Interannual Variations of Planetary Waves in the Southern Hemisphere Stratosphere

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Abstract

A statistical analysis is made of the interannual variability of stratospheric planetary waves in the Southern Hemisphere (SH) in late winter using NCEP/NCAR reanalysis data gathered over a 20 year period from 1979 to 1998. The study investigates the dynamical coupling between the planetary waves and the tropospheric circulation, by focusing on the activity of transient baroclinic disturbances.

With the aid of EOF analysis applied to the 10 hPa monthly mean height anomalies and tropospheric circulation patterns, it is found that the year-to-year variation of quasi-steady planetary waves in late winter (September and October) is characterized by a longitudinal phase shift of zonal wavenumber 1 as well as by a variation in wave amplitudes. An eastward (westward) shift of the stratospheric planetary waves corresponds to a double-jet (single-jet) structure of the upper tropospheric (300 hPa) zonal winds. On the other hand, planetary wave amplitude variation is closely related to that of the meridional heat flux in the upper troposphere.

These results are reinforced by the Eliassen-Palm flux diagnosis composed of the stratospheric EOF scores, and hence the effect of tropospheric transient waves on the forcing to the stratospheric planetary waves is stressed.

1. Introduction

Mainly as a result of the abundance of satellite data gathered during the last two decades, stratospheric observations for both hemispheres have now become available and notable differences in dynamical features between the two hemispheres have become apparent, not only in their seasonal variations but also in their year-to-year variabilities.

For stratospheric circulation in the Southern Hemisphere (SH), Shiotani et al. (1993) made the first analysis on the interannual variation of the seasonal evolution of mean zonal winds and planetary waves, by using a TIROS-N SSU dataset covering the 10 year period from 1980

to 1989. They found that planetary waves in the SH stratosphere in late winter show a significant year-to-year variation in association with changes in the intensity and latitudinal profiles of the polar-night westerlies. They further showed, through the Eliassen-Palm (EP) flux diagnosis, that evidence exists for a dynamical coupling between the stratosphere and troposphere which is related to the mean zonal wind variation.

Following this early work, Aoki et al. (1996) and Aoki and Hirota (1998) made further statistical analyses on the relationships between the SH tropospheric circulation and stratospheric planetary waves in order to identify the characteristic features which are unique to the SH.

Regarding the quasi-steady planetary waves in the stratosphere as seen in the monthly mean geopotential height field in wintertime, it has been widely recognized that the strato-

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spheric wave is generated through an upward propagation of energy from the troposphere (e.g., Charney and Drazin 1961). In the NH, the forcing in the lower atmosphere is considered to be due mainly to the large-scale topographic effect: This can be easily understood from the fact that the geographic location of stationary planetary waves in the NH winter stratosphere is almost always fixed by the topography, as in the cases of the Aleutian High and the Greenland Low, for instance.

In the SH stratosphere, however, the interannual variation of planetary waves is rather larger than that in the NH, in both location (phase) and wave amplitude (Shiotani et al. 1993). Quintanar and Mechoso (1995) suggested that forcing from the Indian Ocean region is an important source of SH planetary waves, in addition to orographic and thermal forcing from the Antarctic continent. Note that this variability can be also seen in the year-to-year change of intensity and in the location of the so-called "Ozone Croissant" over the Antarctic continent in late winter (e.g., Kayano 1997).

In view of the fact that the surface topographies are quite different between the two hemispheres, it is important to examine both the generation mechanism and interannual variability for planetary waves in the SH stratosphere in terms of the dynamical coupling between the troposphere and the stratosphere. In particular, investigation of the contribution of transient, baroclinic disturbances (i.e., the location and intensity of "storm tracks") to the tropospheric forcing in the SH is of considerable interest.

Another important issue concerning the SH stratospheric circulation is the difference of the seasonal march from that observed in the NH. As was shown, for example, by Hirota et al. (1983), planetary wave amplitudes in the SH stratosphere have two maxima in early winter (May and June) and in late winter (September and October), with a minimum in mid-winter (July–August). A number of aspects of this behavior are still open to question, so in what follows we confine our attention to the period of maximum wave activity in late winter.

