1. Introduction

The Northern extratropical stratosphere shows large variations over intraseasonal and interannual time scales during winter (e.g., Labitzke 1982). The large variations reflect the occurrence of anomalously weak/warm or strong/cold states of the stratospheric polar vortex.

Some of such anomalous vortex states correspond to stratospheric sudden warming events (SSWs) and vortex intensification events (VIs). SSWs are a spectacular phenomenon, where the stratospheric polar vortex largely deforms and sometimes even breaks down (e.g., Andrews et al. 1987). The deformation and/or breakdown of the vortex accompany a rapid increase in polar temperatures and a weakening/reversal of the westerly polar night jet. A SSW that accompanies a reversal of the polar night jet is classified as a major SSW (MSSW). In contrast, VIs are basically a mirror image of SSWs: the gross behavior of the stratospheric polar vortex during VIs is opposite in sign to that during SSWs but similar in shape (Limpasuvan et al. 2005). However, it is also shown that the wind and temperature signals of the VIs are generally weaker than those of SSWs.

Anomalous vortex states are also captured in a framework of the so-called polar-jet oscillation (PJO), which is based on a more statistics-oriented approach, such as an empirical orthogonal function (EOF) anal-

On the Asymmetry of Forecast Errors in the Northern Winter Stratosphere between Vortex Weakening and Strengthening Conditions

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Abstract

This study investigates the asymmetry of forecast errors in the Northern winter stratosphere between vortex weakening and strengthening conditions. Previous studies suggest that the stratospheric forecast errors for medium-range time scales of about two weeks are larger for weakening conditions of the polar vortex than for strengthening conditions even though they have anomalies of similar magnitudes.

We explore the asymmetry by comparing the one-month hindcast data of the Japan Meteorological Agency to the Japanese 55-year Reanalysis data. We define the vortex weakening and strengthening conditions of anomalies of similar magnitudes using an empirical orthogonal function analysis of polar stratospheric temperatures. Results indicate that the larger forecast errors in the stratosphere for the vortex weakening conditions originate from those of the planetary wave forcing in the upper troposphere. In particular, it is more difficult to forecast planetary wave amplification that leads to the vortex weakening conditions than forecasting wave attenuation that leads to the vortex strengthening conditions.

Examining forecast errors between major stratospheric sudden warming events (MSSWs) and vortex intensification events (VIs) defined by the zonal mean zonal wind in the high latitude stratosphere, we also show that it is more difficult to forecast the MSSWs than the VIs and further discuss the cause for the difference.

Keywords stratospheric predictability; vortex weakening; vortex strengthening

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ysis, applied to extratropical stratospheric variability (Kuroda and Kodera 1999, 2001; Hitchcock et al. 2013). The PJO is characterized by the poleward and downward propagation of zonal mean zonal wind anomalies, sometimes extending throughout a winter season. The PJO signal can be extracted for daily data, but are more visible in low-frequency components.

Several studies have investigated possible relationships between these fast (SSWs and VIs) and slow (PJO) variations of the polar vortex in the stratosphere. Kodera et al. (2000) show that SSWs tend to occur for a specific phase of the PJO and suggested that the PJO plays a preconditioning role for SSWs. Kuroda (2008) also suggests that the time evolution of the PJO includes SSWs and VIs as parts in its stages.

In recent times, numerous studies have investigated the predictability of the extratropical stratosphere (e.g., Tripathi et al. 2014). Such studies are partly motivated by a suggestion that knowledge of stratospheric conditions may be useful to improve tropospheric weather forecasts for extended-range time scales (Baldwin and Dunkerton 2001). In particular, Taguchi (2014a) investigates basic characteristics of the stratospheric predictability in one-month ensemble hindcast (HC) experiment data of the Japan Meteorological Agency (JMA). Taguchi suggests that the forecast errors of the HC data for the Northern winter stratosphere are asymmetric in sign of the vortex strength and its tendency. Moreover, the errors tend to be larger when the vortex is weaker than normal and/or is under deceleration. Inatsu et al. (2015) show a relevant result stating that the ensemble spread of the JMA operational forecast data in the Northern winter stratosphere tends to be larger when the vortex is weakening.

