Interannual Variations of the Seasonal March in the Southern Hemisphere Stratosphere for 1979–2002 and Characterization of the Unprecedented Year 2002

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ABSTRACT

Dynamical features of the interannual variations of the seasonal march in the Southern Hemisphere stratosphere are investigated with the NCEP–NCAR reanalysis dataset from 1979 to 2002, and the unprecedented year 2002, in which a major stratosphere sudden warming occurred, is characterized by comparing it with the other 23 yr.

A multiple empirical orthogonal function analysis of the stratospheric mean zonal wind and a composite analysis based on the principal component of the leading mode show that the interannual variations are characterized by early or late deceleration of the polar-night jet, which is well correlated with the variation of a time-averaged upward Eliassen–Palm (EP) flux in the lower stratosphere. The stronger wave activity in the lower stratosphere is associated with the earlier "shift down" of the jet. The composite difference of the stratospheric mean zonal wind can be traced down to the lower troposphere in September and October. These features are consistent with the variations of the Southern Hemisphere annular mode, although the main disturbance to maintain the variations is different between the stratosphere and troposphere.

Some scatter diagrams show the extreme situation of the year 2002. It is far from the cluster of the other 23 yr, but the large deviation in 2002 is consistent with the tendency of the fluctuations in the other years except for its extreme nature.

1. Introduction

In September 2002, a major stratospheric sudden warming event occurred in the Southern Hemisphere (SH), and the polar vortex split into two for the first time since satellite observations of the stratosphere began. Midwinter warming events had been observed in 1988 (e.g., Kanzawa and Kawaguchi 1990; Hirota et al. 1990), but they were minor ones. Planetary waves in the winter stratosphere were much more vigorous in 2002 compared with other years (Fig. 4 in Baldwin et al. 2003). Rather periodic strengthening of the wave activity in August and early September corresponds to the sequential occurrence of three minor warming events observed prior to the major warming event in late September. Enhancement of the wave activity can be traced back to the middle troposphere (Fig. 6 in Baldwin et al. 2003). In this paper, we consider these warming events as episodes in the seasonal march of the stratospheric circulation, and investigate interannual variations of the seasonal march to characterize the dynamical conditions in the unprecedented year 2002.

Shiotani et al. (1993) studied interannual variations of the seasonal march in the SH by characterizing 10 winters during the 1980s according to the location of the core of the polar-night jet in midwinter. They defined low-latitude jet (LLJ) years and high-latitude jet (HLJ) years. They showed that the timing of the poleward and downward shift ("shift-down") of the polar-night jet in late winter is earlier in the HLJ years than in the LLJ years, and the planetary wave activity from autumn to late winter is also large in the HLJ years. Aoki et al. (1996) investigated interannual variations of the tropospheric circulation and its relation to the stratosphere by using a dataset for 15 yr from 1979 to 1993. The data showed that a zonal wind index, defined as the difference of the 300-hPa zonal-mean zonal wind speed between 60° and 40°S, is a sensitive indicator of two typical latitudinal profiles of the zonal-mean zonal wind in the SH troposphere noted as the single-jet and double-jet profiles, and found that the seasonal evolution of the zonal wind index from April to October is separated into two groups, double-single (DS) years and single-double (SD) years, where DS (SD) years are characterized by the double jet (single jet) in early winter and the single jet (double jet) in late winter. There is a correspondence between interannual variations in the seasonal march of the stratospheric polar-night jet and those of the tropospheric jet structure as listed in Table 1; DS years appear in LLJ years and HLJ years appear in SD years.

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TABLE 1. Classification of 24 yr (1979–2002) based on the PC1 of the seasonal march variation (Fig. 3) in this study and some other classifications of the interannual variations in the SH in the previous studies. For PC1, E is ED year, L is LD year, e is positive PC1 smaller than the threshold value, and l is negative PC1 larger than the threshold value. For Shiotani et al. (1993; SSH), L is LLJ year and H is HLJ year. For Aoki et al. (1996; ASH), D is DS year, and S is SD year. Dashed marks indicate that years cannot be classified into the categories and blanks indicate that classification of the corresponding years are not investigated in their papers. For Naito (2002) plus the updated period (1997–2000; Naito$^+$), W (E) is the westerly (easterly) phase of the QBO.

Year	79	80	81	82	83	84	85	86	87	88	89	90	91	92	93	94	95	96	97	98	99	00	01	02
PC1	E	l	l	e	l	l	L	E	L	E	l	L	E	E	l	e	L	l	e	L	L	E	L	E
SSH		L	L	—	—	H	H	H	L	—	L													
ASH	—	D	D	S	—	S	S	S	—	S	D	S	D	S	D									
Naito$^+$	E	W	E	E	E	E	W	E	W	E	E	W	E	W	E	E	W	E	W	E	W	E	E	W

A multiple empirical orthogonal function (EOF) analysis of the tropospheric and stratospheric circulation was performed by Kuroda and Kodera (1998) with a 16-yr dataset to extract the interannual variations of the seasonal march from autumn to spring. The leading EOF is characterized by the poleward and downward shift of anomalous zonal-mean zonal winds in the SH stratosphere, associated with anomalous propagation of planetary waves from the troposphere. The mode of the first EOF has a long time scale of nearly half of a year through atmospheric wave–mean flow interactions.

Wave driving from the troposphere is also related to the mean meridional circulation. Fusco and Salby (1999) proposed that the planetary wave activity entering the lower stratosphere in the extratropics is a good dynamical proxy for the diabatic meridional circulation and showed that the interannual variations of the upward Eliassen–Palm (EP) flux in high latitudes at 100 hPa are coherent with those of the tendency of the total ozone in the Northern Hemisphere (NH). Randel et al. (2002) investigated the space–time patterns of this coherency in both hemispheres and showed that the high correlation moves from the midlatitudes during winter to polar latitudes in spring. As to the polar temperature, Newman et al. (2001) demonstrated that in the NH, the upward EP flux at 100 hPa in mid- to late winter (January–February) is highly correlated with the mean polar stratospheric temperature in March.

Following these studies, we investigate the interannual variations of the seasonal march in the stratosphere and their relation to the troposphere. We use the daily data of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis for the 24-yr period from 1979 to 2002 at 17 pressure levels (1000–10 hPa) as in Hio and Hirota (2002) in order to characterize the unprecedented year 2002 in the whole analyzed years.

The present paper is organized as follows: In section 2, year-to-year variations in the seasonal march of the stratospheric mean zonal winds and relation to the tropospheric circulation are described, and similarity or uniqueness of the circulation in 2002 is represented. Interannual variations related to the upward EP flux in the lower stratosphere are investigated in section 3. A discussion is presented in section 4, and conclusions are presented in section 5.

2. Interannual variations of the seasonal march

a. *Zonal-mean zonal wind and planetary waves in the stratosphere*

Figure 1 shows the seasonal evolution of the climatological zonal-mean zonal wind at 60°S, 20 hPa (thick solid line), along with the standard deviation of the interannual variability (shaded region). The mean zonal wind reaches its maximum in late August, and interannual variability is large from August to November. In 2002 (thin solid line), the polar-night jet began to decelerate in early August, and then the largest reduction occurred in late September, corresponding to the polar vortex split. Comparing with the seasonal march of 1988 (dotted line), the year in which the midwinter warming events occurred, we can recognize how the deceleration in 2002 was an extreme event.