Keeping the above-mentioned differences between the two hemispheres in mind, in the present study we first make a statistical EOF

Fig. 1. Time evolution of wave 1 (a) amplitudes (through the annual) and (b) phases (from September to October) at 10 hPa, 65°S within the period 1979–1998.

analysis of the interannual variation of SH stratospheric circulation patterns over a 20 year period. We then discuss the relationship with the tropospheric circulation, and especially with the activity of short-period and synoptic-scale transient disturbances. Suggestions are given for the role of transient waves in the forcing of steady-state planetary waves.

2. Data

In this study we use the daily (12GMT) NCEP/NCAR reanalysis data for the 20-year period from 1979–1998 (see Kalnay et al. 1996 for details of the NCEP/NCAR reanalysis project and its products). The data are transformed into $5^\circ \times 5^\circ$ horizontal latitude and longitude mesh at 17 pressure levels (1000–10 hPa).

Figure 1(a) shows the annual march of wave 1 amplitudes at 10 hPa, 65°S for the 20-year period. The wave amplitudes show maxima in September and October and large interannual variability in terms of their growth timing, as mentioned by Shiotani et al. (1993). Thus,

monthly mean fields covering 40 months (Sept and Oct for 20 years) are used in making a statistical analysis to understand the structure and behavior of quasi-steady planetary waves in late winter.

Figure 1(b) shows the time series of wave-1 phase for the two months. We find the stationary component dominant and recognize no seasonal variation of phase during the period of the late winter (Sept and Oct). Thus it is valid to investigate the interannual variability of stationary waves with the 40 monthly mean datasets by ignoring the intraseasonal variations.

Throughout this paper over-bars ($\bar{A}$) denote monthly means and primes (A') represent deviations from the zonal mean.

3. Results of analysis

3.1 Monthly mean features

Figure 2 shows the SH monthly mean geopotential height anomaly (deviations from the zonal mean) at 10 hPa for September months in the period 1979–1998. It can be easily seen that the zonal wavenumber 1 component is the most prominent with phase and amplitude quite variable over the 20 year period. Similar features are seen for October (not shown here). The variability is represented in Fig. 3, in which both scatter diagrams and histograms of the phase and amplitude of the monthly mean wavenumber 1 component at 10 hPa are provided for (a) 65°S and (b) 65°N for the two winter months when planetary waves are active (Sept, Oct, for SH, and Jan, Feb, for NH). In the SH, the phase varies widely between 40° and 220° and is centered on about 150°, in contrast to the NH where the phase is almost locked around 180°. The amplitude is highly variable in both hemispheres.

Before making a statistical analysis let us give a general view of the connection between the tropospheric and stratospheric circulation of the two specific years. Figure 4 illustrates longitude-height cross sections of monthly mean geopotential height anomaly ($\bar{Z}'$) between 55°S and 65°S for September of 1989 and 1990. In Fig. 3, the data of these two years are plotted with open circles, showing that both phase and amplitude are quite different between the two years. As shown in Fig. 2, the quasi-stationary planetary wave in the stratosphere shifts eastward and has larger ampli-

tude in 1990 than in 1989. In the troposphere, the amplitude of the wavenumber 1 component in 1990 is large, reflecting the structure in the stratosphere, while in 1989 it is rather small.

The cases for the years 1989 and 1990 indicate different tropospheric circulation patterns in association with the stratospheric pattern. In the following sections we first examine the interannual variability of planetary waves in the SH stratosphere in a statistical manner. Then, we investigate how the connections between the troposphere and stratosphere differ in relation to this variability.

3.2 EOF of the stratospheric circulation

In this section we apply the empirical orthogonal function (EOF) analysis (see North et al. 1982 for details), on the basis of the unstandardized covariance matrix of monthly mean geopotential height anomaly at the 10 hPa level between 20°S and 90°S for the 40 months of our analysis period. Taking into account the difference in area on a sphere, monthly mean variability is weighted by the square root of the cosine of latitude.

Figure 5(a) gives the 40-month mean pattern (i.e., climatology) of the 10 hPa height deviation, in which the dominance of zonal wavenumber 1 is observed with the ridge and trough located at 150°E and 50°W respectively along the latitude belt of about 60–70°S.

