Response of the Southern Hemisphere Atmosphere to the Stratospheric Equatorial Quasi-Biennial Oscillation (QBO) from Winter to Early Summer

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(Manuscript received 3 March 2018, in final form 2 September 2018)

Abstract

We investigate the effects of the stratospheric equatorial quasi-biennial oscillation (QBO) on the extratropical circulation in the Southern Hemisphere (SH) from SH winter to early summer. The Japanese 55-year Reanalysis (JRA-55) dataset is used for 1960–2010. The factors important for the variation of zonal wind of the SH polar vortex are identified via multiple linear regression, using equivalent effective stratospheric chlorine (EESC), middle- and lower-stratospheric QBO, solar cycle, El Niño-Southern Oscillation (ENSO), and volcanic aerosol terms as explanatory variables. The results show that the contributions to the SH polar vortex variability of ENSO are important in SH early winter (June) to mid-winter (July), while that of middle-stratospheric QBO is important from spring (September to November) to early summer (December).

Analyses of the regression coefficients associated with both middle- and lower-stratospheric QBO suggest an influence on the SH polar vortex from SH winter through early summer in the seasonal evolution. One possible pathway is that the middle-stratospheric QBO results in the SH low latitudes stratospheric response through the QBO-induced mean meridional circulation, leading to a high-latitude response. This favors delayed downward evolution of the polar-night jet (PNJ) at high latitudes (around 60°S) from late winter (August) to spring (September–November) during the westerly phase of the QBO, consequently tending to strengthen westerly winds from stratosphere to troposphere in the SH spring. The other possible pathway involves the response to lower-stratospheric QBO that induces the SH late winter increase in upward propagation of planetary waves from the...
extratropical troposphere to stratosphere, which is consistent with weakening of the PNJ.

Keywords quasi-biennial oscillation; Southern Hemisphere polar vortex; multiple linear regression

1. Introduction

Globally homogeneous spatial and temporal data coverage of the stratosphere has been provided by satellite observations since the latter half of the 1970s, and year-to-year variability and seasonal variation have been investigated, due to the accumulation of observational data, in both the Northern (NH) and Southern (SH) Hemispheres. The year-to-year variability of the stratospheric polar jet over the NH is very high during NH winter and spring, and 6 times per decade includes a complete reversal of the upper stratospheric westerlies in association with sudden stratospheric warming (SSW) (e.g., Labitzke 1977, 1982; Naujokat 1981; Shiotani et al. 1993). In the SH, such dramatic changes of the stratospheric jet do not occur due to the smaller planetary wave activities than the NH (exceptions include minor warming in 1988 and major warming in 2002; see Hio and Yoden 2005), and the upper-stratospheric westerly jet shifts gradually poleward and downward (hereafter “shift-down”, as in previous studies) from SH late winter to spring at the timing of final warming (the westerly jet in the winter hemisphere is replaced by the easterly jet in the summer hemisphere) (e.g., Harwood 1975; Hartmann 1976; Hirota et al. 1983; Shiotani and Hirota 1985). By using a 10-year short-term dataset from the Stratospheric Sounding Unit onboard the TIROS-N NOAA satellite, Shiotani et al. (1993) first found that the seasonal march of the stratospheric westerlies over the SH varies from year-to-year with planetary wave activity. They showed that the year-to-year variability of the intensity and position of the stratospheric polar jet in the SH midwinter influences the shift-down of the polar jet in the SH late winter and spring associated with the year-to-year variability of upward propagation of planetary waves from the troposphere into the stratosphere.

Several studies have examined possible linkages between the year-to-year variability of the SH polar vortex and long-term variations, such as the equatorial quasi-biennial oscillation (QBO) in tropical stratospheric zonal winds (e.g., Holton and Tan 1980; Baldwin and Dunkerton 1998), 11-year solar cycle (e.g., Kuroda and Kodera 2005), El Niño-Southern Oscillation (ENSO) (e.g., Hitchman and Rogal 2010), and the Antarctic ozone depletion that occurred during the last four decades of the 20th century (e.g., Thompson and Solomon 2002; Son et al. 2008; Akiyoshi et al. 2009; SPARC CCMVal 2010; McLandress et al. 2011; Thompson et al. 2011; World Meteorological Organization 2011; Grise et al. 2013).

Holton and Tan (1980) reported that the NH polar vortex is stronger during the westerly (hereafter QBO-W) phase of the 50 hPa equatorial QBO in NH winter (termed the Holton–Tan relationship). They also showed that the response of the SH polar vortex to the QBO in September–November (SH spring) indicated a stronger westerly of the SH polar vortex during the QBO-W as the NH polar vortex, although the signals are weak. In addition to the zonal wind response, for the SH spring, Garcia and Solomon (1987) suggested that the biennial variation of the October minimum ozone in the polar region from 1979 to 1986 might have been due to the influence of the equatorial QBO defined at 50 hPa. Subsequent works noted that the response of the SH polar vortex to the equatorial QBO (i.e., a stronger westerly during the QBO-W) was best seen in November using the equatorial QBO defined at 20–30 hPa (Baldwin and Dunkerton 1998; Naito 2002; Anstey and Shepherd 2014), and the SH QBO response was significantly observed in September–November using the equatorial QBO defined at 10–20 hPa (Hitchman and Huesmann 2009). Thus, several studies have discussed that the stratospheric QBO is closely associated with the SH polar vortex. However, the mechanism by which the QBO modulates extratropical circulation over the SH is poorly understood, as the various studies employed different height levels and metrics to define the phase of the QBO.

