in the lower mesosphere to accelerate easterly winds because STPWs cannot propagate in easterly wind regions. On the other hand, those by TRPWs diverge there due to the BC/BT instability to accelerate westerly winds. The TRPWs developing there appear as a zonal wavenumber 1 eastward TRPW with a period of 10–40 days propagating downward. Hence, we investigated anomaly fields by separating westward and eastward TRPW components of zonal wavenumber 1 and a period of 10–40 days, TRPW1W and TRPW1E, respectively, for the winter in 2015. Results indicate that TRPW1Ws developing in the stratosphere propagate upward and contribute to form the easterly winds in the mesosphere by dissipating and accelerating easterlies. When the easterly winds are formed in the mesosphere and the polar air is warmed below the easterly wind region, the downward propagating TRPW1Es develop in the high-latitude lower mesosphere to dismiss the BC instability by eliminating warm temperature anomalies below the easterly wind region. The westerly acceleration by the development of this TRPW1E cancels out the easterly acceleration by TRPW1Ws and STPWs to maintain the easterly winds in the mesosphere for more than 10 days. These features were identified in three other winters (2007, 2008, and 2016) as well when the cancellation was clear.

This study suggested the importance of TRPWs developing in the stratosphere and mesosphere for the general circulation in the winter NH. On the other hand, the strength of upward propagation of PWs from the troposphere should be an important factor controlling the circulation in the middle atmosphere. If the strength is moderate as in 2015 or three other winters, easterly winds could be maintained for a long time, but if it is fairly strong for a certain period, the balance could be broken down and major SSWs might occur to change the circulation. In winters with major SSWs, long-lasting easterly winds in the stratosphere prevent PWs from propagating upward after the onsets, resulting in the less noticeable PW cancellation in the lower mesosphere.

As mentioned in Introduction, the importance of TRPWs has been reported in the onset and/or recovery of SSWs in the NH. The TRPW1Ws in this study would be identical to the westward TRPWs, which are known...
to develop after SSW onsets and dissipate to form easterly winds and the elevated stratopause in the upper mesosphere (Chandran, Collins, et al., 2013; Limpasuvan et al., 2012, 2016). The PW cancellation we are suggesting in this study is established after that to maintain the formed easterly winds in the mesosphere by opposite contributions by PWs. The TRPW1Es in this study propagate downward after SSWs, whereas eastward TRPWs of wavenumber 2 (TRPW2E) observed before the major SSW in 2009 propagate upward to form the easterly winds in the upper mesosphere (Iida et al., 2014). Our analyses using the SABER data indicated that the TRPW2E before the SSW 2009 propagated downward and upward in the mesosphere (as seen in Figure 12c), and the latter contributed to form the easterly winds in the upper mesosphere (not shown).

Although this study examined contributions by PWs alone, the importance of GWs is well known in the mesosphere. Simulation data from a GW-resolving GCM showed that there was large easterly acceleration by GWs above the polar night jet in both hemispheres during winter (Watanabe et al., 2008). Furthermore, Watanabe et al. (2009) reported that an unstable condition brought by the easterly acceleration due to GWs in the mesosphere is recovered by 4 days wave generations in the winter Southern Hemisphere. Sato and Nomoto (2015) also suggested a temporal cancellation between easterly acceleration by GWs and westerly acceleration by PWs in the NH winter mesosphere by using the same simulation data. In contrast, the current study revealed a cancellation mechanism between easterly acceleration by STPWs and westerly acceleration by TRPWs in the same region. Moreover, GWs are known to contribute to recover SSWs and to form the elevated stratopause after SSWs (Limpasuvan et al., 2016; Tomikawa et al., 2012). We have to elucidate a whole picture of our results by taking GWs into account.

Data Availability Statement

Version 2.0 retrieved profile data from TIMED/SABER are freely available from the SABER website (http://saber.gats-inc.com/data.php). MERRA-2 data were provided by NASA/GSFC/GMAO (https://goldsmr5.gesdisc.eosdis.nasa.gov/data/MERRA2).

Acknowledgments

This work was supported by JSPS KAKENHI Grant numbers JP20H01973, JP18H01270, and JP18H01280. Figures were prepared using the Grid Analysis and Display System (GrADS).

References

Andrews, D. G., Holton, J. R., & Leovy, C. B. (1987). *Middle Atmosphere Dynamics*. New York: Elsevier.