With the long-term variations of this pattern, the first mode of the analysis (EOF1) shown in Fig. 5(b) indicates the east-west fluctuation of phase, since the dipole pattern is almost orthogonal to the climatological pattern. In the case that the standardized leading principal component of EOF1 (PC1) is positive (negative), the phase shifts westward (eastward) from the climatological pattern.

On the other hand, the pattern of EOF2 (Fig. 5(c)) is almost in phase with Fig. 5(a), which indicates the variation of wave amplitude. Positive (negative) values of PC2 denote the large (small) wave amplitude. Both PC's explain large fractions of the total variance (59% for PC1 and 29% for PC2), showing that the interannual variability of planetary waves in the SH stratosphere is characterized by longitudinal phase shifts and amplitude fluctuations.

The representation shown in Fig. 6 gives scatter diagrams and histograms of the longi-

Fig. 2. Monthly mean height anomaly of $\overline{Z'}$ (unit in meter) at 10 hPa for September months between 1979–1998. The outer circle corresponds to the the 20°S parallel. The contour interval is 200 m and negative values are shaded.

Fig. 3. Scatter diagrams for monthly mean wave 1 at 10 hPa, (a) 65°S and (b) 65°N on the phase and amplitude plane. The data consist of (a) September and October, (b) January and February, of 1979–1998. The upper and right panels are histograms of phase and amplitude respectively. Open circles in (a) indicate the September of 1989 (lower point) and 1990 (upper point) used in Fig. 4.

Fig. 4. Longitude-Height sections averaged between 55°S and 65°S for $\overline{Z'}$ for September in (a) 1989, and (b) 1990. The contour interval is 70 m. Shaded areas correspond to negative values.

Fig. 5. EOF patterns of (b) the first and (c) the second principal components obtained from the monthly mean geopotential height anomaly $\overline{Z'}$ at 10 hPa taken from September and October over the 20-year analysis interval and (a) average of the 40 months. The outer circle is the 20°S parallel. The contour interval is 0.15 and negative values are shaded for (b) and (c). For (a) the contour interval is 100 m.

Fig. 6. Scatter diagrams of the longitude and the values of the minimum $\overline{Z'}$ anomaly at 10 hPa between 50°S and 75°S. (a) Solid (open) circle indicates PC1 index positive (negative) and (b) solid (open) circle indicates PC2 index positive (negative). Histograms of the minimum value and its position are shown in the right and upper panels respectively.

tude and value of minimum $\overline{Z'}$ at 10 hPa, i.e., the location and the depth of the trough according to the sign of the PC's. The classification by PC1 sign (Fig. 6(a)) confirms the separation of the east-west drift of phase with the positive groups shifting westward, while the wave intensity is clearly separated by the sign

of PC2 (Fig. 6(b)). The amplitude for the positive PC2 category is certainly larger.

3.3 Difference of the wave-mean interaction

In order to investigate the variability of stratospheric waves shown in Section 3.2 in terms of dynamical coupling to the troposphere, we

Fig. 7. Latitude-height sections of composite monthly averaged zonal mean wind and the EP flux for the group with (a) positive (b) negative PC1 index, and (c) the difference between the two. The contour interval is 8.0 m s^{-1} for (a) and (b). The contour interval of (c) is 1 m s^{-1} . Light (dark) shaded areas are significant in 1σ (2σ). The vertical component of the EP flux is weighted by the inverse of density. The length of the unit vectors corresponds to $1.0 \times 10^6 \text{ kg s}^{-2}$ for the horizontal component and $1.0 \times 10^6 \text{ m}^3 \text{ s}^{-2}$ for the vertical component for (a) and (b). The length of the unit vector for (c) is one-tenth of that for (a) and (b). Vectors are drawn only when the significance level is above 95% (corresponding to 2σ) with bold arrows and above 68% (1σ) with thin arrows.

made a composite analysis of monthly mean zonally averaged zonal winds and the Eliassen-Palm (EP) flux vectors with both PC1 and PC2 as indices. For this purpose, the sub-dataset in which the PC score is larger (smaller) than 0.5 (-0.5) is used as the positive (negative) PC category. In practice, the numbers of samples N used in our composite analysis are $N^+ = 13$ and $N^- = 11$ for positive and negative PC1, and $N^+ = 14$ and $N^- = 12$ for positive and negative PC2, respectively.