On the other hand, there are some indications of a linkage between the QBO and wave activities over the SH (e.g., Newman and Randel 1988; Naito 2002). Newman and Randel (1988) indicated that the biennial variation of the polar vortex was consistent with variation of planetary wave forcing over the SH. Naito (2002) further analyzed the QBO influences on the Eliassen and Palm (E–P) flux in the SH and showed that the QBO at the 20–30 hPa level leads to the E–P flux signal for planetary scale waves with zonal wavenumbers 1–3 in the stratosphere. However,
they employed composite analysis to determine the atmospheric fundamental characteristics based on the meteorological data, which possibly includes other signals such as solar cycle and ENSO other than response to the stratospheric QBO. Therefore, we examine the dynamic process in the SH, using multiple linear regression to separate the influences of the QBO and other factors on the SH polar vortex. In addition, the middle- and lower-stratospheric QBO are defined by two orthogonal QBO components obtained from empirical orthogonal function (EOF) analysis, and the contributions of both the middle- and lower-stratospheric QBO to the atmospheric field over the SH are evaluated. We focus on the seasonal evolutions of the SH polar vortex variability associated with the QBO and the other factors, and elucidate features of the beginning and development of the QBO-SH polar vortex relationship. During the SH winter to spring transition, the interaction between stratosphere and troposphere can occur through planetary waves under propagation conditions of westerly wind region in the SH stratosphere (see Section 3.1).

The paper is organized as follows. The reanalysis data and the analysis method are described in Section 2. The results are presented in Section 3. A summary is provided in Section 4.

2. Data and analysis method

2.1 Data

The primary data used in this study are the Japanese 55-year Reanalysis (JRA-55) dataset (Kobayashi et al. 2015). The JRA-55 reanalysis system is based on the TL319 horizontal resolution version of the Japan Meteorological Agency (JMA) operational data assimilation system from 1000 to 0.1 hPa with 60 vertical levels. We used the 6-hourly samples of reanalysis products from 1960 to 2010 with 1.25° × 1.25° longitude–latitude grids of 37 vertical levels, from 1000 to 1 hPa, for wind and temperature data.

2.2 Analysis of wave propagation and circulation

To clarify the dynamical processes of the extratropical circulation over the SH associated with the stratospheric QBO, we used the E–P flux and the residual mean meridional circulation derived from the primitive equation on the sphere (e.g., Andrews et al. 1987) with the daily mean of the JRA-55 zonal wind, meridional wind, vertical wind, and temperature data. The divergence of the E–P flux is connected with the eastward acceleration by wave activities. These variables have been used in many studies to investigate connections between the QBO and extratropical circulation (e.g., Holton and Tan 1982; Niwano and Takahashi 1998; Chen and Li 2007; Yamashita et al. 2011; Inoue et al. 2011; Garfinkel et al. 2012; Lu et al. 2013).

2.3 Multiple linear regression analysis

Multiple linear regression analysis was used to examine which factors are important for the variability of atmospheric fields over the SH. This was applied to the 51-year (1960–2010) monthly-mean meteorological variables, such as zonal mean wind and E–P flux, to analyze seasonal evolution in the stratosphere and troposphere, focusing on year-to-year variability of the SH polar vortex.

Based on Crooks and Gray (2005), Austin et al. (2008), Akiyoshi et al. (2009), Yamashita et al. (2010), and Mitchell et al. (2015), we perform multiple regression analysis, including the constant, equivalent effective stratospheric chlorine (EESC), solar cycle phase, two QBO phases, ENSO, and volcanic aerosol terms.

The following regression equation was used:

\[ Y(t) = c_0 + \alpha \text{EESC}(t) + \beta_a \text{QBOa}(t) + \beta_b \text{QBOb}(t) + \gamma \text{Solar}(t) + \delta \text{ENSO}(t) + \epsilon \text{Vol}(t) + R(t), \]

where \( Y(t) \) is the target variable for regression at time \( t \) [month]. The long-term annual-mean constant and seasonally varying intercept are included in the regression coefficient \( c_0 \) as

\[ c_0 = A_0 + \sum_{j=1}^{J} [A_j \cos(2\pi j t / 12) + B_j \sin(2\pi j t / 12)], \]

where \( J = 4 \) and \( A_j \) and \( B_j \) are Fourier coefficients.

The time series of the explanatory variables for the regression analysis that we used here are EESC term (EESC), two QBO components (QBOa and QBOb, which are defined below), solar cycle (Solar), ENSO, and volcanic aerosol (Vol). The seasonally varying coefficients describing the influence of the explanatory variables are denoted as \( \alpha, \beta_a, \beta_b, \gamma, \delta, \) and \( \epsilon. \) \( R(t) \) indicates the residual variation that is not described by the regression model. An error estimation for the regression coefficient was described by Randel and Cobb (1994). The 1\( \sigma \) and 2\( \sigma \) statistical significances are indicated where the statistical fits are significantly different from zero at the 1\( \sigma \) and 2\( \sigma \) levels. We mainly discuss the regression coefficients for two QBO phases, and their results are mainly presented in Sections 3.3 and 3.4.

