A Parameter Sweep Experiment on the Effects of the Equatorial QBO on Stratospheric Sudden Warming Events

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ABSTRACT

The effects of the equatorial quasi-biennial oscillation (QBO) on stratospheric sudden warming (SSW) events are investigated by performing long time integrations with a simple global circulation model under a perpetual winter condition. Zonal momentum forcing is imposed to produce a westerly or easterly phase of the QBO in the equatorial stratosphere, and a parameter that determines strength and direction of the forcing is swept for nine values as an experimental parameter.

The polar night jet is weaker and polar stratosphere is warmer in the runs with easterly “QBO wind” forcing, in qualitative agreement with the observational result. The polar night jet is strongest for a moderate westerly forcing, while it becomes weaker for too strong westerly forcing.

Occurrence of SSW events is irregular, intermittent, and least frequent in the run with the moderate westerly forcing. As a result, frequency distribution of the polar stratospheric temperature shows the largest skewness.

The frequency distribution of polar temperature is systematically dependent on the QBO wind forcing not only in the stratosphere but in the troposphere. The dependence is statistically significant even in the troposphere in spite of heavy overlap of the frequency distributions.

Composite analysis for a large number of SSW events shows downward influence of the upper-stratospheric warming to the lower levels in two characteristic timescales. Short-time cooling response within several days is extended to the summer hemisphere, while long-time response persisting a couple of weeks is confined within the winter hemisphere and extends down to the level near the tropopause. The response of polar temperature in the lower stratosphere shows systematic dependence on the QBO wind forcing; mean warming rate is smallest for the moderate westerly forcing, and warming is more rapid for the easterly forcings.

1. Introduction

The stratospheric sudden warming (SSW) event is a highly nonlinear process on a day-to-day timescale due to planetary wave–mean flow interactions (e.g., McIntyre 1982); major SSW events are associated with breakdown of the polar vortex. Occurrence of SSW events is an important factor in intraseasonal and interannual variations of winter stratospheric circulation in the Northern Hemisphere (NH). The equatorial quasi-biennial oscillation (QBO), on the other hand, dominates the variability of the equatorial stratosphere (e.g., Baldwin et al. 2001). Observational studies on the relationship between the phase of the QBO and the winter stratospheric circulation can be traced back to the pioneering work by Holton and Tan (1980). When the phase of the QBO is defined at the levels around 50 hPa, the stratospheric polar vortex is weaker, warmer and more disturbed during winters in the easterly phase of the QBO (Holton and Tan 1980, 1982), and a major SSW tends to occur in the easterly phase (Labitzke 1982). The weaker polar vortex in the easterly phase of the QBO is associated with larger upward component of the Elsassen–Palm (EP) flux from the troposphere to the stratosphere (Dunkerton and Baldwin 1991). In the Holton–Tan relationship described above, modulation of the propagation route of planetary waves associated with the QBO, for example, latitudinal shifts of the zero-wind line (i.e., a critical line for stationary waves), is considered to play an important role.

Dependence of the winter stratospheric circulation on the QBO phase has been well investigated for time-mean states such as monthly mean or seasonal mean. As for the effects of the equatorial QBO on SSW events, most of the discussions were associated with precondition for the SSW events (e.g., Labitzke 1982). However, characteristics of daily evolution during and after the events depending on the QBO phase have drawn little attention. Dunkerton and Baldwin (1991) made a composite analysis for the events of amplified planetary waves, but they found only slight differences between the easterly and westerly phases: slightly more events, slightly larger events, and a larger mean flow response in the easterly phase.

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Numerical studies have been done to investigate the effect of the QBO on the extratropical circulation in the last decade or so (e.g., Holton and Austin 1991; O’Sullivan and Young 1992; O’Sullivan and Dunkerton 1994; Niwano and Takahashi 1998; Hamilton 1998; Gray et al. 2001, 2003). Their aims were to confirm the Holton–Tan relationship, and to investigate the mechanism that a weaker and warmer polar vortex tends to appear in the easterly phase of the QBO. These studies can be classified into two groups by the model they used: a stratosphere-only model or a troposphere–stratosphere coupled model (Yoden et al. 2002).

In the stratosphere-only models, the planetary waves were prescribed at the bottom boundary placed near the tropopause. Holton and Austin (1991) examined how readily an SSW event occurs within a winter season in the westerly or easterly QBO phases. Their result revealed the Holton–Tan relationship only in the cases of intermediate amplitude of the tropospheric wave forcing; an SSW event occurred in midwinter when the equatorial wind was easterly, while the polar vortex persisted till a final warming occurred when the equatorial wind was westerly. They discussed the prewarming condition for the SSW event with EP flux diagnostics, but did not remark on any aftereffect of the SSW event. O’Sullivan and Young (1992) studied sensitivity of the time-mean states of winter polar circulation to the equatorial zonal wind profile. They interpreted the model’s tropical–extratropical coupling in terms of the propagation of Rossby wave activity toward the low-latitude easterlies, and the development of a nonlinear critical layer north of these easterlies. O’Sullivan and Dunkerton (1994) simulated the seasonal development of middle atmosphere circulation in opposite phases of the QBO and found that seasonal integrations produced a more realistic extratropical response to the QBO, making the “bifurcation” (which divides parameter ranges of wave amplitude with/without a major warming) less conspicuous. In these three studies, the QBO wind profile was superposed on the climatological wind field as an initial condition.