As regards the statistical significance of this composite analysis, we make a test in a following way: From the all dataset (40 month samples), we extract 2 sets (N^+ and N^-) randomly, and make the composite difference between the two sets. By repeating this calculation for 10,000 times with randomly chosen 2 sets, we investigate the statistical distribution of the composite differences for zonal winds and EP

fluxes. As is expected, the result of this test reveals that the composite differences of 10,000 cases show the normal distribution with the median close to zero. Compared with this random test, as shown below in Figs. 7–15, the characteristic differences of our composite analysis with PC1 and PC2 indices are substantially larger than the standard deviation σ or more. Therefore we can safely say that the results of our analysis shown below are statistically significant.

Composite maps of monthly mean zonal winds and EP flux vectors for positive and negative PC1's are shown in Fig. 7(a) and (b), respectively. Vertical components of the EP flux, F_z , are weighted by the inverse of the density in order to emphasize the features in the stratosphere clearly. In both of these categories we can find common features for the late winter circulation: The polar-night jet structure is

Fig. 8. Difference between the positive and negative PC1 category in the monthly EP flux of (a) total waves, (b) zonal wavenumber 1–3, (c) zonal wavenumber 4–12 and its divergence. The contour interval for the divergence is $0.09 \text{ m s}^{-1} \text{ day}^{-1}$. Solid contours indicate the convergent region and dashed lines divergent region. Light (dark) shaded areas are significant in 1σ (2σ). The length of the unit vectors corresponds to $1.0 \times 10^6 \text{ kg s}^{-2}$ for the horizontal component and $5.0 \times 10^2 \text{ kg s}^{-2}$ for the vertical component. Vectors are drawn only when the significance level is above 95% (corresponding to 2σ) with bold arrows and above 68% (1σ) with thin arrows.

strong in the stratosphere and the subtropical jet structure is found in the troposphere. The EP vectors point upward and equatorward and the wave activity shows notable propagation into the high-latitude stratosphere where the westerly winds are strong.

The difference between the two categories is shown in Fig. 7(c). Light and dark shadows correspond to the area whose significance level is larger than 68% (i.e., the value is larger than one σ of the random distribution) and 95% (larger than 2σ), respectively. The difference of EP vectors are drawn whose magnitude ($|\mathbf{F}|$) are statistically significant in 2σ (1σ) with bold (thin) arrows. The maximum of the stratospheric westerly jet is located equatorward (poleward) for the positive (negative) PC1 category. The tropospheric westerly winds also show a different structure: The positive PC1 category corresponds to a strong lower latitude jet, described here as a “single-jet structure”, whereas the negative PC1 category corresponds to enhanced higher latitude jet, the “double-jet structure”. These features will be shown more clearly later in Fig. 11. This correspondence of the jet structure between the troposphere and the stratosphere is in agreement with the result of Aoki et al. (1996). The composite difference of the vertical component of the EP vector

in the stratosphere displays a dipole structure in harmony with the structure of the anomaly in the stratospheric zonal wind, as is expected from wave propagation theory.

The composite EP vectors in the troposphere show different features those in the stratosphere. In order to see the tropospheric processes more precisely, we show the composite difference of the tropospheric total EP flux again and its divergence between the PC1 categories in Fig. 8(a) and their breakdown into those of wave 1–3 and wave 4–12 in Fig. 8(b) and (c), respectively. Again, bold (thin) arrows indicate that difference in the magnitude of the EP vectors is statistically significant at the 95% (68%) level of significance and dark (light) shadow indicates that the difference of EP flux divergence is significant at the 95% (68%) in the area. As the components of the EP flux are not weighted by the inverse of the density in this figure, we can make a direct, visible comparison between wave activity at the different levels. Horizontal vectors in Fig. 8(a) and (c) turn weakly equatorward between 25°S and 35°S , and poleward between 45°S and 60°S . These features indicate that the positive PC1 category tends to have a small poleward momentum flux at higher latitudes and a large poleward flux at lower latitudes compared with

the negative PC1 category. Upward-pointing arrows around 40°S mean larger (smaller) poleward heat fluxes for the positive (negative) category. The contribution from wave 4–12 (Fig. 8(c)) shows similar features so that in Fig. 8(a), indicating that the difference is largely due to the synoptic-scale waves. The composite