Defining two orthogonal QBO phases, we use the
first two principal components (PCs) of monthly mean zonal wind data for the 10–70 hPa levels of the equatorial stratosphere in an EOF analysis, as performed by Crooks and Gray (2005), Taguchi (2010), and Mitchell et al. (2015). We then discuss the response of the atmosphere in the SH to the QBO as defined by the regression terms corresponding to the two QBO indices. The indices of $QBO_a(t)$ and $QBO_b(t)$ are the first two PCs derived from the monthly mean equatorial zonal wind data at the 10, 12, 15, 20, 25, 30, 35, 40, 45, 50, 60, and 70 hPa levels. The zonal wind data are based on rawinsonde observations at Canton Island (2.46°S, 171.43°W), Gan, Maldives (0.41°S, 73.09°E), and Singapore (1.22°N, 103.55°E) provided by the International Global Atmospheric Chemistry (IGAC) / the Stratosphere–Troposphere Processes and their Role in Climate (SPARC) Chemistry–Climate Model Initiative (CCMI) (Morgenstern et al. 2017). To derive the PCs, the time series of zonal wind data was deseasonalized and then smoothed with a 5-month running mean to remove month-to-month variation (for details see Wallace et al. 1993, Taguchi 2010); then, the covariance matrix of zonal wind was solved to determine the EOFs and corresponding PCs. The sum of variances of the first (55 %) and second (41 %) modes account for 96 % of the total variance. We confirmed that the temporal behaviors of the first and second modes of the zonal wind can best describe the evolutions of the equatorial zonal wind at 10–20 hPa and 30–50 hPa, respectively. Thus the regression coefficient of $QBO_a(t)$ indicates the variations synchronized with the middle-stratospheric QBO, while that of $QBO_b(t)$ describes the variations synchronized with the lower-stratospheric QBO.

Since this study focuses on the SH polar vortex in SH spring, the EESC term is included to account for possible effects of ozone depletion over this period (e.g., Newman et al. 2007). The index of $EESC(t)$ was derived from the EESC index displayed in Fig. 3d of Newman et al. (2007) (hereafter N07EESC) as

$$EESC(t) = \left[ N07EESC(t) - AVR(N07EESC) \right] / SD(N07EESC),$$

where $AVR(N07EESC)$ is the mean value of the N07EESC and $SD(N07EESC)$ is its standard deviation.

Since the atmospheric field over the SH is affected by various factors including the solar cycle and ENSO, terms other than the QBO are included in the regression equation. The index of the solar cycle $Solar(t)$ was derived from Ottawa/Penticton 10.7 cm solar radio flux (F10.7) data provided by the NOAA/National Geophysical Data Center as

$$Solar(t) = [F10.7(t) - AVR(F10.7)] / SD(F10.7),$$

where $AVR(F10.7)$ is the mean value of the F10.7 index and $SD(F10.7)$ is its standard deviation. The time series of F10.7 data were smoothed in advance with a 5-month running mean. The index for the ENSO was derived from the 5-month running mean data of the Nino 3.4 index (NINO3.4) provided by the NOAA/Earth System Research Laboratory (http://www.esrl.noaa.gov/psd/data/correlation/nina34.data) and was calculated with $ENSO(t) = [NINO3.4(t) - AVR(NINO3.4)] / SD(NINO3.4)$. The index of volcanic activity was derived from the 5-month running mean for optical thickness $\tau$ at 550 nm integrated from the 15 to 35 km levels, provided by the IGAC/SPARC CCMI as $Vol(t) = [\tau(t) - AVR(\tau)] / SD(\tau)$.

3. Results

3.1 Seasonal behaviors of the SH polar-night jet and its interannual variation

Figure 1a shows the seasonal change of climatological zonal-mean zonal wind at 60°S and 10 hPa. The shading represents one standard deviation of annual zonal wind during the whole period. The zonal wind reaches its maximum in August, and the interannual variability is high from August to November. The results are in agreement with those of Hio and Yoden (2005). Here, we focus on temporal behaviors of the location of the polar-night jet (PNJ) in the climatological field. Figure 1b shows the latitude–height section of the seasonal evolution of the PNJ axis position (circles) from 1 August to 31 December (the first day of each month is shown as a cross), where the PNJ axis is defined as the zonal wind maximum in the latitude–height plane over 20–90°S, 1–100 hPa. Note that almost all levels of the PNJ axis in April–July are located at 1 hPa and higher (in this case, the PNJ axis position is determined at 1 hPa), where the vertical resolution of the JRA-55 data is coarse; thus, the PNJ axis position in April–July is not shown. In August (late winter), the PNJ axis is located at around 60°S, 2 hPa, leading to southward and downward movement of the PNJ axis until spring (October–November) (e.g., Kodera and Kuroda 2002), together with gradually decreasing zonal wind intensity. The timing of the shift-down is consistent with the decrease in zonal-mean zonal wind at 10 hPa from September to December (Fig. 1a). Therefore, we use the zonal-mean zonal wind at 60°S, 10 hPa as the indicator of seasonal...
behavior of the PNJ.

3.2 Seasonal evolution of the multiple regression terms

We estimated respective terms of the multiple linear regression equation for the zonal-mean zonal wind. Table 1 shows the monthly regression coefficients for EESC (EESC; $\alpha$), two QBO components (QBOa and QBOb; $\beta_a$ and $\beta_b$), solar cycle (Solar; $\gamma$), El Niño-Southern Oscillation (ENSO; $\delta$), and volcanic aerosol (Vol; $\varepsilon$) averaged over 40–80°S, 1–200 hPa. From spring to early summer, $\beta_a$, variations synchronized with the middle-stratospheric QBO, are dominant, with a maximum in November. For instance, $\beta_a$ in September is $0.42 \pm 0.22$, which is more than double the other coefficients, $\beta_b$, $\gamma$, and $\delta$. Coefficients $\varepsilon$ and $\beta_b$ are more dominant than $\beta_a$ in late winter (August) and winter (June–August), respectively. Coefficient $\alpha$

![JRA-55 60°S U10](image)

![JRA-55 PNJ](image)

**Fig. 1.** (a) Climatological seasonal evolution of 10 hPa daily zonal-mean zonal wind at 60°S for the period 1960–2010 (solid line), showing one standard deviation of interannual variability (shaded); (b) Latitude–height section of climatological seasonal evolution of polar-night jet (PNJ) axis position (circles) from 1 August to 31 December (the first day of each month is shown as a cross), together with its maximum wind speed (colored).