Recently, a series of multiyear integrations were performed with a stratosphere-only model by Gray et al. (2001). In contrast to the previous studies, the winter extratropical circulation was not sensitive to the QBO phase when the equatorial winds were relaxed toward Singapore radiosonde observations covering only the lower stratosphere, whereas the Holton–Tan relationship was obtained when the equatorial winds were relaxed toward rocketsonde observations extending to the upper stratosphere. Gray et al. (2003) employed another approach by performing two series of ensemble experiments consisting of 20 runs with different initial conditions, so that they could assess not only ensemble mean but also variance among the ensemble members dependent on the lower boundary and on the equatorial “sidewall” boundary. Since the amplitude of the tropospheric planetary waves was prescribed near the tropopause as the bottom boundary condition in these stratosphere-only models, natural tropospheric variations were excluded and any downward influence from the stratosphere to the troposphere was not represented in the models.

The second group of the numerical studies (Niwano and Takahashi 1998; Hamilton 1998) used general circulation models (GCMs), in which many complicated physical processes are parameterized and the troposphere is included in the computational domain. Their results are both consistent with the Holton–Tan relationship. Niwano and Takahashi (1998) examined the effects on extratropics of the QBO that was generated internally in their model with fine vertical resolution and suppressed horizontal diffusion, though the period of their “QBO” (approximately 1.3 yr) was much shorter than that observed. By composite analysis with respect to the QBO phases, they showed that the anomaly pattern of time-mean zonal wind in the troposphere reveals a tripole structure among low, mid-, and high latitudes, similar to the North Atlantic Oscillation (NAO) pattern. Hamilton (1998) imposed zonal momentum forcing in the tropical stratosphere in the Geophysical Fluid Dynamics Laboratory (GFDL) SKYHI model and produced a fairly realistic QBO. He found that the effects of the imposed QBO on the troposphere were very modest; a statistically significant weakening of the wintertime polar vortex was in the middle and upper troposphere. In these GCM studies, multiyear integrations with the descent of the westerly and easterly phases of the QBO were performed but not repeated for many runs, and time-mean fields were mainly analyzed.

Yoden et al. (2002) emphasized that parameter sweep experiments are important to investigate nonlinear phenomena such as the internal variations of the troposphere–stratosphere coupled system in mid- and high latitudes. They discussed hierarchy of numerical models to understand the stratospheric variations, and highlighted the use of three-dimensional mechanistic circulation models (MCMs), in which physical processes are very simplified. Taguchi et al. (2001, hereafter referred to as TYY) performed a parameter-sweep experiment with an MCM extended up to the mesosphere to investigate the internal variability in the troposphere–stratosphere coupled system. Changing the amplitude of a sinusoidal surface topography of zonal wavenumber 1 $h_0$, as an experimental parameter from 0 to 3000 m, they classified the resultant intraseasonal variability in the extratropical winter stratosphere into four regimes dependent on the parameter $h_0$. Intraseasonal variability similar to the Southern Hemisphere (SH) is obtained for small amplitudes of the topography around $h_0 = 500$ m, while that similar to the NH is obtained for larger amplitudes around $h_0 = 1000$ m.

In the present study, the same MCM as was used by TYY is employed to investigate the effects of the equatorial QBO on the wintertime intraseasonal variations of the extratropical stratospheric circulation. Zonal mo-
momentum forcing is imposed in the equatorial stratosphere to produce a westerly or easterly phase of the QBO. The model domain covering the troposphere will enable us to obtain internal irregular variations of planetary waves generated in the troposphere and possible downward influence from the stratosphere to the troposphere. Composite analysis for a large number of the obtained SSW events is made to investigate daily evolution of the temperature field during the event, particularly focusing on the aftereffect to the lower levels. The model is integrated over 10 000 days under a perpetual winter condition in nine runs of different QBO forcings so that not only time-mean states but also time variation can be analyzed with estimation of statistical significance. By sweeping the experimental parameter that determines the strength and direction of the forcing, systematic dependence of the characteristics of SSW events on the equatorial wind profile will be obtained.

The present paper is organized as follows. Section 2 describes the present MCM and the experimental setup. Results are shown in the following three sections. Section 3 overviews time-mean states, and time variations of the polar temperature are described in section 4. The results of composite analysis for SSW events are in section 5. Discussion is in section 6 and conclusions in section 7.