Fig. 9. Same as Fig. 7 (c) but for the difference between the positive and negative PC2 index.

difference of the EP flux divergence contributed from the synoptic scale waves (the contour map of Fig. 8(c)), exhibits a tripole structure in the upper troposphere. The positive region around 40°S is explained by the fact that the EP flux attributed to the synoptic scale waves is weakly (strongly) convergent in this region for the positive (negative) PC1 category, because at this level the EP flux divergence ($\nabla \cdot F$) is normally negative. In the same way, the negative region around 60°S is explained by the strongly (weakly) convergence of the positive (negative) PC1 category. Compared with the difference of zonal winds (Fig. 7(c)), the difference of $\nabla \cdot F$ seems to work to maintain the zonal wind anomaly from the climatological pattern. This is consistent with the results of earlier studies (e.g., Hartmann and Lo 1998).

In Fig. 9 we next categorize the dataset into the sign of PC2 in a similar way to Fig. 7(c). The positive PC2 category (large amplitude year) tends to have a weak polar-night jet wind. It is interesting that zonal mean zonal wind in the troposphere does not show notable difference in the extratropical latitude but so does in the tropical region (around 20°S). The wave activity in the stratosphere varies so as to produce a wave-mean flow interaction. The stronger (weaker) wave activity for the positive (negative) PC2 category is consistent with the features seen in Fig. 6(b). This is also the case for the the upper tropospheric wave activity in the high latitude discussed below.

Figure 10 shows the same representation as

Fig. 10. Same as Fig. 8 but for the difference between the positive and negative PC2 index.

Fig. 11. Composite monthly mean zonal wind $\bar{u}$ at 300 hPa for the group with the (a) positive (b) negative PC1 index, and (c) the difference between the two. The contour is drawn above 20 m s^{-1} with an interval of 2.5 m s^{-1} , and values greater than 25 m s^{-1} are painted for (a) and (b). The contour interval of (c) is 1.25 m s^{-1} and light (dark) shaded areas are significant in 1σ (2σ).

Fig. 8 but for PC2 index. Here, a difference is found in the upper tropospheric vertical component between 55°S and 65°S (Fig. 10(a)) and Fig. 10(b) tells us that this is largely due to planetary-scale waves. It is also notable that the difference is prominent in the upper troposphere above 500 hPa as to planetary-scale waves. The difference in the divergence corresponds to this vertical EP flux anomaly, i.e., positive anomaly around 300 hPa between 55°S and 65°S . As the EP flux is weakly convergent or divergent in this region for both categories (not shown), the difference in the upper tropospheric divergence pattern suggests the generation of planetary waves at this level. The composite difference of EP flux divergence due to the synoptic-scale waves (Fig. 10(c)) shows the dipole pattern in the lower troposphere (cf., Fig. 8(c)).

3.4 Difference of the tropospheric circulation pattern

In this section we inspect the hemispheric pattern of the zonal wind and the eddy flux in the troposphere for the purpose of gaining a more precise understanding of the different flux patterns between the PC's categories seen in the previous section. Figure 11 shows PC1 composite maps of the monthly mean zonal wind $\bar{u}$ at 300 hPa. As was shown in Fig. 7, the positive PC1 group, which means a westward drift of stratospheric waves, corresponds to

the single-jet structure profile (Fig. 11(a)), while the negative PC1 group corresponds to a double-jet structure (Fig. 11(b)) in the upper troposphere. Between the two groups there is a large difference in the Eastern Hemisphere (Fig. 11(c)).