This study seeks to investigate in more detail how the asymmetry in the forecast errors in the Northern winter stratosphere occurs between vortex weakening and strengthening conditions, using the Japanese 55-year Reanalysis (JRA-55) and HC data. The present investigation further explores the possible role of planetary wave forcing from the troposphere in inducing the asymmetry. We mainly define the two vortex conditions in the PJO framework using slowly-varying principal components of an EOF analysis applied to polar stratospheric temperature anomalies. The EOF framework facilitates choosing specific conditions of interest, such as amplitude and phase of stratospheric anomalies. We also examine the asymmetry between MSSWs and VIs defined by the zonal mean zonal wind in the extratropical stratosphere, since they have attracted much attention (e.g., Tripathi et al. 2014). Whereas Taguchi (2014a) overviews the stratospheric predictability in the HC data and suggests the asymmetry, it leaves detailed features and cause/mechanism for the asymmetry open. This study seeks to deal with such questions about the asymmetry.

We widely use the word “strengthening” when the polar night jet is stronger (the vortex is colder) than normal and/or is accelerating. It includes specific conditions defined with results from the EOF analysis. In contrast, we use the word “intensification” (in VIs) to more specifically refer to a counterpart of MSSWs, following the convention of Limpasuvan et al. (2005).

The rest of the paper is organized as follows. Section 2 explains the data analyzed in this study. Section 3 describes results when the vortex conditions are defined with the EOF analysis. Section 4 provides summary and discussion. Appendix includes results for MSSWs and VIs, since they are easier to define than EOFs and already widely used.

2. Data and verification measures

2.1 Data
This study compares the JRA-55 and HC data to examine forecast errors in the Northern winter stratosphere. We utilize daily averages of the JRA-55 data from 1979 to 2012 to represent the real world. The reader is referred to Kobayashi et al. (2015) for details of the JRA-55 data. The climatology is obtained from the 34 year means, and anomalies (denoted by a subscript “a”) are defined as deviations from the climatology. The period of the JRA-55 data is somewhat longer than the period of the HC data described below. The ending year 2012 (close to the latest year for which the data are available) is used to increase the data period.

We also utilize daily averages from the HC experiments conducted by the JMA. The experiments are based on the March 2011 version of the JMA one-month ensemble forecast system. The system includes a global atmospheric model with a horizontal resolution of approximately 110 km and 60 levels up to approximately 0.1 hPa. The forecasts in the experiments are initialized on the 10th, 20th, and last day of each month for 31 years from 1979 to 2009. A set of five ensemble runs are conducted for 34 days from each initial date. Further details of the HC data can be found in Taguchi (2014a).

2.2 Verification measures
We mainly use two measures to verify the HC data. One of the two measures is a bias (denoted by \( \Delta \)),
defined as a difference of the ensemble mean field of a HC set from the JRA-55 counterpart as follows:

$$\Delta = A_{m}(t,i) - A_{mJRA55}(t,i).$$

Here A: quantity of interest, m: index for HC sets, t: forecast time in day (t = 0 day corresponds to the initial day), i: index for spatial gridpoints, and EM: ensemble mean.

The second measure is a root-mean-square error (RMSE) defined as follows (Taguchi 2014a):

$$\text{RMSE}_{m}(t) = \sqrt{\frac{1}{I} \sum_{i=1}^{I} (\Delta A_{m}(t,i))^2}.$$

In particular, RMSE is the root of a spatial average in the extratropical stratosphere (poleward of 20°N, I denotes the number of gridpoints for the region) of the squared bias. We use the geopotential height at 10 hPa (Z10) as A: RMSE in this case is referred to as RMSE$_{Z10}$.

3. Analysis based on an EOF analysis of polar temperatures

In this section, we describe results when vortex weakening and strengthening conditions are defined in the PJO framework using an EOF analysis.

3.1 Method

To extract extratropical stratospheric variability, we apply an EOF analysis to vertical profiles of daily polar temperature anomalies (averaged poleward of 70°N as in Hitchcock et al. 2013) above 100 hPa in the JRA-55 data for an extended winter season from November to March (NDJFM). Figure 1a plots the vertical profiles of the two leading EOFs. They
explain almost all (95.0 %) of the total variance (62.9 % by the first mode and 32.1 % by the second mode). Then, the principal components (PCs) are obtained by projecting, onto the two EOFs, daily polar temperature anomalies that are low-passed with a seven-day running mean. The seven-day running mean is applied so as to capture smooth evolutions of the extratropical variability by removing day-to-day fluctuations.