We characterize interannual variations of the seasonal evolution by making a multiple EOF analysis (e.g., Kuroda and Kodera 1998) for the 24-yr dataset, which is performed by combining 12 months of data in a vector $\mathbf{x}^i$ as

$$\mathbf{x}^i = [u^i(1), u^i(2), \ldots, u^i(12)]^{\mathrm{T}}, \tag{1}$$

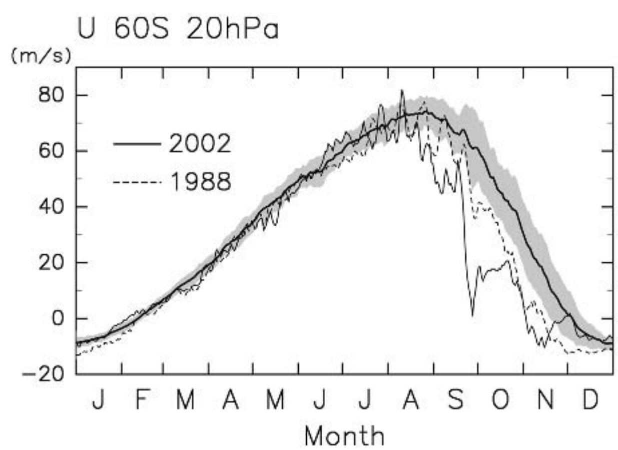

FIG. 1. Climatological annual cycle of 20-hPa zonal-mean zonal wind at 60°S, for 24 yr from 1979 to 2002 (thick solid line), along with the standard deviation of its interannual variability (shaded). Thin solid (dotted) line denotes day-to-day variations in 2002 (1988).

where $u^i(m)$ is the anomalous monthly averaged zonal-mean zonal wind at 60°S, 20 hPa for the mth month of the ith year. The modes of the EOF are defined as the eigenvectors of the covariance matrix calculated from $\mathbf{x}^i$. The first mode (EOF1) shown in Fig. 2 represents interannual variation of the deceleration of the polar-night jet from September to December. Principal component 1 (PC1) explains as much as 66% of the total variance, so that it is safe to give attention to EOF1 only.

Both EOF1 and EOF2 are similar to those of the model experiment by Taguchi and Yoden (2002, their Fig. 7), though they used the stratospheric polar temperature as a measure of the polar vortex intensity instead of the mean zonal wind.

Figure 3 shows year-to-year variation of PC1 for the 24 yr. In the year whose principal component of EOF1 (PC1) is positive (negative), the polar-night jet decelerates earlier (later) than the climatological seasonal march. The two largest PC1s for 2002 and 1988 reflect the large deceleration of the polar-night jet in August and September corresponding to the warming events. The absolute value of the most negative PC1 is not so large as the most positive one (only −1.5 for 1987), and the frequency distribution of PC1 has some skewness as shown in Table 2. To capture the associated interannual variations in some dynamical fields in the stratosphere and troposphere, we make a composite analysis for two groups defined with the value of PC1: early-deceleration (ED) years and late-deceleration (LD) years. Here, ED (LD) consists of the years of the seven largest (smallest) PCs, indicated by closed (open) circles in Fig. 3. The horizontal dashed lines are the threshold values for the definition, PC1 = 0.53 and −0.66.

A composite analysis for ED and LD years is done after applying a 31-day running mean for smoothing.

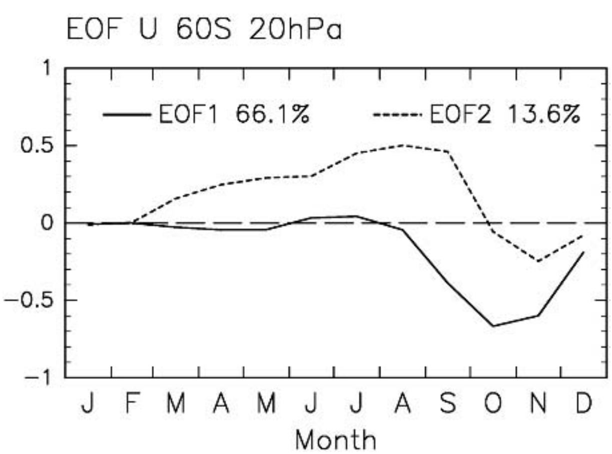

FIG. 2. EOFs of the 12 months variation in zonal-mean zonal wind at 60°S, 20 hPa from the climatological seasonal march on the basis of the covariance matrix of monthly averaged values for the 24-yr dataset from 1979 to 2002.

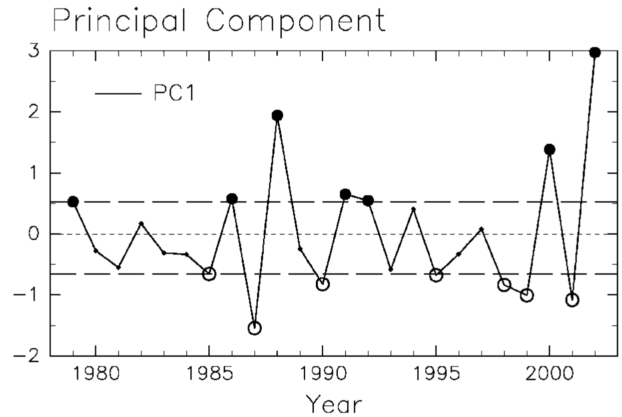

FIG. 3. Year-to-year variations of the principal component of EOF1. Closed and open circles denote ED and LD years, respectively. Dashed lines indicate the thresholds to determine the ED and LD years.

Figure 4 (top) shows the composite seasonal evolutions of the zonal-mean zonal wind at 60°S, 20 hPa for ED (thick solid line) and LD (thick dashed line) years. Shaded regions show the ranges of ±0.5 times standard deviation for each group. When these regions do not overlap, the difference between the two groups is significant at more than 90% based on the t test. The difference in deceleration of the polar-night jet is highly significant from September to December. The zonal-mean zonal wind for 2002 (thin solid line) is much weaker already in August, even if it is compared with the composite for ED years.

Figure 5 (left) displays latitude–time sections of the composite mean zonal wind at 20 hPa for ED years (top), LD years (middle), and the composite difference (bottom). The polar-night jet in ED years is weaker on the equatorward flank of the jet in August and shifts poleward earlier than that in LD years. As a result, the negative composite difference in the midlatitudes moves poleward together with the polar-side positive one from August to October. The large negative difference in November and December reflects the difference in the timing of the shift to summer circulation. There is a significant difference in low latitudes from July to December. The composite mean zonal wind is easterly in ED years and westerly in LD years. Its association with the quasi-biennial oscillation (QBO) of the equatorial mean zonal wind will be discussed later. Note also the significant difference in March and April in high latitudes.

Composites of time evolution of the geopotential amplitude of zonal wavenumber 1 (wave 1) at 60°S, 20 hPa for ED and LD years are shown in Fig. 4 (middle). Latitude–time sections of the wave-1 amplitude at 20 hPa for each group and their difference are also shown in Fig. 5 (center). The amplitude of the wave 1 for LD years has two maxima in early winter and in spring, which are a well known feature in the SH. On the other

TABLE 2. Normalized frequency distributions (in %) with each standard deviation σ of PC1, the zonal-mean zonal wind at 45°S (U), 20 hPa averaged for 1–15 Oct for the 24 yr (1979–2002) in this study, and the 1000-yr middle stratospheric polar temperature in Aug obtained by Taguchi and Yoden (2002; TY) with a troposphere–stratosphere coupled model corresponding to the SH. The distribution of the zonal-mean zonal wind is multiplied by –1 for dynamic consistency with the other distributions. Gaussian frequency distribution is also shown for comparison.

	-3σ	-2σ	$-\sigma$	0	1σ	2σ	3σ	4σ	5σ	6σ
PC1	0.0	12.5	45.8	29.2	8.3	4.2	0.0	0.0	0.0	
U	4.2	4.2	58.3	20.8	8.3	0.0	4.2	0.0	0.0	
TY	0.3	8.7	47.7	32.8	7.0	1.8	1.1	0.2	0.2	
Gaussian	2.1	13.6	34.1	34.1	13.6	2.1	0.1	3.0×10^{-3}	—	

hand, the amplitude for ED years has one maximum in spring; the maximum amplitude is larger and appears earlier. In both groups, the location of the amplitude peak moves poleward from September to November along with the shift of the polar-night jet. The composite difference also shows poleward movement of positive values from late July to early November. A statistically significant difference is also seen in June in high latitudes.