**Table 1.** Monthly values and 1σ errors of regression coefficients in the linear regression equation for zonal-mean zonal wind averaged over 40–80°S, 1–200 hPa. Units for $\alpha$, $\beta_a$, $\beta_b$, $\gamma$, $\delta$, and $\varepsilon$ are m s$^{-1}$ per one standard deviation of the index terms. The bold indicates that the statistical fits are significantly different from zero at the 1σ level.

|       | Jun.       | Jul.       | Aug.       | Sep.       | Oct.       | Nov.       | Dec.       |
|-------|------------|------------|------------|------------|------------|------------|------------|
| $\alpha$ | 0.14 ± 0.16 | 0.15 ± 0.19 | −0.10 ± 0.21 | −0.33 ± 0.21 | 0.21 ± 0.23 | 1.25 ± 0.23 | 1.60 ± 0.18 |
| $\beta_a$ | −0.11 ± 0.17 | 0.12 ± 0.20 | 0.17 ± 0.22 | 0.42 ± 0.22 | 0.89 ± 0.25 | 1.01 ± 0.25 | 0.56 ± 0.19 |
| $\beta_b$ | −0.30 ± 0.17 | −0.26 ± 0.18 | −0.21 ± 0.19 | −0.17 ± 0.18 | 0.12 ± 0.20 | 0.56 ± 0.22 | 0.63 ± 0.18 |
| $\gamma$ | 0.15 ± 0.17 | 0.20 ± 0.20 | 0.06 ± 0.21 | −0.10 ± 0.20 | −0.18 ± 0.22 | −0.20 ± 0.22 | −0.13 ± 0.17 |
| $\delta$ | −0.33 ± 0.25 | −0.24 ± 0.26 | 0.04 ± 0.25 | 0.19 ± 0.21 | 0.02 ± 0.21 | −0.25 ± 0.21 | −0.32 ± 0.16 |
| $\varepsilon$ | −0.02 ± 0.17 | 0.06 ± 0.20 | 0.23 ± 0.21 | 0.22 ± 0.20 | −0.05 ± 0.22 | −0.35 ± 0.23 | −0.40 ± 0.18 |
is more dominant during November–December, \( \delta \) is more dominant during June–July, and \( \epsilon \) is more dominant in August. These results indicate that the impacts of the EESC, ENSO, and volcanic activities on the atmospheric fields in the SH \((i.e., \alpha, \delta, \text{and} \epsilon)\) are not negligible during the analysis period. However, here we focus on the spatial distributions of \( \beta_a \) and \( \beta_b \) to clarify the dynamical contributions of the middle- and lower-stratospheric QBO to the SH circulation, because the contribution of QBO to the SH polar vortex is dominant from spring \((\text{September to November})\) to early summer \((\text{December})\).

Figure 2 shows the latitude–time sections of the multiple regression coefficients (black contours) of two QBO components \((QBOa \text{ and } QBOb)\), together with the 1\(\sigma\) and 2\(\sigma\) statistical significance levels (light and dark shading) where the 1\(\sigma\) and 2\(\sigma\) statistical errors exceed the absolute values of the regression coefficients. Hereafter, the regression coefficients \(\beta_a\) and \(\beta_b\) describing the influence of the explanatory variables of \(QBOa\) and \(QBOb\) are called “QBOa signal” and “QBOb signal,” respectively. It should be noted that similar signals are derived from different time intervals in the JRA-55 dataset of 1958–2017, 1960–1978, 1979–2010, and 1979–2017, and signals similar to the JRA-55 are derived from JRA-55C (an atmospheric reanalysis assimilating conventional observations only) and ERA-interim for the same time interval of 1979–2010 (not shown). Note that the similar signals are derived from the regression with La Niña and El Niño years separately, with slightly early change over mid- and high latitudes in El Niño years (not shown). The green contours in Fig. 2 indicate zonal-mean zonal wind at 10 hPa in the climatological component \((\text{coefficient } c_0)\). Except for late spring and summer \((\text{November–February})\), westerlies are dominant in the mid- and high latitudes at 10 hPa.

The position of the maximum westerly wind, which is representative of the PNJ axis, moves from 70\(^\circ\)S to 60\(^\circ\)S during April–August, with largest magnitude \((> 80 \text{ m s}^{-1})\) occurring in August. The maximum of the westerly wind moves to the polar region with gradual decrease until November, in agreement with the decrease in wind speed at 10 hPa, as shown in Fig. 1a.