An examination of the PCs in the phase space shows a strong tendency for the data points to rotate in the anticlockwise direction (Fig. 1b). Together with the vertical profiles of the two EOFs, this tendency confirms the familiar downward propagation of polar temperature anomalies that are low-passed with a seven-day running mean. The seven-day running mean is applied so as to capture smooth evolutions of the extratropical variability by removing day-to-day fluctuations.

3.2 Results

Figure 2a is a similar scatter plot between the two PCs to Fig. 1b, but only for a subset of the PC data obtained for the JRA-55 data. The subset comprises the PC data 14 days after each initial date of the HC data that ranges from 11/20 (November 20th) to 2/10 (February 10th) for the 31 years. The 14 day time lag is used because the forecast errors of the HC data in the Northern winter stratosphere have large case-to-case variations for such time scales (Taguchi 2014a). The time window for the initial dates (from 11/20 to 2/10) does not match the DJF period but is somewhat shifted backward in time, so that the target days for the PC data (14 days after the initial dates) correspond to the DJF period more closely. Each data point is differently colored by the value of RMSEZ10 averaged for $t = 14 \pm 3$ days. In other words, whereas the location of each data point is based on the observed state of the PJO, the color of the data point represents RMSEZ10 of the 14-day forecast for the PJO state. Figure 2b shows composite values of RMSEZ10 for two-dimensional bins with a bin width of 25 for each axis.

Figure 2 thus demonstrates that the forecast errors or RMSEZ10 depend on the phase angle of the data.

![Fig. 2. (a) As in Fig. 1b, but only plotting the data 14 days after each initial date of the HC data ranging from 11/20 to 2/10. Each data point is differently colored by the value of RMSEZ10 (m) averaged for $t = 14 \pm 3$ days (see the colorbar). Panel (b) shows composite results for two dimensional bins with a bin width of 25 for each axis. The two groups S and W are denoted for further investigation. A star symbol in (a) indicates that a MSSW key day exists in a 15 day time window (before seven days to after seven days) around the target days. A big star is used if the key day exists in a one week window. Crosses are the same, but for the VIs.](image-url)
points even when they have similar amplitudes (distances from the origin). One sees that RMSE$_Z$$_{10}$ tends to be largest around the first quadrant (approximately 0°–90°). This quadrant corresponds to when the polar temperature anomalies are positively large near 10 hPa (Figs. 1c, d). RMSE$_Z$$_{10}$ also depends on the distance of the data points from the origin. Namely, it tends to be large when the distance is large. In contrast, RMSE$_Z$$_{10}$ tends to be smallest for the second and third quadrants (approximately 90°–270°). In these quadrants, the polar temperature anomalies decrease and are negatively large near 10 hPa (Figs. 1c, d). The dependence of the forecast errors on the phase angle agrees with the asymmetry pointed out by Taguchi (2014a).

Figure 2a includes the information on the relationship of the EOF-based target days (data points plotted in the figure) to key days of MSSWs or VIs (defined in Appendix). Such a relationship is discussed in Section 4.2.

To further explore the asymmetry, we define two contrasting groups (S and W, for strong and weak vortex conditions, respectively) and examine differences between them. The two groups are denoted in Fig. 2. They are defined to represent when polar temperature anomalies are largest in the middle stratosphere near 10 hPa. The two groups consider only data points that have similar distances from the origin. The groups S and W comprise 28 and 24 HC sets, respectively.

Figures 3a, b plot time series of anomalous zonal mean temperature $[T]_a$ (70°N–90°N and 10 hPa) and zonal wind $[U]_a$ (60°N and 10 hPa) in the JRA-55 data as a function of forecast time of the HC data for the two groups. Here the square brackets denote the zonal mean. The 70°N–90°N band is used here for
These anomalies in the zonal mean states are well related to those in planetary wave activity entering the stratosphere (Fig. 3c), as Polvani and Waugh (2004) show that the Northern annular mode index at 10 hPa is highly correlated with poleward eddy heat flux at 100 hPa particularly when the latter is cumulated or averaged in time for a few weeks or longer. Here the vertical component of the Eliassen-Palm (EP) flux (FZ) of waves 1–3 in 40°N–90°N and 100 hPa is used as a measure of the wave activity entering the stratosphere. Note that FZ is proportional to the poleward eddy heat flux. The 40°N–90°N band is used to facilitate the budget analysis of the EP flux below, as the

[T]a as in the EOF analysis of the polar temperatures, whereas the 60°N latitude is used for [U]a since it is often used, e.g., in defining MSSWs (Appendix). Note that the figure plots the JRA-55 data but the time axis is measured with respect to each relevant initial date. The figure also includes the composite and 95% confidence intervals for each group based on a Student t test. Ellipses denote representative distributions determined by an EOF analysis. The magnitude of each ellipse corresponds to one standard deviation of each PC of the EOF analysis.