Taking into the consideration the notable difference in the horizontal component of the EP flux (Fig. 8(c)), the hemispheric pattern of eddy poleward momentum flux ($-\overline{u'v'}$) due to the synoptic scale waves (wavenumber 4–12) at 400 hPa is displayed in Fig. 12. The wave activity is strong between 40°S and 50°S over the Atlantic and Indian Oceans. We have also plotted the eddy momentum flux of the 10-day highpass filtered component (not shown) and found a similar pattern, with the exception of smaller fluxes over the Atlantic. These features therefore roughly represent the storm tracks in the SH (Trenberth 1991). For the PC1 negative category, there is another large poleward flux region to the south of Australia, and this difference results in the difference in the horizontal component of the EP flux (Fig. 8(c)). Compared with the zonal mean wind map (Fig. 11(b)), this region corresponds to the northward side of the exit of the polar-front jet for the PC1 negative category. The eddy momentum flux therefore acts to maintain the jet structure downstream of jet core, which is consistent with the earlier work about the jet stream and the storm tracks (e.g., Holton 1992).

Fig. 12. Same as Fig. 11 but for the poleward momentum flux $\overline{-u'v'}$ due to the zonal wavenumber 4–12 at 400 hPa. The contour interval of (a) and (b) is $6.0 \text{ m}^2 \text{ s}^{-2}$, and values greater than $24 \text{ m}^2 \text{ s}^{-2}$ are painted for (a) and (b). The positive (negative) values refer to the mean poleward (equatorward) momentum fluxes. The contour interval of (c) is $3.0 \text{ m}^2 \text{ s}^{-2}$ light (dark) shaded areas are significant in 1σ (2σ).

Fig. 13. Same as Fig. 11 but for the group with (a) positive (b) negative PC2 index and (c) the difference between the two.

The PC2 composite maps of zonal winds (Fig. 13) at 300 hPa do not show marked differences between the positive and negative category except for the subtropical jet in the Western Hemisphere. This agrees with the features seen in Fig. 9.

The poleward eddy heat flux patterns ($\overline{-v'T'}$) contributed from the planetary scale waves at 300 hPa (Fig. 14) reveal a remarkable contrast not only in strength but also in distribution. A strong poleward heat flux appears at high latitude and along the 55°S zone: There are two maxima, one between 160°E and 180° and the other around the Atlantic. In the former region,

there is a big difference in strength for the PC2 fluxes. For the latter region, the maximum of the positive PC2 fluxes lies in a broader region. The two maxima mean that the stationary wavenumber 1 component is dominant in the stratosphere.

Figure 15 shows the PC2 composite maps of eddy heat flux due to the synoptic scale waves (wavenumber 4–12) at 300 hPa. Taking a zonal mean, the difference in the magnitude is quite small as shown in Fig. 10(c). For the horizontal pattern, however, a large difference in the distribution is observed. The positive PC2 category shows a maximum in poleward eddy heat

Fig. 14. Same as Fig. 13 but for the poleward heat flux $-\overline{v'T'}$ due to the zonal wavenumber 1–3 at 300 hPa. The contour interval of (a) and (b) is 1.5 K m s^{-1} and values greater than 6.0 K m s^{-1} are painted. The positive (negative) values refer to the mean poleward (equatorward) heat fluxes. The contour interval of (c) is 1.0 K m s^{-1} . Light (dark) shaded areas are significant in 1σ (2σ).

Fig. 15. Same as Fig. 14 but for the poleward heat flux $-\overline{v'T'}$ due to the zonal wavenumber 4–12 at 300 hPa. The contour interval of (a) and (b) is 1.5 K m s^{-1} and values greater than 6.0 K m s^{-1} are painted. The contour interval of (c) is 0.75 K m s^{-1} . Light (dark) shaded areas are significant in 1σ (2σ).

flux around the Atlantic, while the negative category maximizes around the Indian Ocean.

In Fig. 10(b) it was demonstrated that for planetary-scale waves, the positive PC2 category has the larger vertical component of EP flux than the negative category, especially above 400 hPa. As the upper tropospheric zonal winds in midlatitude do not appear to have a marked difference (Fig. 9), it is interesting to consider the cause of this difference in wave activity. One possible explanation is that the nonlinear effect cannot be negligible in the tro-

posphere: The difference shown in Fig. 15 may be significant for the production of planetary scale waves through the nonlinear effect. Further examination is needed to clarify the temporal and spatial characteristics of the waves involved in the nonlinear effect (wave-wave interaction) in relation to the generation of planetary waves.