We first discuss the QBOa westerly signal of zonal wind. The temporal variation of the QBOa coefficient shows that the westerly signal at 0–30\(^\circ\)S in midwinter \((\text{July})\) appears to extend poleward toward the Antarctic in late winter and spring \((\text{August–November})\). The westerly wind signal has a maximum of 3 m s\(^{-1}\) around 65\(^\circ\)S in November at a statistical significance level of 2\(\sigma\) (Fig. 2a). Figure 3a shows the latitude–height section of the QBOa coefficient in November, when a strong westerly wind is shown in mid- and high latitudes in Fig. 2a. The westerly maximum at

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**JRA-55 10 hPa Zonal Wind**

![Fig. 2. Latitude–time section of: (a) QBOa \((\beta_a)\) and (b) QBOb \((\beta_b)\) coefficients (black lines: m s\(^{-1}\) per one standard deviation of the index terms) of the regression equation at 10 hPa, together with the 1\(\sigma\) and 2\(\sigma\) statistical significances (light and dark shading). Green lines indicate climatological zonal-mean zonal wind at 10 hPa for the period 1960–2010. Contour interval is 10 m s\(^{-1}\) (solid lines: zero line and positive values; broken lines: negative values).**
10–20 hPa around the equator corresponds to positive values of $\beta_a$ near 10–20 hPa. This means that the QBOa signal found in mid- and high latitudes can be regarded as the influence of the equatorial middle-stratospheric QBO. $\beta_a$ is positive (i.e., westerly wind signal) throughout the extratropical stratosphere. In addition, the maximum westerly signal is located near the zero-wind line of the climatological zonal wind (65°S and 5–10 hPa). These findings imply that the westerly wind area in the extratropical stratosphere is extended to more upper levels in the QBO-W condition, and the PNJ over the SH remains strong for a longer time. In other words, the westerly phase of middle-stratospheric QBO favors a delay in the PNJ shift-down. The westerly signal extending from the SH mid-latitudes to the Antarctic at the late timing of the shift-down is similar to the results of Anstey and Shepherd (2014), who showed the seasonality in the response of the SH westerly wind to the QBO-W defined 20 hPa.

Figures 2b and 3b show regression coefficients of the QBOb ($\beta_b$, black contour) for zonal-mean zonal wind in the time–latitude and latitude–height sections, respectively. We find the QBOb signal of zonal wind at the equator, with a westerly maximum at 30–50 hPa (Fig. 3b). This means that the QBOb signal found in mid- and high latitudes is regarded as the response to the equatorial lower-stratospheric QBO. Similarly to the QBOa signal, the positive values of $\beta_b$ (i.e., the westerly wind signal) are found in the mid- and high-latitude stratosphere in SH spring (October–November), with statistical significance (Fig. 2b), suggesting that the shift-down of the PNJ is delayed in the QBO-W condition at 30–50 hPa. In contrast to the QBOa effects, the westerly signal that is found around 30–40°S in June–July disappears in August, and further southward movement of the westerly signal is not detected in late winter and spring (Fig. 2b).

3.3 Middle-stratospheric QBO westerly effects on the seasonal evolution of the PNJ in SH winter to spring
To investigate how the response of SH polar vortex to the middle-stratospheric QBO arises in winter and spring, the seasonal evolutions of the zonal wind, temperature, residual mean meridional circulation, and wave propagation using E–P flux were analyzed. Figures 4a–d show latitude–height sections for the QBOa signals of zonal wind from July to October; Figs. 4e–h show those of temperature (contour) with residual mean meridional circulation (vector); and Figs. 4i–l show those of the zonal wavenumber 1–3 (WN1–3) components of E–P flux (vector) with its divergence (contour).

In July, there is an easterly signal ($\beta_a < 0$) around 1 hPa that lies over a westerly signal ($\beta_a > 0$) around 10–20 hPa in the equatorial region (10°N–30°S) (Fig. 4a). Residual circulation analysis revealed a two-cell structure of the meridional circulation signal in the equatorial region with a southward flow signal at 2 hPa, an equatorward flow signal at 10–20 hPa, and a southward signal at 50–100 hPa (Fig. 4e). The upper...
JRA-55 QBOa coefs.

(a) Jul. U  (e) Jul. T, (v*, w*)  (i) Jul. WN1–3

(b) Aug. U  (f) Aug. T, (v*, w*)  (j) Aug. WN1–3

(c) Sep. U  (g) Sep. T, (v*, w*)  (k) Sep. WN1–3

(d) Oct. U  (h) Oct. T, (v*, w*)  (l) Oct. WN1–3

Fig. 4. (a–d) As in Fig. 3, but for QBOa coefficients of zonal wind in June, July, August, September, and October, together with climatology; (e–h) As in (a–d), but for coefficients of temperature (contour: K), residual mean meridional circulation (vector: m s$^{-1}$), together with the 1σ and 2σ statistical significances of temperature (shading). The units are per one standard deviation of the index terms. Residual circulation vectors below 100-hPa level are multiplied by 0.3; (i–l) As in (a–d), but for zonal wavenumber 1–3 (WN1–3) components of the E–P flux (vector: kg m$^{-1}$ s$^{-2}$) and its divergence (contour: m s$^{-1}$ d$^{-1}$). E–P flux vectors below 100-hPa level are multiplied by 0.1. The vertical components of E–P flux and residual circulation are magnified ×100 relative to the horizontal component; scales for the horizontal vector are shown at the bottom-right of the panel.
cell corresponds to the upward flow signal around the equator, and the downward flow signal at 20–50°S around 2–10 hPa levels (Fig. 4a) corresponds to the QBO’s easterly wind shear in vertical; the lower cell corresponds to the downward flow signal around the equator, and the upward flow signal at 20–50°S around 20–50 hPa levels (Fig. 4a) corresponds to the QBO’s westerly wind shear in vertical, suggesting that those responses are consistent with the expected QBO-induced mean meridional circulation (e.g., Plumb and Bell 1982). Note that the divergence signal of E–P flux is shown around 20–70°S and 2–20 hPa (Fig. 4i), which corresponds to the westerly acceleration by the planetary waves and is related to the equatorward flow signal of meridional circulation. Thus, the influences of the planetary waves are not ruled out. Consistent with the downward flow signal around 2–10 hPa and the upward flow signal around 20–50 hPa at 20–50°S, high-temperature signals are observed around 2–10 hPa and low-temperature signals around 20–50 hPa at the same latitude ranges. By the thermal wind relationship, these temperature signals should be related to the easterly wind signal around 1 hPa and the westerly wind signal around 10–20 hPa in the SH mid-latitudes (40–60°S) (Fig. 4a).