Fig. 4. Scatter plots between the anomalous FZ of waves 1–3 in 40°N–90°N and 100 hPa averaged from $t = -10$ to 14 days (x-axis) and each of the four quantities (y-axis) for the groups S (closed circles) and W (open circles) in the JRA-55 data. Panels (a,c) plot the anomalous polar temperature in 70°N–90°N and 10 hPa and zonal mean zonal wind at 60°N and 10 hPa for $t = 14 \pm 3$ days, respectively. Panels (b,d) plot their time changes from $t = -10 \pm 3$ to $14 \pm 3$ days. Crosses are mirror images of the closed circles. Horizontal and vertical lines denote 95% confidence intervals for each group based on a Student t test. Ellipses denote representative distributions determined by an EOF analysis. The magnitude of each ellipse corresponds to one standard deviation of each PC of the EOF analysis.
EP flux vanishes at 90°N. It also considers the equatorward extension of FZ. The results are robust irrespective of the exact choice of the band (not shown), consistent with Polvani and Waugh (2004). Here FZ for each day is averaged in time for 25 days prior to the day (inclusive) to consider the importance of cumulating or averaging the heat flux in time. The results are also insensitive to the choice of the averaging period of 25 days. The contribution from waves 1–3 are considered, since they are the main components that can propagate to the stratosphere (Charney and Drazin 1961). The FZ anomalies are negative for S and positive for W when the zonal mean anomalies are large in magnitude.

Time evolutions of $\text{RMSE}_{\text{Z10}}$ for the two groups are plotted in Fig. 3d. As expected, it is close to zero near $t = 0$ day and increases with time. $\text{RMSE}_{\text{Z10}}$ tends to experience more rapid increases after the initialization and have larger values for W than for S. The 95% intervals are separated for approximately $t = 10–20$ days.

Since it is shown that the planetary wave activity in the lower stratosphere is a key for the stratosphere, Fig. 4 further examines relationships of the FZ anomalies in 40°N–90°N, 100 hPa with stratospheric anomalies at 10 hPa for the two groups. The FZ anomalies are averaged for 25 days from $t = -10$ to +14 days. Figures 4a, c plot the zonal mean temperature and zonal wind anomalies for $t = 14 \pm 3$ days, respectively. It is shown that the distributions for S and W largely overlap each other when the data points for S are reversed in sign to consider the sign difference. Such overlaps also occur in Figs. 4b, d when time changes in the temperature and wind anomalies for
the 25-day window (from $t = -10 \pm 3$ to $-14 \pm 3$) are used.

In contrast, the two groups are different in terms of the forecast errors (Fig. 5). Using the temperature and zonal wind biases for $t = 14 \pm 3$ days in the $y$-axis, Figs. 5a, b show that the distributions and 95 % intervals of these biases are separated between S and W, whereas they are similar in the FZ anomalies ($x$-axis). Such a separation is also the case with RMSE$_{210}$ for $t = 14 \pm 3$ days (Fig. 5c). The bias of FZ at 100 hPa is also different between S and W in terms of the 95 % intervals (Fig. 5d). Note that the anomaly and bias of FZ tend to be opposite in sign.

Figure 6 summarizes anomalies (Fig. 6a) and biases (Fig. 6b) of the EP flux budget in the extratropical region (40°N–90°N and 10–100 hPa) for the two groups. The budget considers the EP flux exiting from the region (FZ at 10 hPa and meridional component FY at 40°N) in addition to that entering it (FZ at 100 hPa) to examine the possible roles of the former terms. The budget calculations are performed by integrating each EP flux component in distance (latitude or height): FZ at 10 or 100 hPa is integrated from 40 to 90°N, whereas FY at 40°N from 100 to 10 hPa. The wave driving is calculated as a net of the three fluxes (a positive value of the wave driving corresponds to a net inward flux). All numbers for S are reversed in sign in Fig. 6 to facilitate the comparison.

The figure thus demonstrates that the 95 % intervals of the anomalies of the EP flux and wave driving roughly overlap between the two groups. It is also noted that the anomalous FZ mainly determines the anomalous wave driving, whereas FY plays a minor role.