2. Model and experimental setup

The model used in this study is basically the same as in TYY, and the details are documented there. The model is a three-dimensional global primitive-equation model (Swamp Project 1998), which explicitly describes large-scale motions with a horizontal resolution of T21 spherical-harmonics truncation and a vertical representation of 42 σ levels (σ = pl/psurface; p is pressure) from the surface to the mesopause. The model includes simplified physical processes such as Newtonian heating/cooling to a perpetual winter condition in the model NH, dry atmosphere without moist processes, Rayleigh friction at the surface, and so on. A sinusoidal surface topography of zonal wavenumber 1 and amplitude h0 = 1000 m is included in the model NH; the result by TYY showed that the parameter range around h0 = 1000 m was corresponding to a regime close to the intraseasonal variations in the NH winter stratosphere with intermittent occurrence of SSW events.

To produce a westerly or easterly phase of the QBO, an additional source term is imposed in the zonal momentum equation:

\[ \frac{\partial u}{\partial t} = \cdots - \alpha_{QBO}(u - U_{QBO}), \]

where \( u(\lambda, \phi, \sigma, t) \) is the local zonal wind at any point of longitude \( \lambda \), latitude \( \phi \), and level \( \sigma \); and a relaxation rate \( \alpha_{QBO} \) and a profile \( U_{QBO} \) of the perpetual "QBO wind" forcing are defined as prescribed functions of \( \phi \) and \( \sigma \) as follows:

\[
U_{QBO}(\phi, \sigma) = a \times 45 \times \gamma(\phi, \sigma) \cos\left(\frac{\xi - 3}{6}\right) \quad \text{[m s}^{-1}\text{]},
\]

\[
\alpha_{QBO}(\phi, \sigma) = |a| \times \frac{1}{30} \times \gamma(\phi, \sigma) \quad \text{[day}^{-1}\text{]},
\]

where \( \xi = -\log\sigma \); and \( \gamma(\phi, \sigma) \) is a weighting function to confine both \( U_{QBO} \) and \( \alpha_{QBO} \) to the equatorial lower stratosphere defined by

\[
\gamma(\phi, \sigma) = \begin{cases} 
\exp\left[-\left(\frac{\phi}{17}\right)^2\right], & \text{for } \xi < 2.7; \\
\exp\left[-\left(\frac{\xi - 2.7}{0.68}\right)^2\right], & \text{for } 2.7 < \xi < 3.6; \\
\exp\left[-\left(\frac{\xi - 3.6}{1.37}\right)^2\right], & \text{for } \xi > 3.6
\end{cases}
\]

(cf. Horinouchi and Yoden 1997).

The factor \( a \) in Eqs. (2) and (3) is changed as an experimental parameter from \( -2 \) to \( +2 \) with an increment of 0.5. The case of \( a = 0 \) is the control run (CTRL) in which no QBO wind forcing is given. The equatorial zonal wind in the lower stratosphere is forced to be westerly in the four runs of \( a > 0 \) (W0.5, W1.0, W1.5, and W2.0), while it is forced to be easterly in the other four runs of \( a < 0 \) (E0.5, E1.0, E1.5, and E2.0). Here the number after "W" or "E" in each experimental code denotes the strength of the forcing \( |a| \). Figure 1 shows the nine vertical profiles of \( U_{QBO} \) at \( \phi = 0^\circ \) (the equator). The cases of \( a = \pm 1 \) (W1.0 and E1.0) roughly correspond to the range of the QBO in the real atmosphere. In each of nine runs, time integration is performed with \( \Delta t = 15 \) min. for 12 000 days with a perpetual winter condition. Daily data for the last 10 800 days, excluding the initial transient time, are used in the present analysis. The analysis is done after the data are interpolated to surfaces of constant pressure.

3. Time-mean states

Figure 2 shows time-mean fields of the zonal-mean zonal wind \([u]\) in the nine runs from W2.0 to E2.0. Here, square brackets denote the zonal mean and an overbar does the time mean. The central panel for CTRL is equivalent to Fig. 1e of TYY. As they discussed, general features in CTRL agree qualitatively with those observed in the real atmosphere: the westerly subtropical jet in the both hemispheres, the westerly polar night jet in the extratropical stratosphere in the NH, and the easterly jet in the stratosphere in the SH. Similar features are also seen in the other runs. In response to the QBO
wind forcing, $\overline{u}$ in the equatorial stratosphere is westerly in the runs with westerly forcing and is easterly in the runs with easterly forcing. The maximum of $\overline{u}$ in the equatorial stratosphere is about $+40$ m s$^{-1}$ in W1.0 and $-40$ m s$^{-1}$ in E1.0; these values are roughly corresponding to (but somewhat stronger than) the range of the QBO in the real atmosphere. The maximum reaches about twice of these values in W2.0 and in E2.0, respectively. Note that the zero wind line in the equatorial lower stratosphere shifts to higher latitudes from CTRL to E2.0.