Andrews, D. G., & McIntyre, M. E. (1976). Planetary waves in horizontal and vertical shear: The generalized Eliassen-Palm relation and the mean zonal acceleration. *Journal of the Atmospheric Sciences*, 33, 2031–2048. https://doi.org/10.1175/1520-0469(1976)033<2031:EPWITH>2.0.CO;2

Bosilovich, M. G., Akella, S., Coy, L., Cullather, R., Draper, C., Gelaro, R., et al. (2015). MERRA-2: Initial evaluation of the climate. NASA Tech. Rep. series on global modeling and data assimilation, 43, NASA/TM-2015-104606. NASA. Retrieved from http://gmao.gsfc.nasa.gov/pubs/docs/Bosilovich2015.pdf

Chandran, A., Collins, R. L., Garcia, R. R., Marsh, D. R., Harvey, V. L., Yue, J., & de la Torre, L. (2013). A climatology of elevated stratopause eastward PWs in the middle and upper atmosphere following the stratospheric sudden warming event of January 2012. *Geophysical Research Letters*, 40, 1861–1867. https://doi.org/10.1002/2012GL053786

Charlton, A. J., & Polvani, L. M. (2007). A new look at stratospheric sudden warmings. Part I: Climatological and modeling benchmarks. *Journal of Climate*, 20, 449–469. https://doi.org/10.1175/JCLI3996.1

Fleming, E. L., Chandra, S., Barnett, J. J., & Corney, M. (1990). Zonal mean temperature, pressure, zonal wind and geopotential height as functions of latitude. *Advances in Space Research*, 10, 11–39. https://doi.org/10.1016/0273-1177(90)90386-E

Hayashi, Y. (1971). A generalized method of resolving disturbances into progressive and retrogressive waves by space fourier and time cross-spectral analyses. *Journal of the Meteorological Society of Japan*, 49, 125–128. https://doi.org/10.2151/jmsj1965.49.2_125

Iida, C., Hirooka, T., & Eguchi, N. (2014). Circulation changes in the stratosphere and mesosphere during the stratospheric sudden warming event in January 2009. *Journal of Geophysical Research - D: Atmospheres*, 119, 7104–7115. https://doi.org/10.1002/2013JD021252

Kawatani, Y., Hirooka, T., Hamilton, K., Smith, A. K., & Fujisawa, M. (2020). Representation of the equatorial stratopause semiannual oscillation in global circulation model simulations in the whole atmosphere community climate model. *Journal of Geophysical Research - D: Atmospheres*, 118, 1234–1246. https://doi.org/10.1002/2019JD034195

Lait, L. R. (1994). An alternative form for potential vorticity. *Journal of the Atmospheric Sciences*, 51, 1754–1759. https://doi.org/10.1175/1520-0469(1994)051<1754:AFAFPP>2.0.CO;2

Limpasuvan, V., Orsolini, Y. J., Chandran, A., Garcia, R. R., & Smith, A. K. (2016). On the composite response of the MLT to major sudden stratospheric warming events with elevated stratopause. *Journal of Geophysical Research - D: Atmospheres*, 121, 4518–4537. https://doi.org/10.1002/2015JD024401

Limpasuvan, V., Richter, J. H., Orsolini, Y. J., Stordal, F., & Kviissel, O.-K. (2012). The roles of planetary and gravity waves during a major stratospheric sudden warming as characterized in WACCM. *Journal of Atmospheric and Solar-Terrestrial Physics*, 78–79, 84–98. https://doi.org/10.1016/j.jastp.2011.03.004

Lu, X., Chu, X., Chen, C., Nguyen, V., & Smith, A. (2017). First observations of short-period eastward propagating planetary waves from the stratosphere to the lower thermosphere (110 km) in winter Antarctica. *Geophysical Research Letters*, 44, 10744–10753. https://doi.org/10.1002/2017GL075641
Lu, X., Chu, X., Fuller-Rowell, T., Chang, L., Fong, W., & Yu, Z. (2013). Eastward propagating planetary waves with periods of 1-5 days in the winter Antarctic stratosphere as revealed by MERRA and lidar. *Journal of Geophysical Research - D: Atmospheres*, 118, 9565–9578. https://doi.org/10.1002/jgrd.50717

Lu, X., Wu, H., Chu, X., Oberheide, J., Mlynczak, M. G., & Russell, J. M., III. (2019). Quasi-biennial oscillation of short-period planetary waves and polar night jet in winter antarctica observed in SABER and MERRA-2 and mechanism study with a quasi-geostrophic model. *Geophysical Research Letters*, 46, 13526–13534. https://doi.org/10.1029/2019GL084759

Manney, G. L., Orsolini, Y. J., Pumphrey, H. C., & Roche, A. E. (1998). The 4-day wave and transport of UARS T racers in the Austral polar vortex. *Journal of the Atmospheric Sciences*, 55, 3456–3470. https://doi.org/10.1175/1520-0469(1998)055<3456:TDWATR>2.0.CO;2