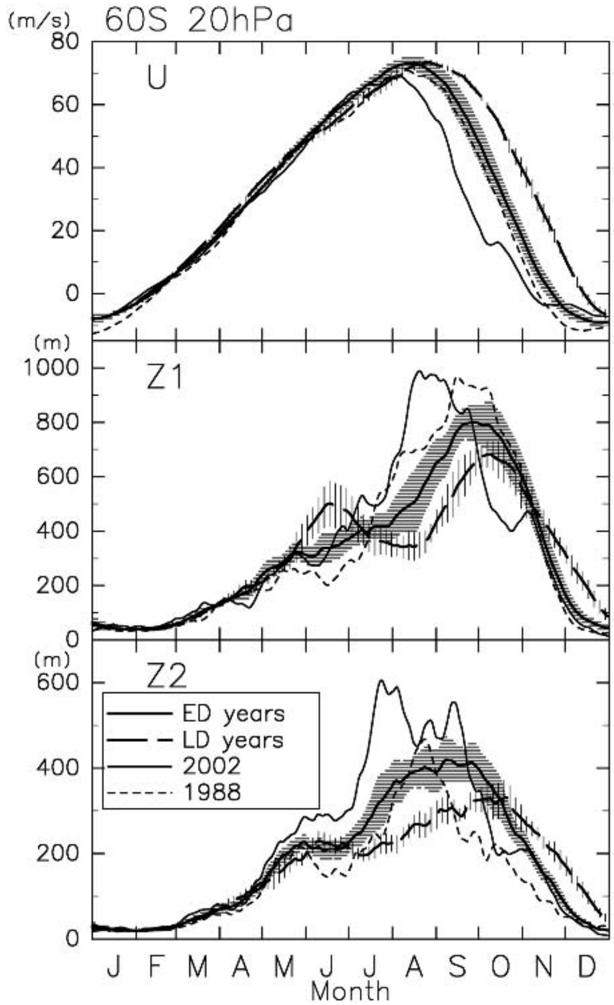

FIG. 4. Time series of the composites for ED years (thick solid lines) and LD years (thick dashed lines) of the zonal-mean zonal wind at (top) 20 hPa, 60°S and 20-hPa geopotential amplitude of (middle) wave 1 and (bottom) wave 2 at 60°S, along with the ±0.5 times std dev (shaded region for ED years and vertical bars for LD years). Thin solid line and dotted line denote the time series for 2002 and 1988, respectively. A 31-day running mean is applied to each quantity beforehand.

Similar features are seen in the amplitude of wave 2 as shown in Fig. 4 (bottom) and Fig. 5 (right). The ED years have a larger amplitude in August and September, while they have a smaller one in November and December.

As for the year 2002 shown in Fig. 4 with the thin solid line, wave 2 amplified sharply in July, and then wave 1 amplified in August. Both amplitudes are much larger even if compared with the composite for ED years. The amplitude of wave 1 in 1988 (dotted line) is very large in September, although that of wave 2 is not so large except for August.

b. *Composite differences of the zonal-mean zonal wind and wave activity in the stratosphere and troposphere*

Contours in Fig. 6 denote latitude–height sections of the composite difference in monthly averaged zonal-mean zonal wind between ED and LD years for 9 months from April to December. Significant differences are found as early as May around 30°S above 30 hPa. The dipole pattern moves poleward and expands downward in the stratosphere from June to October. This month-to-month variability corresponds to the earlier shift-down of the polar-night jet in ED years. The difference in the zonal-mean zonal wind is largest in November over 24 m s^{-1} at 65°S, 20 hPa. Significant differences in the troposphere are also seen from August to December. The mean zonal wind around 60°S is weaker over 3 m s^{-1} in ED years in the upper troposphere in September and October and in the whole troposphere in November and December. On the other hand, that around 35°S is significantly stronger in ED years from September to November. In low latitudes, the difference of the mean zonal wind between ED and LD years is negative in the middle stratosphere, suggesting that the QBO phase at these levels tends to be easterly (westerly) for the ED (LD) years.

The composite difference of the monthly averaged EP flux is also shown in Fig. 6 with arrows, which are drawn only at the positions where the difference in ei-

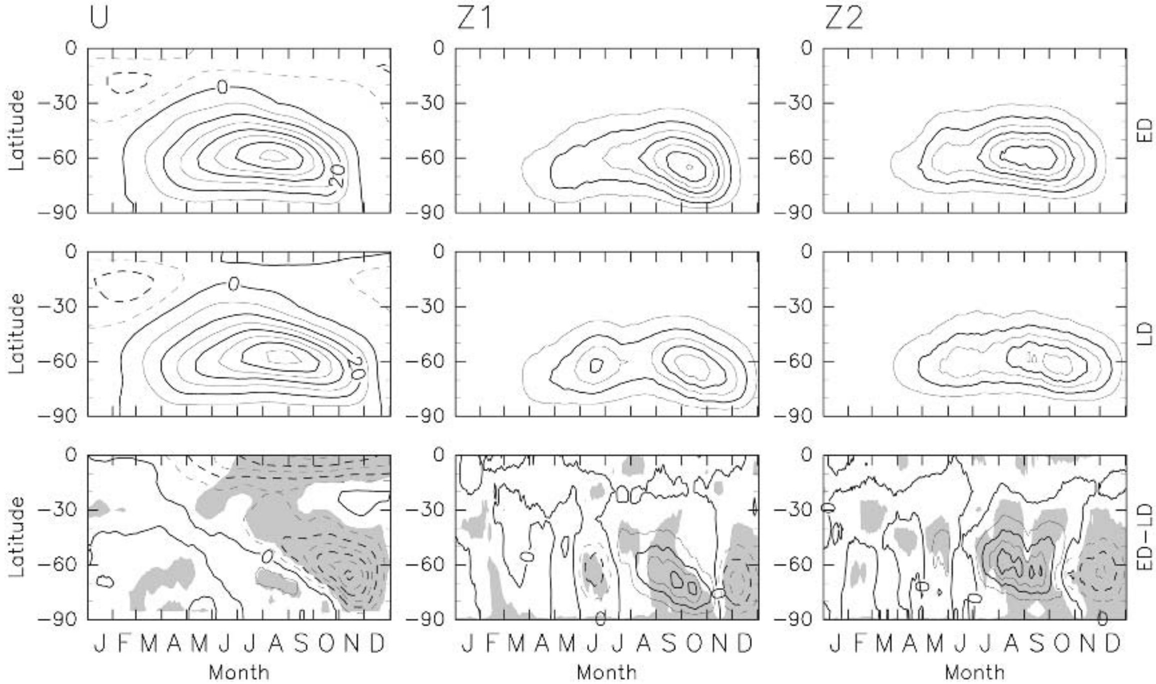

FIG. 5. Latitude–time sections of the (left) zonal-mean zonal wind and geopotential height amplitude of (center) wave 1 and (right) wave 2 for the composites of (top) ED years and (middle) LD years, and (bottom) the differences between the two groups. Dotted contours denote negative values. Areas where the statistical significance of the difference is larger than 90% based on the *t* test are shaded in the bottom panels. Contour intervals for the mean zonal wind are (top and middle) 10 and (bottom) 4 m s^{-1}. Contour intervals of the amplitude of wave 1 are (top and middle) 120 and (bottom) 60 m, while those of the amplitude of wave 2 are (top and middle) 60 and (bottom) 30 m. A 31-day running mean is applied to each quantity beforehand.

ther component is statistically significant at the 90% level. From July to October, the upward EP flux for ED years is larger in the high-latitude stratosphere. This is consistent with the difference in the geopotential amplitudes of wave 1 and wave 2 shown in Figs. 4 and 5. Along with the earlier evolution of the polar-night jet, the difference of the EP flux also moves poleward and downward. The equatorward EP flux in the midlatitude upper troposphere is weaker in ED years in August and September, and synoptic-scale waves are mostly responsible for this difference (not shown). A significant difference in the EP flux is also seen in the midlatitude lower troposphere in September and October. The composite difference in the mean zonal wind in the troposphere is mainly associated with the difference in the EP flux by synoptic-scale waves.

c. *Characterization of the year 2002*

Differences of the monthly averaged zonal-mean zonal wind and the monthly averaged EP flux in 2002 from the composite for LD years are shown in Fig. 7 for 9 months from April to December. Many of the patterns of the differences resemble those between ED and LD years (Fig. 6). A poleward movement of the dipole pattern in the stratospheric mean zonal wind is also seen from June to September with much larger values and is accompanied with a much larger upward EP flux in the stratosphere. Particularly, the dipole pattern in August expands down into the high-latitude troposphere, and the upward EP flux is also larger in the polar region of the whole troposphere where the mean zonal wind is stronger. In October and November, the difference of the mean zonal wind is much larger compared with those in Fig. 6, indicating a much earlier seasonal march in 2002. Note that in October, the difference of the EP flux is downward in the stratosphere showing the weaker upward flux in 2002, which is consistent with the smaller amplitudes of wave 1 and wave 2 in 2002 as shown in Fig. 4.