In contrast, it is important to notice that the biases of the EP flux and wave driving are different in magnitude between S and W. In particular, the FZ bias at 100 hPa has the larger composite difference than
the other fluxes. The composite difference of the 100 hPa FZ bias is comparable to that of the wave driving bias. When one examines FZ in the upper troposphere, such as 300 hPa, it shows similar features in the anomaly and bias to FZ at 100 hPa.

To further understand upper tropospheric features leading to the FZ anomalies and biases for the two groups, Fig. 7 plots geographical patterns of anomalies and biases of the geopotential height of waves 1–3 at 300 hPa. Here time means are taken for a shorter window of $t = 0$ to 14 days to emphasize the biases. The anomalous wave patterns (Figs. 7b, c) tend to negatively or positively interfere with the climatology (Fig. 7a) for S and W, respectively. In particular, the anomalies for S have positive values around 150°E, which is statistically significant (significantly different from zero) at the 95 % level according to a Student t test. This leads to the negative interference, weakening the climatological trough around the longitude. In contrast, the group W has similarly signed anomalies to the climatological waves (particularly wave 1), leading to the positive interference. Such negative interference for S leads to weakened wave activity entering the stratosphere, which yields the colder- and stronger-than-normal polar vortex in the stratosphere. The situation is opposite for W. The importance of such interference in modulating wave activity entering the stratosphere, and hence the strength of the polar vortex in the stratosphere is pointed out by previous studies (e.g., Garfinkel et al. 2010; Fletcher and Kushner 2011; Nishii et al. 2011).

It is notable that the height biases roughly have opposite signals to the anomalies in each group (Figs. 7d, e). For the example of W, the bias has negative values near 0°E and positive values near 180°E, which are opposite in sign to the anomalies. In other words, the HC data tend to underestimate the

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**Fig. 7.** (a) Climatological DJF stationary waves 1–3 in the 300 hPa geopotential height. Panels (b,c) plot anomalous 300 hPa height in the JRA-55 data composited for the groups S and W ($t = 0$–14 days), respectively. Panels (d,e) are similar to (b, c), but for composite bias. Contour interval is 40 m in (a) and 20 m in (b–e). Negative values are shaded. Dots in (b–e) indicate that the values are significantly different from zero at the 95 % level. The 0°E longitude is located at the bottom in each panel.
observed weakening (for S) or strengthening (for W) of waves 1–3.

Since the negative values near 0°E for W are the most notable features in the height biases, we compare all cases for S and W in a scatter plot (Fig. 8). The scatter plot is made between anomalies and biases of the 300 hPa height of waves 1–3 over the region (30°W–30°E, 50°N–70°N). A notable feature is that two outliers in the vertical direction exist for W, with anomalies above +150 m and biases below −100 m. Such outliers are absent from S. The first outlier (with the bias below −200 m) corresponds to the case initialized on 1987/1/1. The second outlier case (with the bias below −100 m) is initialized on 1992/1/10.

It follows from the analysis procedure that the 300 hPa height features for the two outliers (Fig. 9) confirm the composite features in Fig. 7. The two cases are characterized by positive height anomalies of waves 1–3 near 0°E (Figs. 9c, f). This feature corresponds to the flow in the total height field that meanders poleward (Figs. 9b, c; e.g., see thick contours drawn at 9000 m). The bias plots also confirm that such positive anomalies (or meandering flow) are underestimated in the HC data (Figs. 9g, h).

4. Summary and discussion

4.1 Summary

This study has investigated the asymmetry in forecast errors in the Northern winter stratosphere between weakening and strengthening conditions of the polar vortex. Previous studies suggest the asymmetry, or larger forecast errors for weakening conditions of the vortex (Taguchi 2014a; Inatsu et al. 2015), but leave the detailed features and cause/mechanism for it open. To this end, we examine the JMA one-month HC experiment data in comparison to the JRA-55 data. We mainly define the two contrasting vortex conditions using the antiphase domains in the PJO framework or phase space of the EOF analysis of polar stratospheric temperatures (Section 3). We also define the two vortex conditions by MSSWs and VIs using the zonal mean zonal wind in the extratropical stratosphere (Appendix).

Results show that whereas the vortex weakening and strengthening conditions defined with the EOF analysis have similar magnitudes of the anomalous zonal mean states in the stratosphere and the anomalous wave forcing from the troposphere, the weakening conditions experience larger forecast errors of these quantities. This indicates that planetary wave amplification in the troposphere leading to the vortex weakening conditions is more difficult to forecast than wave attenuation leading to the strengthening conditions.