The previous paragraph described the westerly wind signal around 1–5 hPa, 40–60°S in July. We next describe the temporal behaviors of the westerly wind signal over the SH from SH midwinter to spring. The westerly wind signal around 1–5 hPa, 40–60°S in July is located on the equatorward side of the polar jet axis, meaning that the polar vortex prefers to shift equatorward under a QBO-W condition as compared to a QBO-E condition in July. In general, planetary waves propagate upward from the troposphere into the stratosphere during the SH winter and spring (e.g., Charney and Drazin 1961). These planetary waves are refracted away from the strong polar vortex in July. In other words, climatological E–P flux is upward from the extratropical troposphere into the stratosphere and upward/equatorward in the stratosphere in July. Thus, the equatorward shift of the polar vortex in the QBO-W condition corresponds to suppression of the climatological wave propagation route toward the equatorial upper stratosphere, with a southward/downward signal of the E–P flux around 40–60°S and 10–100 hPa (Fig. 4i), thereby indicating the suppression of climatological planetary wave (WN1–3) propagation in the QBO-W condition. This downward signal corresponds to the E–P flux divergence signal (westerly acceleration due to the suppression of, e.g., wave dissipation) in the mid- and high latitudes in the stratosphere around 20–70°S and 2–20 hPa in July (Fig. 4i). The area of westerly wind signal around 1–2 hPa, 40–60°S in July moves southward and downward from July to October (Figs. 4a–d), and the center of this signal is always located over the upward/equatorward area of the PNJ core. Note that the timing of the shift-down of the PNJ in the QBO-W condition is later than that in the QBO-E condition. This monthly shift of βz (Figs. 4a–d) is consistent with the monthly southward movement of the westerly signal (Fig. 2a). As shown in Figs. 4i–1 and 6a, the area of downward E–P flux around 40–80°S and 10–100 hPa itself moves downward from July to October due to the equatorward shift of the PNJ in QBO-W as the slowing of the shift-down of the PNJ (Figs. 4a–d). This downward movement of the area of downward E–P flux corresponds to that of the E–P flux divergence signal in the stratosphere averaged over 40–80°S (Fig. 6a). Note that the E–P flux for WN1–3 largely (more than 80 %) accounts for the net E–P flux derived from the total wavenumbers (not shown).

The divergence area is related to the upward circulation signal around 50–80°S and downward circulation signal around 20–40°S at 2–10 hPa in July and August (Figs. 4e, f). The upward circulation signal is widespread, from Antarctic (50–90°S) to the mid-latitudes (20–40°S) at 2–200 hPa in September–October (Figs. 4g, h), as the divergence area is widespread, from 90°S to 40°S in the upper stratosphere, with the seasonal evolution of the PNJ. Via adiabatic cooling, this upward circulation signal might be responsible for the low-temperature signal over the Antarctic region around 2–50 hPa in September, and around 50–90°S and 10–100 hPa in October (Figs. 4g, h).

3.4 Lower QBO westerly effects on the SH polar-night jet

To investigate the features of the seasonality and spatial distribution of the QBOb signal, we analyzed the latitude–height sections of the monthly mean QBOb signals. Figure 5 shows the latitude–height sections of regression coefficients βz of zonal-mean zonal wind, temperature, residual mean meridional circulation, and the zonal wavenumber 1–3 (WN1–3) components of E–P flux with its divergence. A westerly wind signal is seen at mid-latitudes (30–50°S) around 5 hPa in the upper stratosphere in July (Fig. 5a), which is located in the equatorward area of the PNJ core. In August, this westerly wind signal has become ill-defined in the upper stratosphere at 5–10
hPa around 40°S (Fig. 5b), accompanying the deepening of the easterly wind signal at 2–100 hPa around 60°S. These signals correspond to the convergence signal of the E–P flux around 5–10 hPa, 20–60°S (Fig. 5i). In September, two weak westerly wind signals appear separately again at 1 hPa near 60°S and at 10 hPa around 30°S (Fig. 5c). These two westerly signals merge into a single large, strong signal at the
extratropics in the upper stratosphere, centered on the upward/equatorward area of the PNJ core in October (Fig. 5d), suggesting that the shift-down of the PNJ in the QBO-W condition occurs later than that in the QBO-E condition in October.

We also examined meridional circulation. In July, we observed a southward signal at 2–10 hPa, an equatorward signal at 30–50 hPa, and a southward signal near 100 hPa in the tropics with downward signal at 20–50°S around 10–20 hPa (Fig. 5e). These QBO-induced meridional flows are observable from July to October, with the high-temperature signal associated with adiabatic heating (Figs. 5e–h). However, the downward signal of the meridional circulation at 20–50°S around 10–20 hPa is weakened in August, with temperature signal (maximum about 0.5 K) approximately half that in July (maximum about 1 K), or rather the meridional circulation is extended from 40°S to 80°S around 5–10 hPa in August, indicating the downward signal over the SH high latitudes with the high-temperature signal there (Fig. 5f). This suggests that the behavior of the QBO-induced meridional circulation (i.e., two-cell structure) alone cannot explain this weakening of the westerly wind signal in August. Upward signals of the E–P flux, in relation to enhanced wave propagation from troposphere into stratosphere, are seen around 40–80°S at 10–200 hPa with the convergence signal in the stratosphere in August (Figs. 5j, 6b). These E–P flux signals for WN1–3 largely account for the E–P flux signal for the total wavenumbers (not shown). The convergence signal of the E–P flux could induce the extended meridional circulation signal from the equator to 80°S around 5–10 hPa in August with the weakening of the QBOb westerly wind signal around 40°S at 5–10 hPa and the deepening of the easterly wind signal at 2–100 hPa around 60°S. With the increasing magnitude of the upper stratospheric westerly wind signal around 60°S from September to October (Figs. 5c, d), the wave activity around 60°S becomes refracted away from the strong westerly wind of the polar vortex, indicating the downward signal of the E–P flux at 10–100 hPa with the divergence signal (e.g., less dissipation of the planetary waves) in October (Fig. 5l).