The differences of the tropospheric mean zonal wind from August to November are also larger than those in Fig. 6. In agreement with these differences, the difference of the vertical component of the EP flux in the lower troposphere is upward around 50°S and downward around 70°S in September, October, and November. In low latitudes, the differences of the stratospheric mean zonal wind have the opposite sign from those in Fig. 6 and shift downward from April to December. This reflects the fact that the QBO phase of the equatorial mean zonal wind is westerly in 2002 while it tends to be easterly in ED years. We shall come back to these in more detail in the discussion section.

It is also notable that the mean zonal wind in high

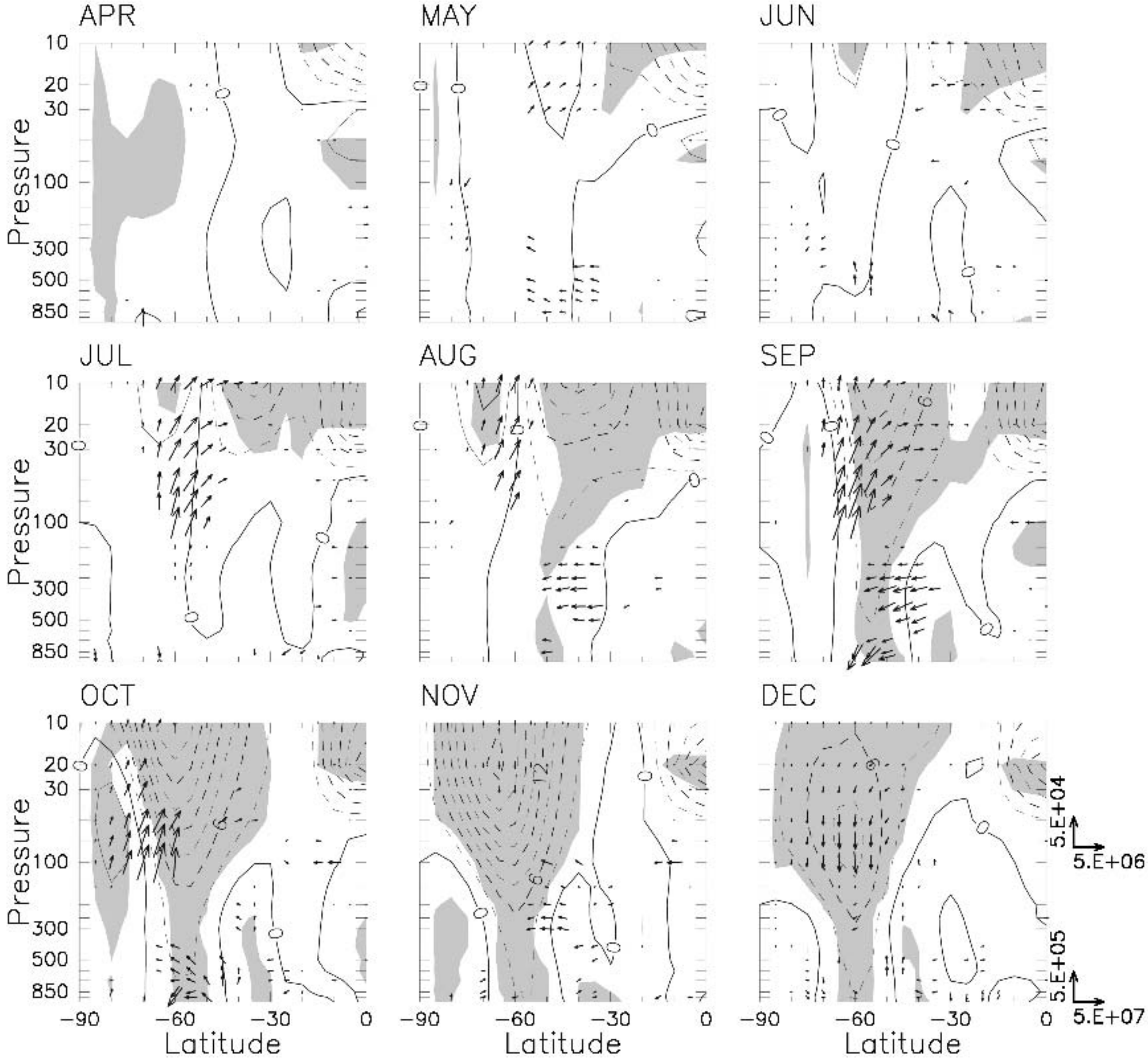

FIG. 6. Latitude–height sections of composite difference of the monthly averaged zonal-mean zonal wind between ED and LD years for 9 months from Apr to Dec. The contour interval is 3 m s^{-1}, and dotted contours denote negative values. Areas where the statistical significance of the difference is larger than 90% based on the *t* test are shaded. Composite difference of the EP vectors is also drawn by an arrow at the location where the significance level of the difference in either component is above 90%. The length of the unit vector is displayed on the right-hand side of the graph for Dec. The scaling of the unit vector below 100 hPa is 10 times as large as that above 100 hPa.

latitudes in May is weaker in 2002 both in the troposphere and the lower stratosphere, with the largest difference in the upper troposphere associated with the poleward and upward EP flux anomaly in the same region.

3. Relation to the upward EP flux in the lower stratosphere

a. Correlation of the upward EP flux to the zonal-mean zonal wind

Figure 8 shows the seasonal evolution of the climatological upward EP flux at 100 hPa averaged over 45°–75°S (thick solid line) along with the standard deviation of the interannual variability (shaded region). The climatological EP flux has a maximum in late September, and interannual variations are large through winter and spring. In 2002 (thin solid line), the upward EP flux is very large in August and September, while it is small in October. These features were already pointed out with the differences of the EP vectors shown in Fig. 7. In 1988 (dotted line), the larger upward EP flux compared with the climatology is seen in July and August, though it is not so large as that in 2002.

We use the upward EP flux at 100 hPa between 45° and 75°S as a proxy of the wave activity in the lower stratosphere to investigate the relationship to some dy-