We show that the asymmetry also holds for the MSSWs and VIs. The HC data show larger errors for the MSSWs than for the VIs, when they are defined to have mean zonal wind anomalies of similar magnitudes around the key day. In addition to the above factor, this is contributed by the fact that the MSSWs experience stronger decelerations of the stratospheric jet, and hence stronger wave forcing anomalies from the troposphere, reflecting a stronger-than-normal preceding stratospheric wind condition.

4.2 Discussion

The above results are based on the two definitions of the contrasting vortex conditions, whereas previous studies suggest an in-phase relationship between the PJO and MSSWs/VIs (Kodera et al. 2000; Kuroda 2008). Such a relationship is seen in Fig. 2. It uses a star symbol, when a MSSID key day exists in a 15 day time window around the target day. Crosses are similarly used for VIs. As expected, the stars are mainly located in the first quadrant when polar temperature anomalies are largely positive at 10 hPa (Figs. 1c, d),
whereas the crosses mainly exist in the third quadrant. The stars are located farther from the origin than the crosses. This implies that the MSSWs accompany larger polar temperature anomalies than do the VIs, although they are defined to have anomalous zonal mean zonal wind of similar magnitudes at 60°N and 10 hPa around the key days.

Whereas this study compares the average pictures of the two contrasting conditions of the stratospheric polar vortex, it is suggested that stratospheric predict-
ability, e.g., during MSSWs, varies from one case to another as seen in Figs. 2a, A2, and A3 (e.g., Taguchi 2014a, b; Tripathi et al. 2014). Such case-to-case variations need further study. We are conducting such an analysis and will report the results in a separate paper.

Exploring the asymmetry in stratospheric forecast errors, this study attributes the larger errors for the vortex weakening conditions including the MSSWs to the difficulty in forecasting planetary wave amplification in the troposphere. This leads to a next problem, how/why planetary wave amplification in the troposphere is more difficult to forecast than planetary wave attenuation even when they lead to anomalous planetary wave forcing (FZ) of similar magnitudes entering the stratosphere. This will be relevant to the widespread recognition that tropospheric blocking events are difficult to forecast (e.g., Tibaldi and Moteni 1990; Kimoto et al. 1992; Pelly and Hoskins 2003). In fact, the two cases, which have strongly positive anomalies and strongly negative biases around 0°E (Fig. 9), correspond to the occurrence of blocking events over a Euro-Atlantic sector (e.g., Taguchi 2008).

The asymmetry examined here means that strengthening conditions of the stratospheric polar vortex can be forecasted better than weakening conditions. Since previous studies show that both anomalous conditions in the stratosphere exert effects on the troposphere (Baldwin and Dunkerton 2001; Kuroda 2008; Lipps suvan et al. 2005), this suggests that tropospheric anomalies associated with strengthening conditions of the stratospheric polar vortex (VIs) can be forecasted better. This suggestion can be explored in the future study, whereas the possible importance of SSWs on improved tropospheric weather forecasts has drawn much attention (e.g., Mukougawa et al. 2009; Sigmond et al. 2013).

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Appendix

In Appendix, we briefly describe results when we use MSSWs and VIs as two contrasting conditions of the stratospheric polar vortex.

### A.1 Method: Definitions of MSSWs and VIs

We adopt a similar definition of MSSWs as outlined in Charlton and Polvani (2007), who follow the adaptation of the World Meteorological Organization definition of MSSWs. In short, a MSSW is detected when the zonal mean zonal wind at 60°N and 10 hPa changes its direction from a westerly wind to an easterly wind (in particular, the wind drops below 0 m s\(^{-1}\)) during the DJF period. The key day of the MSSW is defined as the first day on which the zonal wind is easterly. The key day is also referred to as lag = 0 day. To identify two MSSWs for
one season, their key days must be separated by 20 days or more. Thus, this study identifies 20 MSSWs during DJF for the entire 34 years from 1979 to 2012.

In contrast, we identify VIs using a 60 m s$^{-1}$ threshold of the zonal mean zonal wind at 60°N and 10 hPa. In particular, the key day of each VI (also denoted as lag = 0 day) is the first day on which the wind rises above 60 m s$^{-1}$. The value of 60 m s$^{-1}$ is determined so that the MSSWs and VIs have mean zonal wind anomalies of similar magnitudes from the DJF climatology (approximately 30 m s$^{-1}$) around the respective key days. The median of the daily zonal wind during DJF is approximately 33.0 m s$^{-1}$. The 20 day separation is also used to identify two VIs for one season, as is the case for the MSSWs. As a result, we obtain 11 VIs in the 34 years.