Influence of the QBO wind forcing to the extratropical stratospheric circulation in the NH appears as a systematic change in strength and position of the winter polar night jet. The strength of the polar night jet is summarized in Fig. 3 (solid line with closed circles). The polar night jet is strongest in W1.0, and it becomes weaker monotonically to E2.0. This dependence on the equatorial wind is consistent with the Holton–Tan relationship. The rate of weakening of the polar night jet also decreases in the parameter range from W1.0 to E2.0. As the polar night jet becomes weak, the position of the jet maximum slightly shifts poleward and downward in this parameter range (Fig. 2). On the other hand, in the parameter range from W1.0 to W2.0, the polar
night jet becomes weaker as the westerly forcing becomes stronger.

Figure 4 shows the difference of time-mean zonal wind between W1.0 and E1.0. A positive value over the equator in the middle and lower stratosphere is the difference of the direct responses to the QBO wind forcing, and the positive one poleward of 30°N in the upper stratosphere indicates the stronger polar night jet in W1.0 than in E1.0, as was seen in Fig. 2. In addition, the difference is negative in the high-latitude lower stratosphere in the NH, which is due to the slight shift-down of the polar night jet in the easterly forcing runs mentioned above. Note that the negative difference in the polar region of the lower stratosphere extends down to the surface, though the difference is very small in the troposphere.

In order to see the propagation characteristics of planetary waves, squared refractive index (see Andrews et al. 1987) $n^2$ for the stationary planetary wavenumber 1 is computed from the time-mean wind fields $\bar{u}$ (Fig. 5). In E1.0 and E2.0, $n^2$ in the equatorial region is negative because of the easterly wind, which indicates that the stationary planetary waves cannot propagate there. In W1.0 and W2.0, on the other hand, it is positive in the equatorial lower stratosphere and allows the planetary waves to propagate into the equatorial region. In the parameter range from W1.0 to E2.0, latitudinal width of the positive $n^2$ region in the NH becomes narrower as the easterly equatorial wind becomes stronger. It should be noted that $n^2$ in the equatorial lower stratosphere in W2.0 is positive but not so large as that in W1.0, implying more refraction of stationary waves to higher latitudes. Accordingly, the planetary waves are confined to middle and high latitudes more easily in the runs with very strong westerly forcing.

The time-mean field of the zonal-mean temperature $\bar{T}$ basically satisfies a thermal wind balance with $\bar{u}$ (see Fig. 1b of TYY for $\bar{T}$ equivalent to our CTRL). The difference of $\bar{T}$ between W1.0 and E1.0 is shown in Fig. 6 in order to see the QBO effects. Over the equator, a warm anomaly in the W1.0 appears around $p = 100$ hPa and a cold anomaly appears around $p = 10$ hPa. Over the subtropics, on the other hand, two cold anomalies appear around $p = 100$ hPa in both hemispheres and two warm anomalies appear around $p = 10$ hPa. This six-box pattern of temperature anomalies in the low-latitude stratosphere is in a thermal wind balance with the anomaly of $\bar{u}$ associated with the vertical shear above and below the maximum of the QBO wind response as shown in Fig. 4. Mean meridional circulation is expected to maintain these temperature anomalies as was pointed out by Plumb and Bell (1982). In the winter polar region in the NH, a large cold anomaly in W1.0 appears in a height range from $p = 100$ hPa to 1 hPa and a warm anomaly appears above $p = 0.5$ hPa, and vice versa in the midlatitudes, although the cold anomaly above $p = 1$ hPa is subtle. This four-box pattern of temperature anomalies is consistent with the stronger polar night jet in W1.0. Stronger mean meridional circulation in E1.0 induced by larger Rossby wave driving is expected to maintain this pattern of temperature anomalies with opposite sign. Difference of $\bar{T}$ in the troposphere is very small, but there are a warm anomaly around 50°N and cold ones in the polar region and around 30°N.

Polar temperature in the upper stratosphere for the nine runs is also summarized in Fig. 3 (dashed line with stars). Dependence of the polar stratospheric temperature on the QBO wind forcing is consistent with that of the polar night jet: the coldest polar stratosphere associated with the strongest polar vortex in W1.0, monotonic increase of $\bar{T}$ in the parameter range from W1.0 to E2.0, and increase from W1.0 to W2.0. Note that the difference of $\bar{T}$ between W1.0 and E1.0 is nearly 15 K, but it is only 2 K between E1.0 and E2.0.
Fig. 5. Latitude–height sections of squared refractive index $n^2$ evaluated for a stationary wave of zonal wavenumber 1 from [u] in five runs of W2.0, W1.0, CTRL, E1.0, and E2.0. They are multiplied by a factor of $a^2$, where $a$ is radius of the earth. Contour interval is 5.

Fig. 6. A latitude–height section of the difference of $T$ between W1.0 and E1.0. Contour interval is 2 K. Dashed contour lines denote negative values, that is, higher temperature in E1.0 than in W1.0. Values more than 6 K are shaded with dots and those less than −6 K are shaded with diagonal lines.