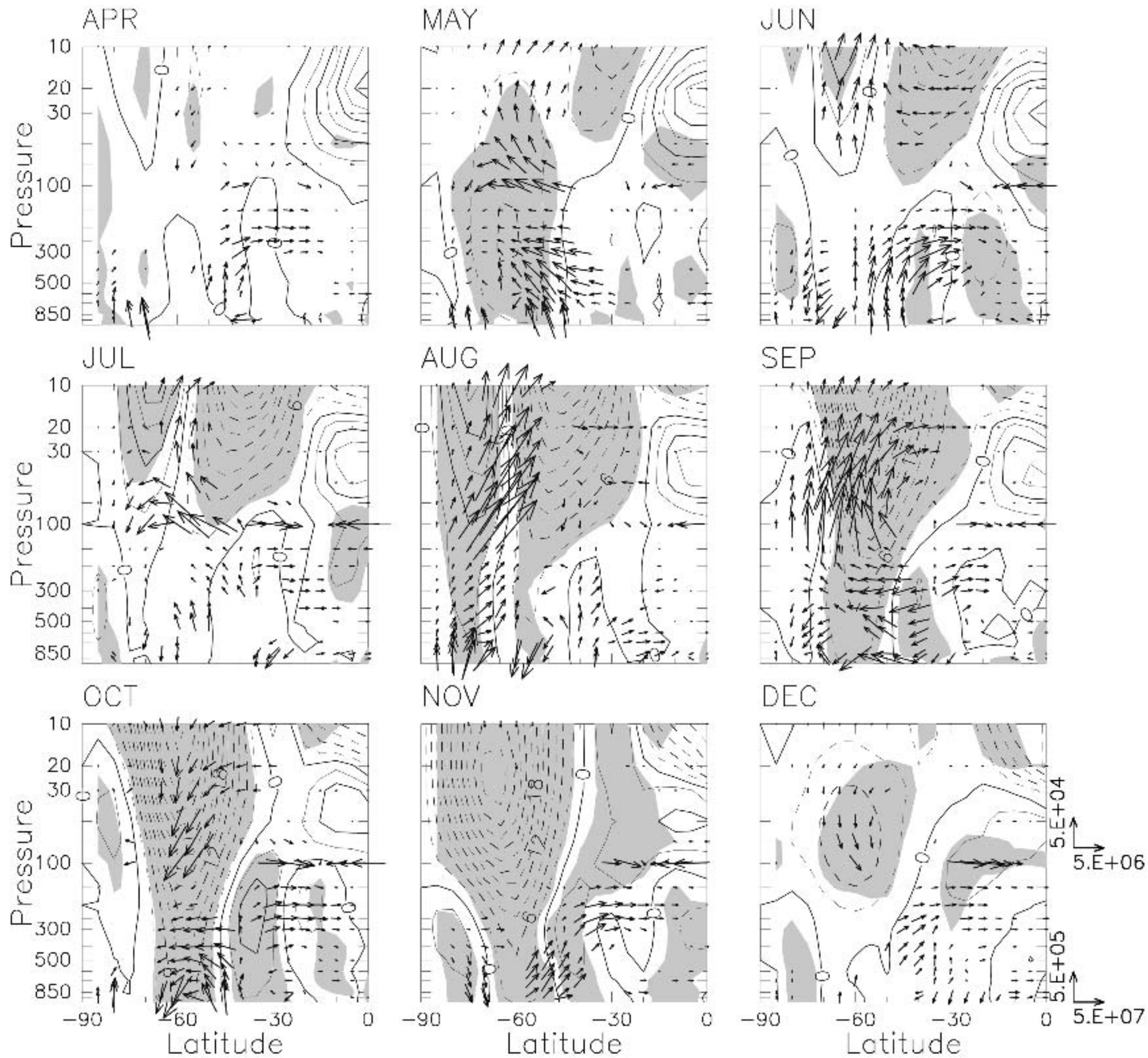

FIG. 7. Latitude–height sections of the difference of the monthly averaged zonal-mean zonal wind between 2002 and LD years (2002 minus LD) for 9 months from Apr to Dec. The contour interval is 3 m s^{-1}, and dotted contours denote negative values. Shaded regions correspond to the areas where the anomaly is greater than the 1.5 times the std dev of LD years. Composite difference of the EP vectors is also drawn by an arrow. The length of the unit vector is displayed on the right-hand side of the graph for Dec. The scaling of the unit vector below 100 hPa is 10 times as large as that above 100 hPa.

namical fields in the stratosphere and troposphere. Figure 9 shows correlations of the 24-yr time series of the zonal-mean zonal wind averaged for the first 15 days of each month with the upward EP flux in the lower stratosphere during the preceding 45 days for 9 months from April to December. This average period for the EP flux is the same as that used in Newman et al. (2001), which was based on the thermal damping rate in the middle stratosphere. The correlations that exceed ±0.4 (0.65) are significant at the 95% (99.9%) level for 24 data values and are lightly (heavily) shaded in the figure. From July to September, strong negative correlations are found in the equatorward flank of the climatological core of the polar-night jet, and positive ones are found in the poleward flank. This dipole pattern moves poleward and downward with the seasonal shift of the climatological polar-night jet, indicating that the stronger wave activity in the lower stratosphere is associated with the earlier shift-down of the polar-night jet. The significant positive correlations in the polar stratosphere disappear in October, while the negative correlations spread in mid- and high latitudes in November and December. In the troposphere, a dipole pattern with positive correlations in the poleward flank of the subtropical jet and negative ones in the equatorward flank is found in October; this relation suggests that the larger (smaller) upward EP flux in the lower stratosphere in late August and September is related to the poleward (equatorward) shift of the subtropical jet in October.

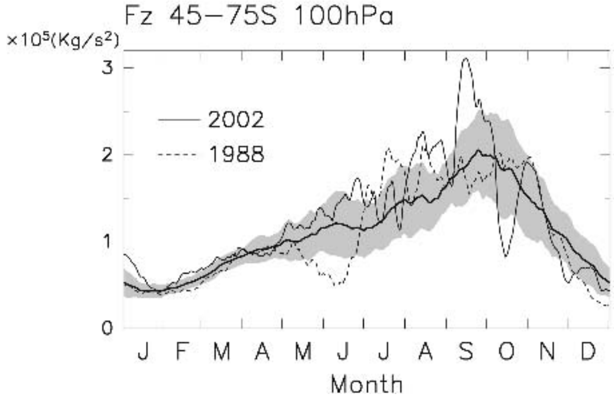

FIG. 8. Same as in Fig. 1, but for the upward EP flux at 100 hPa averaged for 45°–75°S. A 31-day running mean is applied to each quantity beforehand.

The patterns of the correlation in the stratosphere from July through November are similar to those of the composite difference of the zonal-mean zonal winds between ED and LD years shown in Fig. 6. This means that the EP flux in the lower stratosphere is a good indicator of the interannual variations in the seasonal march of the stratospheric circulation in the SH through winter and spring. Large negative correlations are found on the equator-side of the core of the polar-night jet, even in the formation stage of the jet from May to June.

b. Scatter diagrams to characterize the year 2002

Figure 10 (top left) shows a scatter diagram for the pair that gives the maximum correlation in the stratosphere during the nine months shown in Fig. 9 (at the point marked with a cross). The upward EP flux averaged over the period from 16 August to 30 September is plotted against the zonal-mean zonal wind at 45°S, 20 hPa averaged for 1–15 October for the 24-yr data. The correlation with the stratospheric mean zonal wind is very high whether the year 2002 is included in the calculation ($r = -0.86$) or not (-0.73). For these periods in 2002, the mean zonal wind is easterly, and the upward EP flux in the lower stratosphere is extremely large compared with the other 23 yr. These features are also found in the relationship between the late-September polar temperature at 50 hPa and the preceding upward EP flux in the lower stratosphere (Newman and Nash 2005).

Now we introduce a confidence ellipse in the assumption of the bivariate normal distribution (see Wilks 1995) to characterize the year 2002. The solid-line ellipse denotes a 90% confidence limit for the full dataset with the year 2002, while the dotted line represents a 90% confidence ellipse for the other 23 yr without 2002. The major and minor axes of the confidence ellipses are oriented in the direction of the first and second eigenvectors of the variance–covariance matrix, respectively, and are stretched in these directions by an amount proportional to the square roots of the respective eigen values. That is, when the correlation between the two quantities is so high that the first eigen value is much larger than the second one, the ellipse is stretched much more along the major axis. The stretched ellipses in Fig. 10 (top) are another description of the high correlation between the upward EP flux in the lower stratosphere and the stratospheric mean zonal wind. The confidence ellipse is stretched more in almost the same direction when the year 2002 is included in the calculation, even though the year 2002 is outside of both confidence ellipses. These features indicate that the point for 2002 is far from the clusters of the other 23 yr satisfying the fundamental relationship obtained from the other years.

In Fig. 9, another peak of the correlation of the 45-day averaged upward EP flux with the mean zonal wind averaged for the following 15 days is found in the troposphere at the point marked with a plus sign. Figure 10 (bottom left) is the scatter diagram for the EP flux in the lower stratosphere and the mean zonal wind at 35°S, 200 hPa. The correlation is high ($r = 0.73$) for the full dataset. If the year 2002 is excluded from the calculation, the correlation is reduced to 0.47, but the confidence ellipse does not change the angle of the major axis. Both the EP flux and the mean zonal wind are largest in 2002 in the 24 yr.

Figure 11 shows the correlation map of the vertical component (left) and the horizontal component (right) of the EP flux with the upward EP flux at 100 hPa between 45° and 75°S for the same period from 16 August to 30 September. Both components highly correlate over a wide area in the stratosphere with the largest value along 60°S for the vertical component and in lower latitudes at the 20-hPa level for the horizontal component. This means that the propagated EP flux converges in the middle stratosphere at the lower latitudes, which is consistent with the result shown in Fig. 9 that the maximum correlation of the mean zonal wind to the upward EP flux at the 100 hPa between 45° and 75°S is located on the equator side of the reference latitudes of the upward EP flux.