3.5 EOF of the tropospheric circulation

Finally, we make an EOF analysis on the tropospheric circulation, in order to reconfirm

Fig. 16. (a) Latitude height section (850–200 hPa) of the first EOF mode for monthly mean zonal wind averaged longitudinally for the Eastern Hemisphere. Negative values are shaded. (b) Same as Fig. 6, but solid (open) circles denote positive (negative) TUPC1.

the connection between the troposphere and stratosphere. Since a large difference in tropospheric zonal winds between the PC1 sign categories appears in the Eastern Hemisphere (Fig. 11), here we confine our attention to the monthly mean tropospheric zonal winds in eight SH tropospheric levels (850–200 hPa) averaged over 0° – 180° longitude. Before making the EOF analysis, the climatology for each calendar month (20-year average) was removed from the each dataset to extract the seasonal variation. The data are weighted by the square-root of the density at each level to give anomalies equal weights in terms of kinetic energy.

The first mode shown in Fig. 16(a) exhibits a meridional dipole pattern with a node centered around 45°S , which explains up to about 50% of the total variance. When the PC of this mode (denoted as TUPC1 hereafter) is positive (negative), the tropospheric zonal wind has a single-jet (double-jet) structure. Classification by the sign of TUPC1 reconfirms that the phase of planetary waves at 10 hPa is clearly separated as shown in Fig. 16(b), consistent with the separation based on the stratospheric PC1 sign shown in Fig. 6(a).

The difference in the tropospheric zonal wind structure between the PC1 category seems to

be similar to the mode of the low frequency variation, i.e., a zonally symmetric and equivalent barotropic dipole pattern centered at 40°S and 60°S (e.g., Yoden et al. 1987; Shiotani 1990). We made an EOF analysis of the zonally averaged monthly mean zonal wind in the troposphere and found that the first component of this PC is certainly well correlated with TUPC1 (correlation coefficient is 0.78). However, we could not determine whether it categorizes the phase of the stratospheric wave. Therefore, these results suggest that the difference seen in the zonal wind structure between the PC1 category (Fig. 11) is not the same as the tropospheric zonally symmetric variation presented by earlier authors.

4. Concluding remarks

We have made an EOF-based statistical study of the interannual variations of planetary waves in the wintertime SH stratosphere using NCEP/NCAR reanalysis dataset covering a 20 year period. The results were used to interpret some aspects of upper tropospheric circulation such as the mean zonal flow and various scale waves. Resulting from this work, the characteristic features of the interannual variations of the stratospheric planetary waves and their

connection with the tropospheric dynamics can be summarized as follows:

- The year-to-year variation of monthly mean planetary waves at 10 hPa in late winter can be characterized by an east-west drift of phase, and by a well defined wave amplitude variation.
- The variation of the phase is related to the structure in the tropospheric jet. The westward (eastward) drift of the stratospheric wave corresponds to a single-jet (double-jet) structure in the upper troposphere over the Eastern Hemisphere. The activities of synoptic-scale waves show a variety of different features which to maintain the jet structure.
- The vertical component of the EP flux in the troposphere is larger when stratospheric planetary waves have the large amplitudes and this inclination is most pronounced around the 300 hPa level. In relation to the hemispheric pattern of monthly mean activities of the planetary scale waves in the troposphere, we found a difference in the maximum location and the intensity of the heat flux according to the difference in the amplitude variability. The hemispheric pattern of the eddy heat flux due to the synoptic waves also shows a different distribution.

Building on the basis of the present statistical study, future work will aim to clarify the dynamical process to explain the connection between the troposphere and the stratosphere. Scinocca and Haynes (1998) introduced the Transformed Fourier Decomposition (TFD) equations to define the nonlinear forcing. The TFD equations express the forcing of a specific wavenumber by wave-wave interactions and employ a relation between the Fourier decomposed quasi-geostrophic potential vorticity equation and the total eddy-vorticity forcing. In this sense, the TFD equations would be a powerful tool to investigate the effect of synoptic waves on the generation of planetary scale waves propagating into the stratosphere.

Since the distributions of the storm tracks and the zonal wind structure are tightly correlated in the SH (e.g., Karoly 1990), we also need further studies to define the relationship between the zonal wind and horizontal eddy fluxes due to the synoptic waves, in terms of the interannual variability.

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