4. Time variations of the polar temperature

Time variations of $T$ around the North Pole in the upper stratosphere are shown in Fig. 7 for the last 2000 days in each run. Many SSW events are seen for all the runs. Here we define an SSW event from this time series as a period during which $T$ is continuously above 235 K and its maximum is greater than 270 K (cf. Yoden et al. 1999). Denoted by a circle is the day of maximum $T$ in each event period. Total number $N$ of the defined SSW events in 10 800 days is shown on the right side of the panel for each run in Fig. 7. In W1.0, the occurrence of SSW events is least frequent of all the runs, and it is irregular and intermittent. The total number $N$ increases monotonically from W1.0 to E2.0, and the occurrence is no longer intermittent in E2.0. On the other hand, in the parameter range from W1.0 to W2.0, $N$ increases as the westerly QBO wind forcing becomes stronger. The dependence of $N$ on the QBO wind forcing is directly related with the time-mean temperature $T$ shown in Fig. 3.

Frequency distributions of $T$ for 10 800 days are plotted in Fig. 8 for four runs (W2.0, W1.0, E1.0, and E2.0) overlaid on that of CTRL (shaded with dots). All of these distributions are far from the Gaussian distribution. The distributions in the parameter range from W2.0 to CTRL have similar characteristics; most part of 10 800 days are in a low-temperature state around 220 K and a limited number of days exhibit high temperature up to 300 K associated with SSW events. Skewness of the distributions is large, and the largest skewness is found in W1.0, which is the run with least frequent occurrence of SSW events. The skewness decreases monotonically from W1.0 to E2.0. In E1.0 and E2.0, a hint of bimodal distribution can be seen with frequent occurrence of SSW events.

In the troposphere, frequency distributions of $T$ around the North Pole are close to Gaussian distribution...
in all the nine runs. Figure 9 shows two examples for W1.0 and E1.0. These two distributions heavily overlap each other. The difference between the time means for these two runs is about 1 K, which is smaller than the standard deviation of either distribution. Here, statistical significance of the difference between the two mean values is tested by a method described in the appendix, which is based on the textbook by Hoel (1984). By using the sample size (10 800 for both datasets), the difference between the two mean values (1 K) and the standard deviations (1.87 K for W1.0 and 1.75 K for E1.0, respectively), the threshold for a standard normal variable Z is obtained as 40.6. It implies that the probability that the difference between means of two samples reaches 1 K is much less than $10^{-5}$% if two populations from which the samples are extracted, respectively, have the same mean as each other. Since this probability is quite small, the difference between the two time means of tropospheric polar temperature in W1.0 and E1.0 is significant. The large number of the sample size gives the statistical significance in spite of heavy overlap of the two distributions.

In order to compare statistical characteristics of the occurrence of SSW events dependent on the QBO wind forcing, cumulative frequency distributions for the interval of two consecutive SSW events are plotted in Fig. 10. This plot resembles Fig. 6 of Taguchi and Yoden (2002), which displays intervals of two consecutive anomalously warm winters in the 1000-yr integration of the TYY model with a seasonal cycle. All the distributions are roughly exponential so that they are fitted with straight lines on the linear-log scale plot. As the interval between successive events has an exponential distribution, the number of events that occur in a given unit time is supposed to have a Poisson distribution; that is, the events are supposed to occur at random. The gradient of the slope of the fitting line determines the mean interval $\tau$ of SSW events. The value of $\tau$ becomes smaller in the parameter range from W1.0 to E2.0, reflecting shorter intervals and more frequent occurrence of the events in the easterly forcing runs. The mean error $\epsilon$ is smallest in CTRL, and it is larger in the easterly forcing runs than in the westerly forcing runs. For short intervals ($<30–60$ days), the cumulative frequency does not decrease very much as the interval increases, indicating the fact that an SSW event tends not to occur within such a short interval.
FIG. 8. Frequency distributions of $[T]$ at $\phi = 86^\circ$N and $p = 2.6$ hPa for 10 800 days in four runs: (a) W2.0, (b) W1.0, (c) E1.0, and (d) E2.0 (shaded with diagonal lines), compared with CTRL (shaded with dots). Note that (b) has a different vertical scale from (a), (c), and (d). Width of a bin is 2 K. A solid vertical line denotes the time-mean temperature $[T]$ in each of the four runs, and a dashed one does $[T]$ in CTRL. Skewness of the distribution for each of the four runs is shown in each, and that for CTRL is 0.720.
5. Composite analysis for the SSW events

Figure 11 shows composites of the time variation of polar temperature before, during and after SSW events for three stratospheric levels. The key day, corresponding to lag = 0 day in the abscissa, is defined as the day of maximum $[T]$ at $p = 2.6$ hPa in each event (denoted by a circle in Fig. 7). At this level, the SSW events appear as a sharp increase of $[T]$ by $\sim 50$ K from lag $\sim -10$ days, rather independently of the QBO wind forcing except for W2.0. Cooling rate just after the key day is comparable to the radiative cooling rate (relaxation time is about 5 days), and soon after that the cooling rate becomes smaller than the radiative cooling rate, implying that large fraction of the radiative cooling is cancelled by the effects of dynamical processes. There is some variation depending on the QBO wind forcing before lag = $-20$ day, which is related to the occurrence of the previous SSW event.