A scatter diagram of the horizontal EP flux at 50°S in the upper troposphere averaged between 300 and 150 hPa, and the tropospheric mean zonal wind at the same position as used in Fig. 10 (left) is shown in Fig. 10 (right) to see the relationship more directly in the upper troposphere. The height range for the averaged horizontal EP flux is shown in Fig. 11, with the dashed line showingwhere the equatorward EP flux is weakly correlated with the upward EP flux at 100 hPa. The tropospheric equatorward EP flux at the poleward side of the subtropical jet has a negative correlation with the tropospheric mean zonal wind, though it is not highly significant. The year 2002 is outside of both confidence ellipses, representing the uniqueness of this year. The equatorward EP flux is smallest, while the mean zonal wind is strongest.

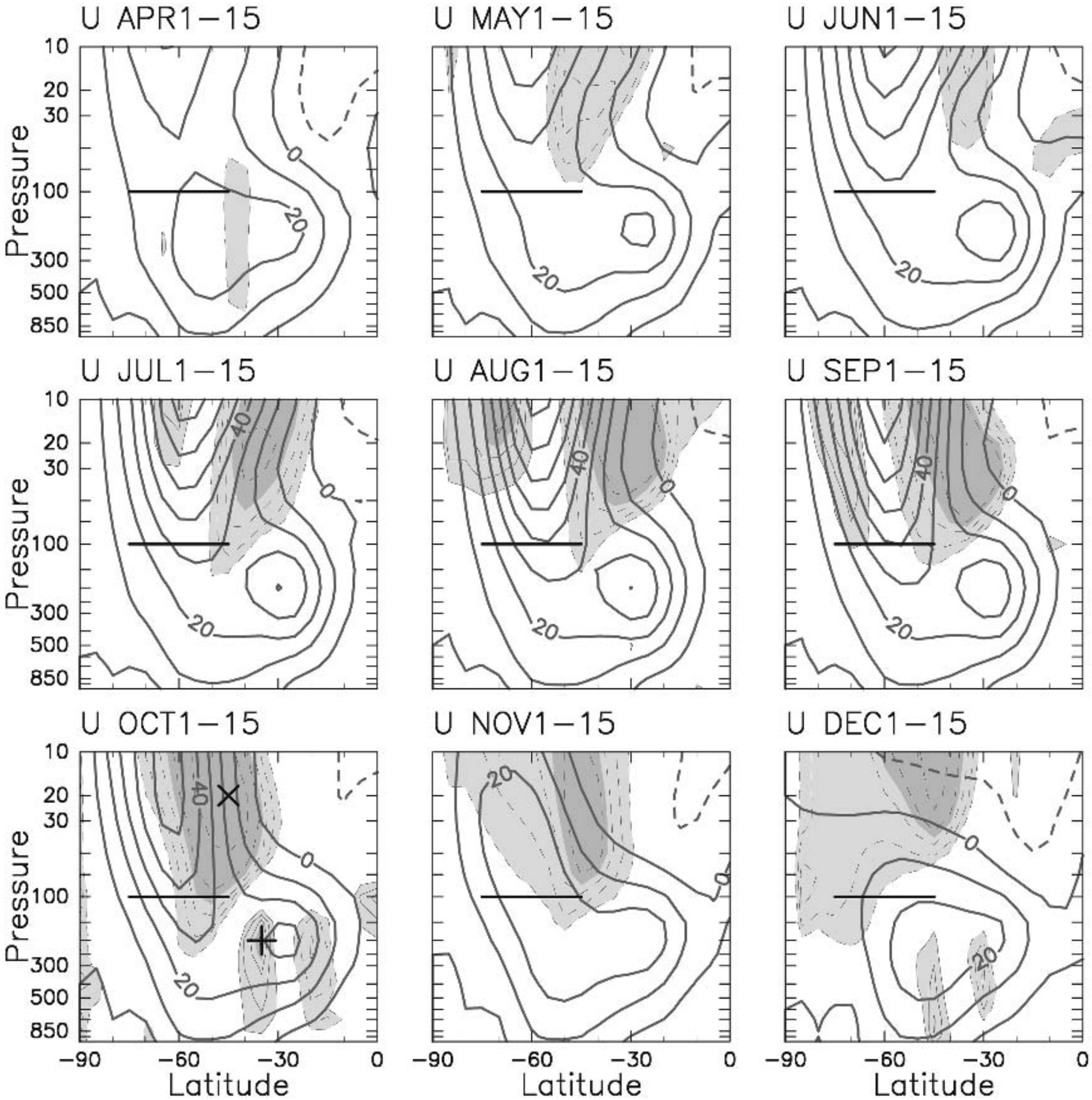

FIG. 9. Correlation of the 24-yr time series of the zonal-mean zonal wind averaged for the first 15 days of each month with the upward EP flux at 100 hPa for 45°–75°S during the period preceding 45 days for 9 months from Apr to Dec. Thin line contours are drawn above 0.4 or below −0.4 with an interval of 0.1, and values over ±0.4 (0.65) are lightly (heavily) shaded. Dotted contours denote negative values. Climatological zonal-mean zonal wind averaged for the same first 15 days is superimposed with thick gray lines. The contour interval for the mean zonal wind is 10 m s^{-1}.

In these scatter diagrams, we can see the extreme situation of the year 2002. The points for 2002 are far from the clusters of the other 23 yr and are outside of the 90% confidence ellipses. As for the relationship between the upward EP flux and the mean zonal wind, angles of the major axis of the confidence ellipses for the full 24 yr are almost the same as those for the 23 yr without the year 2002, indicating that the large deviations in 2002 are consistent with the tendency of fluctuations in the other years.

Note that the other two years with very large PC1 in Fig. 3 (i.e., 1988 and 2000) are also found on the same side as the year 2002: large upward EP flux in the lower stratosphere, weak mean zonal wind in the stratosphere, strong mean zonal wind in the upper troposphere, and small equatorward EP flux in the upper troposphere.

4. Discussion

Shiotani et al. (1993) introduced HLJ and LLJ years based on the location of the polar-night jet in midwinter and pointed out some relevance to the timing of the deceleration of the polar-night jet in late winter, such