Since wave propagation from troposphere into stratosphere in the QBO-W is enhanced during August and is related to the convergence signal of the E–P flux around the PNJ (Figs. 5i, 6a), the planetary waves influence the weakening of the westerly wind signal in the equatorward area of the PNJ core in August (Fig. 5a). The enhancement of planetary wave propagation from troposphere to stratosphere could be controlled by the strong stratospheric wind, as well as enhancement of the source of tropospheric wave activity. In future work, the behavior of tropospheric waves associated with the QBO will be investigated using both

**JRA-55 Fz & EPF Div.**

![Fig. 6. Time–height section of: (a) QBOa (βa), and (b) QBOb (βb) coefficients of the vertical E–P flux for the WN1–3 component averaged over 40–80°S (black lines: $\times 10^{-3}$ kg m$^{-1}$ s$^{-2}$ per one standard deviation of the index terms), together with the E–P flux divergence, shown as red (positive) and blue (negative) (shaded: m s$^{-1}$ d$^{-1}$).](image)
observations and numerical modeling, to identify the relative importance of the strong stratospheric wind and the tropospheric wave source.

4. Summary

The mechanism by which the QBO modulates extratropical circulation over the SH is poorly understood, as the various studies employed different height levels and metrics to define the phase of the QBO. In this study, we examined the response of atmospheric circulation over the SH to the stratospheric QBO westerly, defined by the equatorial zonal wind phases at middle- and lower-stratospheric levels. Previous studies employed composite analyses that possibly included signals other than the response to the stratospheric QBO. In this study, multiple linear regression analysis of the EESC, two orthogonal QBO components, ENSO, and the volcanic aerosol terms was conducted using the JRA-55 Reanalysis data set over a 51-year period from 1960 to 2010. Defining two orthogonal QBO components, the first two PCs (QBOa and QBOb signals) of the equatorial zonal wind were used for the QBO-related signals derived from an EOF analysis. These QBOa and QBOb signals correspond to the middle- and lower-stratosphere QBO westerly responses, respectively. From SH spring to early summer, the effects of the middle-stratospheric QBO are dominant, whereas the effects of ENSO are dominant in SH early- to mid-winter.

The seasonal evolution of the climatological SH polar vortex indicates poleward and downward movement of its central position (shift-down) from SH late winter (August) to spring (September–November), as shown in previous studies. We suggest possible pathways of the QBO response in the polar vortex evolution at these middle- and lower-stratospheric levels.

In association with the middle-stratospheric QBO at 10–20 hPa (QBOa), the shift-down of the polar vortex in the QBO-W condition occurs later than that in the QBO-E condition from SH late winter to spring. This is associated with the westerly wind signal over the equatorward/upward area of the PNJ core in the SH mid-latitudes in the upper stratosphere. This westerly wind signal is consistent with the high-temperature anomaly in the SH low latitudes, synchronized with the QBO-induced mean meridional circulation that occurs from July to October. The delay in the shift-down of the strong polar vortex corresponds to the suppression of planetary wave propagation from the troposphere into the stratosphere.

The shift-down of the polar vortex in the lower-stratospheric QBO-W condition at 30–50 hPa (QBOb) is also more delayed than that in QBO-E condition in SH spring, and the QBO-induced mean meridional circulation is also seen from July to October. In contrast to the QBOa, the westerly wind response of the SH polar vortex to the lower-stratospheric QBO is unclear in August. This corresponds to the enhanced planetary wave propagation from the troposphere to the stratosphere in August.

Note that the diagnostic study in this work cannot establish a firm conclusion, but hypotheses can be drawn about the roles of QBO influences. Hence, these results suggest that there are two possible pathways of stratospheric response to the QBO in terms of the seasonal evolution: one possible pathway is that the middle-stratospheric QBO modifies the distributions of air temperature related to zonal wind and planetary wave activities, and meridional circulation in the low latitudes. This favors delayed evolution of the PNJ at high latitudes (around 60°S) from late winter (August) to spring (September–November) during the westerly phase of the middle-stratospheric QBO, consequently tending to strengthen westerly winds from stratosphere to troposphere in spring. Lower-stratospheric QBO is also related to temperature in low latitudes and zonal wind in low- and mid-latitudes in the stratosphere. The other possible pathway involves an increase in upward propagation of planetary waves from the mid- and high latitudes troposphere to stratosphere in August responding to the westerly phase of the lower-stratospheric QBO, which is consistent with weakening of the PNJ. The numerical models that include the internal QBO and the relationship between the QBO and the polar vortex will be useful to clarify the two possible pathways of stratospheric response to the QBO.