Some dependence of the time variation on the QBO wind forcing is seen in the middle stratosphere ($p = 12$ hPa). Before the sudden warming starting around lag = $-10$ day, $[T]$ decreases from relatively warm state in the easterly forcing runs, while gradual increase of $[T]$ from relatively cold state starts earlier than lag = $-10$ day in the westerly forcing runs. There is little difference during the warming period among the nine runs, but the maximum of $[T]$ is lower (over 5 K) in the westerly forcing runs (except for W2.0) than in the easterly forcing runs. In all the nine runs, the decay after the event is more gradual than that at $p = 2.6$ hPa, and much more gradual than the radiative cooling at this level (the relaxation time is still 12 days at this level).
Fig. 11. Composite time series of \([T]\) for the SSW events in the nine runs, at the latitude \(\phi = 85^\circ\text{N}\) and (top) \(p = 2.6\) hPa, (middle) 12 hPa, and (bottom) 120 hPa. The key day corresponds to lag = 0 day.

The effect of the QBO-wind forcing is more evident and systematic in the lower stratosphere \((p = 120\) hPa), even though the temperature variation is small. In the easterly forcing runs, \([T]\) still increases rather rapidly and the maximum of \([T]\) appears around lag \(\approx 5\) day. There is little difference among the three easterly forcing runs. On the other hand, in the westerly forcing runs, no signature of the “sudden” warming can be seen; \([T]\) increases gradually for a month or so before and after the key day. Mean warming rate of \([T]\) during a period \(|\text{lag}| \leq 10\) day is estimated by the least squares method for each of the nine runs from W2.0 to E2.0 as 0.170, 0.123, 0.097, 0.141, 0.161, 0.168, 0.211, 0.217, and 0.247 (K day\(^{-1}\)). The warming rate monotonically increases in the parameter range from W1.0 to E2.0.

Statistical significance of the difference between a pair of composites for W1.0 and E1.0 is tested by the method described in the appendix. In the present case, the sample size is the number of SSW events: 57 events in W1.0 and 168 events in E1.0. Figure 12 shows time variation of a statistic \(Z\) for each lag at the same three stratospheric levels as in Fig. 11, together with two lines of the composites with a shade showing the range of \(\pm 0.2\) times standard deviation. If the two shaded regions do not overlap each other, significance of the difference between two means is estimated as more than 99.55% (see the appendix). In the upper stratosphere \((p = 2.6\) hPa), the statistic \(Z\) is less than 2.5 for the period \(-25\) day \(\leq\) lag \(\leq 30\) day. The difference is significant only for lag \(\leq -25\) day, reflecting the occurrence of the previous SSW event with a shorter interval in E1.0. The statistical significance during SSW events increases as the pressure increases. In the lower stratosphere \((p = 120\) hPa), \(Z\) reaches 5.2, indicating the significance of
Figure 13 shows the latitude-time sections of the composite SSW events for the three runs of W1.0, CTRL, and E1.0 in the middle and lower stratosphere. In order to exclude the climatological temperature gradient in direct response to the equatorial QBO wind forcing, the time-mean value for “cold periods” is subtracted from the composites. Here the cold periods are defined for each run from the time series in Fig. 7 as the days of $T < 235$ K. Downward influence of the SSW events to the lower stratosphere has two characteristic timescales: a short-time cooling response (several days after the key day) and a long-time response (a few weeks after). The short-time cooling response extends to the SH at $p = 12$ hPa, and becomes obscure at the lower levels. It is more evident in W1.0 than in E1.0. The long-time cooling response is very confined to the NH. It is clearly seen at all three levels, and stronger in the westerly forcing run.

6. Discussion

An MCM was employed to investigate the effects of the equatorial QBO on SSW events. In contrast with stratosphere-only models, tropospheric variations are produced internally and downward influence from the stratosphere to the troposphere is also included within the present model. Simplification of physical processes in this model makes it easier to perform a parameter sweep experiment of long time integrations compared with GCMs.

By sweeping a parameter of the QBO wind forcing, systematic dependence on the parameter was found in composites of the polar temperature in the lower stratosphere just after the key day of SSW event (Fig. 11); the estimated mean warming rate becomes larger as the wind forcing becomes stronger easterly within the parameter range from W1.0 to E2.0. A sufficiently large number of SSW events—more than 50 in each run—were obtained in the long time integrations over 10 800
days, which enabled us to show a statistically significant difference of the temperature composite for the SSW events dependent on the QBO wind forcing (Fig. 12).