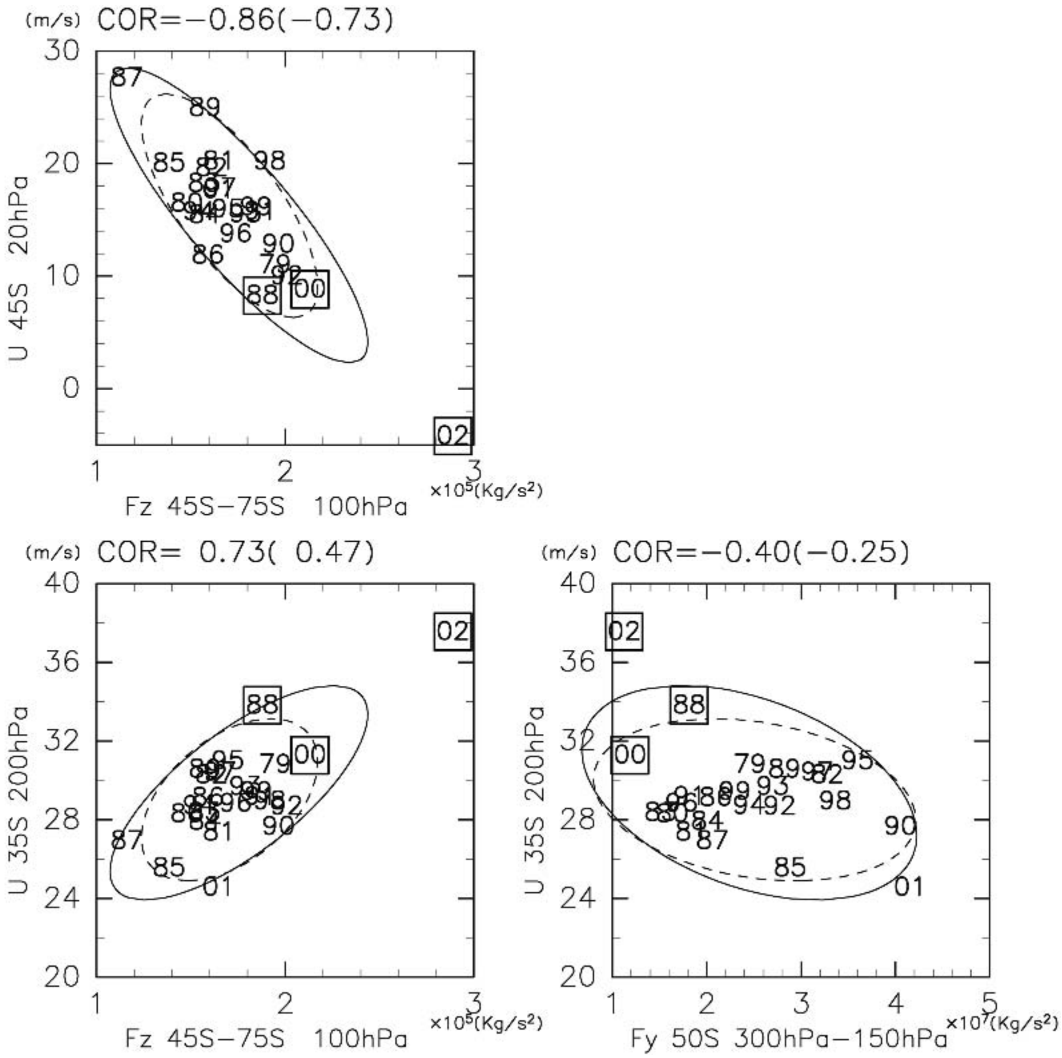

FIG. 10. Scatter diagrams (top) between the upward EP flux at 100 hPa averaged for 45°–75°S and the zonal-mean zonal wind at 45°S, 20 hPa, (bottom left) between the same upward EP flux and the zonal-mean zonal wind at 35°S, 200 hPa, and (bottom right) between this tropospheric mean zonal wind and the horizontal EP flux at 50°S averaged for 300–150 hPa. The time-averaging period for the mean zonal winds is 15 days from 1 Oct to 15 Oct, while that for the EP fluxes is 45 days from 16 Aug to 30 Sep. The two values above each panel are the correlation for the 24 and 23 yr without 2002 (written in parentheses). The solid-line (dotted-line) ellipse denotes the 90% confidence limit for the 24 yr (23 yr).

that the jet decelerates earlier in HLJ years than in LLJ years. LLJ years appear in LD years defined in this paper. However, the correspondence of ED years to HLJ years is not clear, as shown in Table 1. On the other hand, the leading mode of the multiple EOFs for the variations of the seasonal march obtained by Kuroda and Kodera (1998) reveals very similar features as our composite difference between ED and LD years shown in Fig. 6. The year-to-year variation of the seasonal march is basically explained by the variation in the timing of slow poleward and downward shift of the polar-night jet during winter.

The composite difference of the zonal-mean zonal wind at 20 hPa (Fig. 5) shows the significant difference in the equatorial wind from July to December. It is easterly (westerly) in ED (LD) years. This relation reminds us of the result of a composite analysis by Naito (2002) based on the QBO phase of the equatorial mean zonal wind in July averaged between 20 and 30 hPa for the 17 yr from 1979 to 1996. Figure 2 in Naito's paper resembles our Fig. 5 (bottom left), and the time series of the QBO category listed in our Table 1 shows a good correspondence with our time series of ED or LD years. However, such a correspondence is less clear for the recent years following 1997. In the year 2002, the QBO category is westerly, although the seasonal march is much earlier than in the other 23 yr.

In section 2, it was found that the tropospheric EP

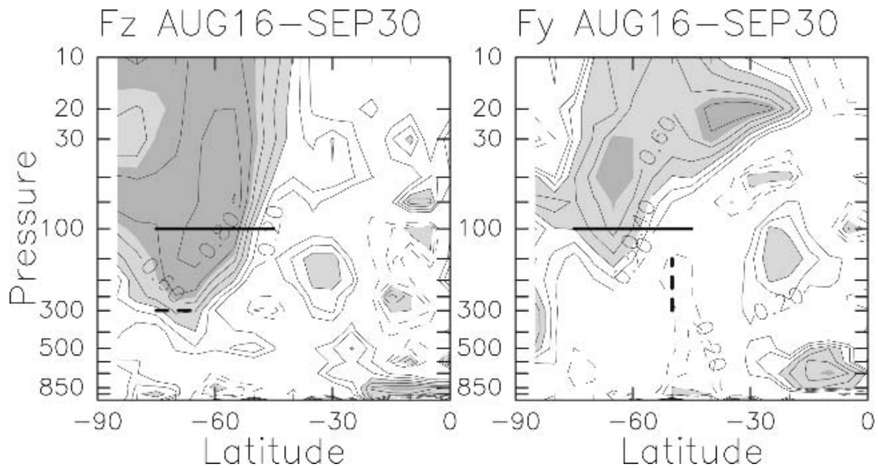

FIG. 11. Correlation of the 24-yr time series of the (left) upward and (right) equatorward EP flux averaged 45-day period from 16 Aug to 30 Sep with the upward EP flux at 100 hPa for 45°–75°S for the same period. Contours are drawn above 0.2 or below −0.2 with an interval of 0.1, and values over ±0.4 (0.65) are lightly (heavily) shaded. Dotted contours denote negative values.

flux anomalies as well as the stratospheric ones show the significant difference between the year 2002 and the average for LD years (Fig. 7). Here we note the horizontal scale of waves to make the difference by decomposing the EP flux into each wavenumber component. As for the difference of the upward EP flux in August, wave 1 is responsible for the large positive anomaly around 80°S, indicating that vigorous wave 1 in the stratosphere in 2002 is generated in the lower troposphere at fairly high latitudes. On the other hand, EP flux anomalies in the lower and middle troposphere in October and November are largely related to synoptic-scale waves. The active latitudes of the synoptic-scale waves shift equatorward in 2002 compared with those in LD years, while the location of the tropospheric jet shifts poleward.

These features about the tropospheric jet and synoptic-scale waves are consistent with the previous studies about the zonal flow vacillation in the SH troposphere (e.g., Karoly 1990; Hartmann and Lo 1998). Limpasuvan and Hartmann (2000) also showed that the SH annular mode (SAM) is maintained by the momentum flux anomaly of synoptic-scale eddies. The SAM in the stratosphere, which is associated with the upward EP flux propagating into the stratosphere, extended down into the troposphere in October 2002 (Fig. 7 in Baldwin et al. 2003). The SAM in the troposphere was maintained mainly by the horizontal EP flux anomaly caued by the synoptic-scale waves as described above. The present result suggests that the main disturbance to maintain the SAM is different between the stratosphere and the troposphere, even though we sometimes regard the SAM as a single "mode."

The correlation map shown in Fig. 11 indicates that the variations of the upward EP flux in the lower stratosphere have some association with those of wave activity in the troposphere. In Fig. 12 (top), the scatter diagram of the EP flux at the 100-hPa level between 45° and 75°S with the tropospheric vertical component (left) or the horizontal component (right) of the EP flux is shown; as is the upward EP flux at 300 hPa between 65° and 75°S or the equatorward EP flux at 50°S between 300 and 150 hPa (described as the dashed lines in Fig. 11). The upward EP flux in the upper troposphere in 2002 is the largest, while the equatorward EP flux in 2002 is the smallest compared with the other 23 yr. As a result, the year 2002 is isolated from the other years in both of the scatter diagrams. The difference in the orientation angle of the confidence ellipses with and without the year 2002 also shows the uniqueness of 2002. The scatter diagram of the equatorward EP flux with the upward EP flux in the upper troposphere (Fig. 12, bottom) shows significant correlation ($r = 0.50$) for the 23 yr from 1979 to 2001. However, the year 2002 is far from this relationship, which is located near the minor axis rather than the major one of the confidence ellipse for the other 23-yr dataset, and the correlation coefficient is reduced to 0.31 if this year is included in the calculation. Newman and Nash (2005) investigated the relationship between the upward EP flux in the lower stratosphere and the quasigeostrophic potential vorticity meridional gradient averaged between 400 and 100 hPa in midlatitudes and found that these quantities averaged over the winter season are highly correlated. This suggests that the valve for the upward propagation of the planetary waves near the tropopause is related to the wave energy that has entered the stratosphere and the tropospheric forcing. Further studies on the interannual variations in the troposphere that are related to the upward EP flux in the lower stratosphere, as the analysis on hemispheric geopotential height patterns done by Salby and Callaghan (2002) for the NH, will be