It is noted that the present definitions of the MSSWs (SSWs) and VIs are different from those of Limpasuvan et al. (2004, 2005). They use ±1 standard deviation thresholds of the leading PC time series of an EOF analysis applied to the daily zonal mean zonal wind at 50 hPa and obtain 39 SSWs and 41 VIs for 44 years (November to April) from 1958 to 2001. We adopt the simpler 0 and 60 m s$^{-1}$ thresholds of the zonal wind at 60°N and 10 hPa, because the MSSW threshold is widely used (Charlton and Polvani 2007).

A.2 Results

Figure A1a displays time series (with respect to the key day) of the zonal mean zonal wind at 60°N and 10 hPa for the MSSWs and VIs defined above. By definition, the zonal wind drops below 0 m s$^{-1}$ on lag = 0 day for the MSSWs and rises above 60 m s$^{-1}$ for the VIs. The composite zonal wind still decreases just after lag = 0 day for the MSSWs, whereas it is close to the peak on lag = 0 day for the VIs. The composite zonal wind well before the key day (such as lag < −25 days) is approximately 40 m s$^{-1}$ for both of the MSSWs and VIs. It follows that the wind deceleration for the MSSWs (approximately 40 m s$^{-1}$ from 40 to 0 m s$^{-1}$) is larger in magnitude than the wind acceleration for the VIs (approximately 20 m s$^{-1}$ from 40 to 60 m s$^{-1}$). It turns out that this difference in the wind changes is a key for that in the forecast errors between the MSSWs and VIs.

Figure A1b similarly plots FZ of waves 1–3 in the lower stratosphere for the MSSWs and VIs. In this figure, FZ on each day is averaged for preceding 25 days as in Fig. 3c. As expected, the MSSWs are associated with increased FZ around the key day, whereas the VIs are associated with decreased FZ.

The forecast errors in the stratosphere, or RMSE$_{Z10}$, for the MSSWs and VIs are examined in Fig. A1c. This figure uses the HC data that have lead times ranging from lag = −30 to 0 day of any one of the MSSWs and VIs. It is notable that the MSSWs have larger errors around lag = 0 day than do the VIs, although the time evolutions are different depending on the lead time.

Such a difference in RMSE$_{Z10}$ between the MSSWs and VIs is extracted in Fig. A2. It plots RMSE$_{Z10}$ averaged for lag = 0 ± 3 days (y-axis) for all HC data used in Fig. A1c. The x-axis denotes the lead time of each HC data. It is common between the MSSWs and VIs that RMSE$_{Z10}$ increases as the HC are initialized earlier, but there is also a systematic difference.
between them. RMSE_{Z10} is larger for the MSSWs than for the VIs, when we compare the HC data of similar lead times such as lag = −20 to −10 days.

Figures A3a, b examine the relationship of the tropospheric wave forcing and stratospheric zonal wind in the JRA-55 data for the MSSWs and VIs. One sees that, whereas the MSSWs and VIs are roughly similar in the magnitudes of the zonal wind anomalies around the key day as a result of their definitions, they are different in those of the FZ anomalies (Fig. A3a). Since the preceding zonal wind is similarly approximately 40 m/s before about lag = −20 days (Fig. A1a), this follows that the time changes in the zonal wind anomalies are different in magnitude (Fig. A3b).

Figures A3c, d show similar scatter plots between the anomalous FZ in the JRA-55 data and RMSE_{Z10} (Fig. A3c) or FZ bias (Fig. A3d). These panels use a HC set for each of the MSSWs or VIs that has a maximum lead time smaller than lag = −11 days. The errors are also larger for the MSSWs than for the VIs.

In summary, whereas the MSSWs and VIs are defined to have extratropical zonal wind anomalies of similar magnitudes, the wind deceleration for the MSSWs is stronger than the acceleration for the VIs, reflecting the preceding wind condition. This corresponds to the stronger wave forcing anoma-
lies required for the MSSWs. The stronger wave forcing anomalies are more difficult for the HC data to reproduce, leading to the larger errors (FZ bias and RMSEZ10) for the MSSWs.

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