In the statistical tests, a large sample method can be used instead of a small sample method with Student’s t test, so that it is not necessary to assume the equality of the variances of two populations. As for the observation data of the real atmosphere, it would be difficult to use the large sample method for finding a statistically significant difference of characteristics of daily evolution during SSW events between the westerly and easterly phases of the QBO. Since major SSW events occur every two or three years, over 100-yr records for each phase of the QBO may be required to obtain comparable number of SSW events to the present analysis. Besides, other processes which possibly influence SSW events, such as 11-yr solar cycle, volcanic aerosols, El Niño–Southern Oscillation, and so on, may bring additional difficulty in extracting the QBO effect in the real atmosphere.

When the polar temperature in the upper stratosphere increases during SSW events, the mean meridional circulation induced by the westward zonal momentum forcing due to planetary wave breaking is expected according to the “downward control principle” (Haynes et al. 1991). Figure 13 shows that the lower-level response of temperature during and after an SSW event has two characteristic timescales: short-time response extended to the summer hemisphere, and long-time response confined to the winter hemisphere and extended down to near the tropopause. It can be said in other words that descent rate of the latitudinally narrower response is slower than the wider one. This is consistent with the result of Haynes et al. (1991), who considered the downward control by a time-dependent zonal momentum forcing and showed that the meridional mass circulation cell below the forcing penetrates downward more slowly for the smaller meridional scale of response.

Composites of the polar temperature in the midtroposphere (Fig. 14a) also exhibit systematic dependence on the QBO wind forcing; the temperature around lag = 0 day monotonically decreases in the parameter range from CTRL to E2.0. The composite difference between W1.0 and E1.0 is statistically significant after the key day and the maximum of the significance appears at lag = 12 day (Fig. 14b). Figure 14c shows latitude-time sections of the deviation of the composite temperature from the cold-period mean at $p = 449$ hPa. In all the three runs, a warm anomaly around 55°N and a cold anomaly around 30°N persist about a month before and around the key day, and another cold anomaly poleward of 70°N also persists around lag = −20 day. These temperature anomalies in the troposphere are large before the SSWs in contrast to those in the lower stratosphere as shown in Fig. 13.
The time mean of polar temperature in the troposphere also shows systematic dependence on the QBO wind forcing, though the difference between the two runs of W1.0 and E1.0 is small (~1 K). The large size of samples (10 800 for each run) enabled us to show a statistically significant difference between the two QBO phases for these distributions in spite of heavy overlap of the frequency distributions (Fig. 9). To find such statistically significant evidence in the real atmosphere, much longer observation would be required; regarding a winter as 90 days, for example, observation over roughly 240 yr would be required to obtain 10 800-day records for two phase of the QBO.

Difference of the time-mean zonal wind between W1.0 and E1.0 (Fig. 4) is also significant even in the troposphere; the time-mean wind in the polar region is stronger in E1.0. In the real atmosphere, on the other hand, the polar anomaly in the stratosphere indicating stronger time-mean westerly wind in the westerly phase of the QBO penetrates the polar troposphere without changing the sign (Baldwin and Dunkerton 1991). The stratospheric and tropospheric anomalies tend to be of the same sign during downward propagation of the annual mode (Dunkerton and Baldwin 1991), while they might be of the opposite sign during an SSW event including preconditioning state, because of the poleward shift of the tropospheric jet and the weaker polar night jet in the stratosphere. The present difference in the troposphere is statistically significant, but the reality might be limited because some tropospheric processes are highly idealized; sinusoidal surface topography, dry atmosphere, no heat contrast at the surface, and so on.

The interval between successive SSW events under a perpetual winter condition is distributed roughly exponentially (Fig. 10), implying that the occurrence of SSW events itself is described by a Poisson distribution, that is, the events are supposed to occur at random. The mean interval $\tau$ is about 4 months in W1.0 and about 1 month in E1.0, which is comparable with the observational fact that an SSW event tends to occur in December or January in the easterly phase of the QBO while an SSW event tends not to occur within winter season (for more than 3 months) in the westerly phase. It should be noted that $\tau$ is just a “mean” interval; some SSW events in W1.0 occur within a shorter interval than a month and some in E1.0 occur after a longer interval than 3 months. If SSW events occur at random, a trial whether an SSW event occurs within a winter season is also expected to be a random process. This idea explains the resemblance of Fig. 6 of Taguchi and Yoden (2002) and Fig. 10 in the present analysis.

7. Conclusions

In order to investigate the effects of the equatorial QBO on the stratospheric sudden warming (SSW) events, a simple global circulation model was integrated over 10 800 days under a perpetual winter condition for nine runs with a sweeping parameter of the equatorial “QBO wind” forcing.

Systematic dependence on the QBO wind forcing is commonly seen in several climatological variables. In the parameter range with realistic strength of the equatorial QBO wind, the dependence is monotonic and general features are consistent with the Holton–Tan relationship observed in the real atmosphere; weaker polar night jet and warmer polar stratosphere are found in the easterly phase of the QBO than in the westerly phase. Similar but weaker dependence is seen in the parameter range with stronger easterly wind. On the other hand, weaker polar night jet is found with very strong westerly equatorial wind. It was suggested by computing the refractive index that the propagation of planetary waves is more confined to middle and high latitudes not only in the easterly phase of the QBO but also in very strong westerly phase.