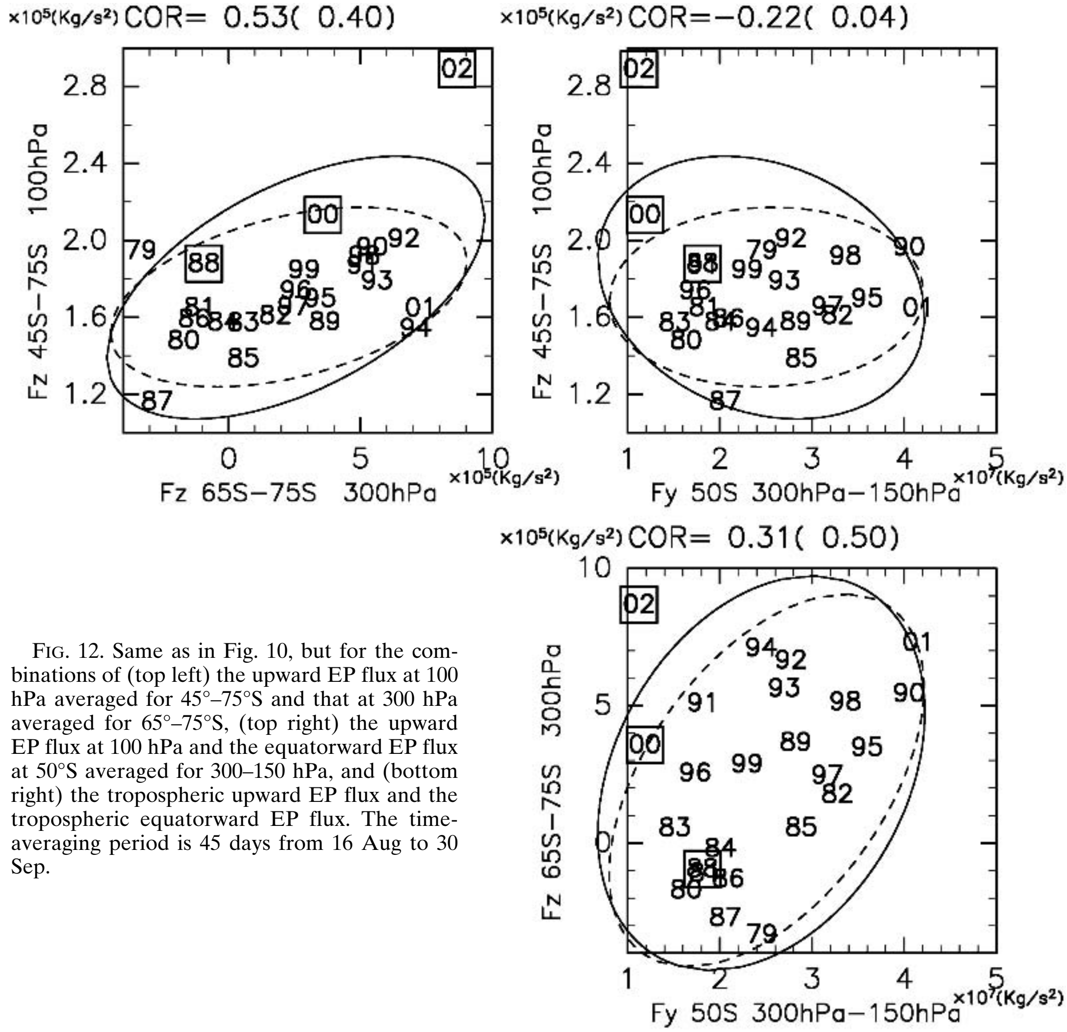

FIG. 12. Same as in Fig. 10, but for the combinations of (top left) the upward EP flux at 100 hPa averaged for 45°–75°S and that at 300 hPa averaged for 65°–75°S, (top right) the upward EP flux at 100 hPa and the equatorward EP flux at 50°S averaged for 300–150 hPa, and (bottom right) the tropospheric upward EP flux and the tropospheric equatorward EP flux. The time-averaging period is 45 days from 16 Aug to 30 Sep.

important to clarify what kind of disturbances in the troposphere modulate the upward EP flux in the lower stratosphere.

The frequency distribution of the PC1 score (Fig. 3) for the 24 yr shows a positively skewed distribution as listed in Table 2. A similar distribution (with opposite sign) is obtained for the mean zonal wind in the stratosphere. The distribution of the SAM index at 10 hPa shown by Thompson et al. (2005) also appears to have the skewness. Taguchi and Yoden (2002) also obtained a highly skewed distribution of the polar temperature in the upper stratosphere for the months through winter and spring in the 1000-yr integrations of a simple troposphere–stratosphere coupled model corresponding to the SH. A large number of the sampling guarantees the non-Gaussian distribution, and about 0.2% of the 1000 yr are extremely warm with monthly temperature anomalies exceeding five times the standard deviation. The skewed probability distribution requires careful treatment in the statistical arguments of the rarity. Here caution is necessary on the skewness of the observed data because of the small sample number. Correlation analysis with a small number of the sampling also requires careful treatment; the correlation coefficients in some scatter diagrams in Figs. 10 and 12 change drastically by excluding the year 2002 from the calculation.

5. Conclusions

A major stratospheric sudden warming event occurred in September 2002 for the first time in the Southern Hemisphere (SH) since satellite observations of the stratosphere began. We investigated the interannual variations in the SH for the 24 yr from 1979 to 2002, in terms of the seasonal evolution and the troposphere–stratosphere coupling, and characterized the dynamical features of the unprecedented year 2002 by comparing it with the other 23 yr.

Interannual variations of the seasonal march were extracted with a multiple EOF analysis of the zonal-mean zonal wind in the stratosphere. The leading mode explains the variations in the timing of the deceleration of the polar-night jet (Fig. 2). We defined two groups, early-deceleration (ED) years and late-deceleration (LD) years, according to the value of the principal component of EOF1 (Fig. 3) and made a composite analysis.

The seasonal march characterized by the poleward and downward shift of the polar-night jet from winter to spring is earlier in ED years as shown in Figs. 4, 5, and 6. The amplitude of the planetary waves in late winter is also larger for ED years, which is consistent with the wave–mean flow interaction theory. The weaker mean zonal wind in the high-latitude stratosphere in ED years, which is associated with the stronger upward EP flux in the stratosphere, can be traced down into the lower troposphere in September and October (Fig. 6). The location of the subtropical jet and the horizontal EP flux in the midlatitude troposphere also shows some variation between ED and LD years.

These features are consistent with the variations of the SH annular mode. However, the main influence for maintaining the variationsis different between the stratosphere (planetary waves) and the troposphere (synoptic-scale waves). The present results were also compared with some previous studies of the interannual variations of the SH (Table 2). The longer dataset shows relationship with the QBO of the equatorial mean zonal wind less clearly.

In 2002, the upward EP flux in the lower stratosphere was much larger in August even compared with the other ED years, and the large wave driving caused the earlier and larger deceleration of the mean zonal wind in the stratosphere (Figs. 7 and 8). The tropospheric circulation in 2002 also showed outstanding differences from LD years, as shown in Fig. 7. The polar-front jet is weak, and the subtropical jet shifts poleward in September and October. The upward EP flux in the high-latitude troposphere is weak in association with the weak polar-front jet.

The upward EP flux in the lower stratosphere is a good indicator of the interannual variations of the seasonal march of the stratospheric circulation through winter and spring (Fig. 9). The stronger wave activity in the lower stratosphere is associated with the earlier shift-down of the polar-night jet. Some scatter diagrams show the extreme situation of the year 2002 (Fig. 10): the upward EP flux is large in the lower stratosphere, the zonal-mean zonal wind is weak in the midlatitude stratosphere and strong in the upper troposphere, and the equatorward EP flux is small in the upper troposphere. However, the large deviations in 2002 are consistent with the fluctuations in the other 23 yr except for their extreme nature. Two other years (1988 and 2000) in ED years were also found in a similar situation to the year 2002 but have smaller deviations.

As stated in the last paragraph in section 4, observational studies like the present one always suffer from the limitation of the data length. The dataset for 24 yr is not long enough to obtain a stable frequency distribution, but there is no other data of the real atmosphere. It will be important for us to use numerical models to supplement the statistical arguments based on much longer datasets.

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