Manney, G. L., & Randel, W. J. (1993). Instability at the winter stratopause: A mechanism for the 4-day wave. *Journal of the Atmospheric Sciences*, 50, 3928–3938. https://doi.org/10.1175/1520-0469(1993)050<3928:ISATWS>2.0.CO;2

Muller, H. G., & Nelson, L. (1978). A travelling quasi 2-day wave in the meteor region, (40 pp. 761–766). Elsevier. https://doi.org/10.1016/0021-9169(78)90136-8

Pancheva, D., Mukhtarov, P., Mitchell, N. J., Merzlyakov, E., Smith, A. K., Andonov, B., et al. (2008). Planetary waves in coupling the stratosphere and mesosphere during the major stratospheric warming in 2003/2004. *Journal of Geophysical Research*, 113, D12105. https://doi.org/10.1029/2007JD009011

Plumb, R. A. (1983). Baroclinic instability of the summer mesosphere: A mechanism for the quasi-two-day wave? *Journal of the Atmospheric Sciences*, 40, 262–270. https://doi.org/10.1175/1520-0469(1983)040<0262:BIOSIMS>2.0.CO;2

Plumb, R. A. (2002). Stratospheric transport. *Journal of the Meteorological Society of Japan*, 80, 793–809. https://doi.org/10.2151/jmsj.80.793

Randel, W. J. (1987). The evaluation of winds from geopotential height data in the stratosphere. *Journal of the Atmospheric Sciences*, 44, 3097–3120. https://doi.org/10.1175/1520-0469(1987)044<3097:TEOWFG>2.0.CO;2

Randel, W. J. (1994). Observations of the 2-day wave in NMC stratospheric analyses. *Journal of the Atmospheric Sciences*, 51, 306–313. https://doi.org/10.1175/1520-0469(1994)051<0306:O2DWA>2.0.CO;2

Rensberg, E. E., Marshall, B. T., García-Comas, M., Krueger, D., Lingenfelser, G. S., Martin-Torres, J., et al. (2008). Assessment of the quality of the Version 1.07 temperature-versus-pressure profiles of the middle atmosphere from TIMED/SABER. *Journal of Geophysical Research*, 113, D17101. https://doi.org/10.1029/2008JD009013

Siskind, D. E., Eckermann, S. D., McCormack, J. P., Coy, L., Hoppel, K. W., & Baker, N. L. (2010). Case studies of the mesospheric response to recent minor, major, and extended stratospheric warmings. *Journal of Geophysical Research*, 115, D00N03. https://doi.org/10.1029/2010JD014114

Smith, A. K., Garcia, R. R., Moss, A. C., & Mitchell, N. J. (2017). The semiannual oscillation of the tropical zonal wind in the middle atmosphere derived from satellite geopotential height retrievals. *Journal of the Atmospheric Sciences*, 74, 2413–2425. https://doi.org/10.1175/JAS-D-17-0067.1

Stray, N. H., Orsolini, Y. J., Espl, P. J., Limpasuvan, V., & Hibbins, R. E. (2015). Observations of planetary waves in the mesosphere-lower thermosphere during stratospheric warming events. *Atmospheric Chemistry and Physics*, 15, 4997–5005. https://doi.org/10.5194/acp-15-4997-2015

Tomikawa, Y., Sato, K., Watanabe, S., Kawatani, Y., Miyazaki, K., & Takahashi, M. (2012). Growth of planetary waves and the formation of an elevated stratopause after a major stratospheric sudden warming in a T213L256 GCM. *Journal of Geophysical Research*, 117, D16101. https://doi.org/10.1029/2011JD017243

Venne, D. E., & Stanford, J. L. (1979). Observation of a 4-day temperature wave in the polar winter Stratosphere. *Journal of the Atmospheric Sciences*, 36, 2413–2425. https://doi.org/10.1175/1520-0469(1979)036<2413:OATWTP>2.0.CO;2

Watanabe, S., Kawatani, Y., Tomikawa, Y., Miyazaki, K., Takahashi, M., & Sato, K. (2008). General aspects of a T213L256 middle atmosphere general circulation model. *Journal of Geophysical Research*, 113, D12110. https://doi.org/10.1029/2008JD009013

Watanabe, S., Tomikawa, Y., Sato, K., Kawatani, Y., Miyazaki, K., & Takahashi, M. (2009). Simulation of the eastward 4-day wave in the Antarctic winter mesosphere using a gravity wave resolving general circulation model. *Journal of Geophysical Research*, 114, D16111. https://doi.org/10.1029/2008JD009013