The long time integrations enabled us to study on higher moment statistics, such as skewness, in addition to the time-mean values. Frequency distribution of polar temperature in the upper stratosphere shows largest skewness in the moderate QBO wind in the westerly phase and the skewness monotonically decreases as the easterly equatorial wind becomes stronger. This is a result due to more frequent occurrence of SSW events.

Time mean of the polar temperature in the troposphere also shows systematic dependence on the QBO wind forcing. The difference between the runs with westerly or easterly forcing is statistically significant due to the large size of samples, even though the frequency distributions heavily overlap each other and the difference between two means is only 1 K.

Composites of polar temperature for many SSW events obtained in long time integrations shows systematic dependence on the QBO wind forcing in the lower stratosphere. The associated warming is more gradual in the runs with westerly forcing than in those with easterly forcing. The latitudinal structure of the composite temperature deviation from the cold periods has two characteristic timescales: short-time cooling response within several days is extended to the summer hemisphere, while long-time response persisting a couple of weeks is confined to the winter hemisphere and extends down to the level near the tropopause. Some statistically significant differences can be seen even in the troposphere.

Acknowledgments. The present graphic tools we used were based on the codes in the GFD-DENNOU Library (SGKS Group 1999). Model integrations were done on VPP800 of the Kyoto University Data Processing Center. Calculation of the thresholds for $Z$ was done with the program on the Web site by Aoki (1996). This work was supported by the Grand-in-Aid for Scientific Research of the Ministry of Education, Culture, Sports, Science, and Technology of Japan.
APPENDIX

Estimation of Statistical Significance with a Large Sample Method

Statistical significance of difference between two means can be tested as described below, based on the textbook by Hoel (1984). Supposing two samples (such as the datasets for W1.0 and E1.0) are extracted from two populations \( X \) and \( Y \), respectively, then it can be shown that a statistic

\[
Z = \frac{\bar{X} - \bar{Y} - (\mu_x - \mu_y)}{\sqrt{\frac{\sigma_x^2}{N_x} + \frac{\sigma_y^2}{N_y}}} \quad (A1)
\]

is a standard normal variable if \( N_x \) and \( N_y \) are sufficiently large (roughly 50 is sufficient). Here, \( N_x \) and \( N_y \) are sample sizes, \( \bar{X} \) and \( \bar{Y} \) are sample means, \( \sigma_x^2 \) and \( \sigma_y^2 \) are sample variances, \( \mu_x \) and \( \mu_y \) are population means of \( X \) and \( Y \), respectively. [Mean and variance of \( \bar{X} - \bar{Y} \) are equivalent to \( \mu_x - \mu_y \) and \( (\sigma_x^2/N_x) + (\sigma_y^2/N_y) \), respectively.] To test a hypothesis that \( \mu_x - \mu_y \) is identically zero, it suffices to check probability that \( Z \) reaches a threshold \( (\bar{X} - \bar{Y})/\sqrt{(\sigma_x^2/N_x) + (\sigma_y^2/N_y)} \), which is calculated from the two samples. In Table A1, some thresholds for \( Z \) are shown together with the corresponding values of the probability and significance.

In the case of the tropospheric polar temperature in section 4, the threshold 40.6 is obtained by substituting the parameters \( N_x = N_y = 10800, \bar{X} = 226.8 \, K, \bar{Y} = 225.8 \, K, \sigma_x = 1.87 \, K, \) and \( \sigma_y = 1.75 \, K \) into Eq. (A1). Since the probability that a standard normal variable \( Z \) reaches 40.6 is very small (\( \approx 10^{-5} \)), the hypothesis \( \mu_x - \mu_y = 0 \) is rejected.

In the case of the composites of temperature for SSW events in section 5, \( N_x = 57 \) and \( N_y = 168 \) are adopted. If \( \sigma_y = \sigma_y = \sigma \) is assumed, the definition of \( Z \) together with \( N_x = 57 \) and \( N_y = 168 \) provides us with

\[
Z \sim \frac{\bar{X} - \bar{Y}}{\sigma/\sqrt{1/N_x} + 1/N_y} = \frac{\bar{X} - \bar{Y}}{0.153\sigma}; \quad (A2)
\]

hence, if \( \bar{X} - \bar{Y} > 2 \times 0.2\sigma \), then \( Z > (2 \times 0.2)/0.153 = 2.609 \). Probability of \( Z > 2.609 \) is 0.45% according to the table for the standard normal distribution. After all, the difference between two means has 99.55% significance if the two regions shaded with the width 0.2\( \sigma \) do not overlap each other. This result is not changed very much even if \( \sigma_x = 0.8\sigma_y \), for instance, is assumed.

REFERENCES


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