Acknowledgments

The authors thank anonymous reviewers for their helpful comments and Dr. H. Akiyoshi of the National Institute for Environmental Studies, Japan, for many helpful discussions and comments. This work was supported by the Japan Society for the Promotion of Science (JSPS) KAKENHI Grant Number JP16K16186 and JP18K03748. The datasets used for this study were provided from the Japanese 55-year Reanalysis (JRA-55) project conducted by the Japan Meteorological Agency (JMA), and the data set can be accessed via the JMA website (http://jra.kishou.go.jp/). Generic Mapping Tools (GMT) and Grid Analysis and Display System (GrADS) were used to draw the figures.
References

Akiyoshi, H., L. B. Zhou, Y. Yamashita, K. Sakamoto, M. Yoshiki, T. Nagashima, M. Takahashi, J. Kurokawa, M. Takigawa, and T. Imamura, 2009: A CCM simulation of the breakup of the Antarctic polar vortex in the years 1980–2004 under the CCMVal scenarios. *J. Geophys. Res.*, 114, D03103, doi:10.1029/2007JD009261.

Andrews, D. G., J. R. Holton, and C. B. Leovy, 1987: *Middle Atmosphere Dynamics*. International Geophysics Series, Vol. 40, Academic Press, 489 pp.

Anstey, J. A., and T. G. Shepherd, 2014: High-latitude influence of the quasi-biennial oscillation. *Quart. J. Roy. Meteor. Soc.*, 140, 1–21.

Austin, J., K. Tourpali, E. Rozanov, H. Akiyoshi, S. Bekki, G. Bodeker, C. Brühl, N. Butchart, M. Chipperfield, M. Deushi, V. I. Fomichev, M. A. Giorgetta, L. Gray, K. Kodera, F. Lott, E. Manzini, D. Marsh, K. Matthes, T. Nagashima, K. Shibata, R. S. Stolarski, H. Struthers, and W. Tian, 2008: Coupled chemistry climate model simulations of the solar cycle in ozone and temperature. *J. Geophys. Res.*, 113, D11306, doi:10.1029/2007JD009391.

Baldwin, M. P., and T. J. Dunkerton, 1998: Quasi-biennial modulation of the southern hemisphere stratospheric polar vortex. *Geophys. Res. Lett.*, 25, 3343–3346.

Charney, J. G., and P. G. Drazin, 1961: Propagation of planetary-scale disturbances from the lower into the upper atmosphere. *J. Geophys. Res.*, 66, 83–109.

Chen, W., and T. Li, 2007: Modulation of northern hemisphere wintertime stationary planetary wave activity: East Asian climate relationships by the Quasi-Biennial Oscillation. *J. Geophys. Res.*, 112, D20120, doi:10.1029/2007JD008611.

Crooks, S. A., and L. J. Gray, 2005: Characterization of the 11-year solar signal using a multiple regression analysis of the ERA-40 dataset. *J. Climate*, 18, 996–1015.

Garcia, R. R., and S. Solomon, 1987: A possible relationship between interannual variability in Antarctic ozone and the quasi-biennial oscillation. *Geophys. Res. Lett.*, 14, 848–851.

Garfinkel, C. I., T. A. Shaw, D. L. Hartmann, and D. W. Waugh, 2012: Does the Holton-Tan mechanism explain how the quasi-biennial oscillation modulates the Arctic polar vortex? *J. Atmos. Sci.*, 69, 1713–1733.

Grise, K. M., L. M. Polvani, G. Tselioudis, Y. Wu, and M. D. Zelinka, 2013: The ozone hole indirect effect: Cloud-radiative anomalies accompanying the poleward shift of the eddy-driven jet in the Southern Hemisphere. *Geophys. Res. Lett.*, 40, 3688–3692.

Hartmann, D. L., 1976: The structure of the stratosphere in the Southern Hemisphere during late winter 1973 as observed by satellite. *J. Atmos. Sci.*, 33, 1141–1154.

Harwood, R. S., 1975: The temperature structure of the Southern Hemisphere stratosphere August–October 1971. *Quart. J. Roy. Meteor. Soc.*, 101, 75–91.

Hio, Y., and S. Yoden, 2005: Interannual variations of the seasonal March in the Southern Hemisphere stratosphere for 1979–2002 and characterization of the unprecedented year 2002. *J. Atmos. Sci.*, 62, 567–580.

Hirota, I., T. Hirooka, and M. Shiotani, 1983: Upper stratospheric circulations in the two hemispheres observed by satellites. *Quart. J. Roy. Meteor. Soc.*, 109, 443–454.

Hitchman, M. H., and A. S. Huesmann, 2009: Seasonal influence of the quasi-biennial oscillation on stratospheric jets and Rossby wave breaking. *J. Atmos. Sci.*, 66, 935–946.

Hitchman, M. H., and M. J. Rogal, 2010: ENSO influences on Southern Hemisphere column ozone during the winter to spring transition. *J. Geophys. Res.*, 115, D20104, doi:10.1029/2009JD012844.

Holton, J. R., and H.-C. Tan, 1980: The influence of the equatorial quasi-biennial oscillation on the global circulation at 50 mb. *J. Atmos. Sci.*, 37, 2200–2208.

Holton, J. R., and H.-C. Tan, 1982: The quasi-biennial oscillation in the Northern Hemisphere lower stratosphere. *J. Meteor. Soc. Japan*, 60, 140–148.

Inoue, M., M. Takahashi, and H. Naoe, 2011: Relationship between the stratospheric quasi-biennial oscillation and tropospheric circulation in northern autumn. *J. Geophys. Res.*, 116, D24115, doi:10.1029/2011JD016040.

Kobayashi, S., Y. Ota, Y. Harada, A. Ebita, M. M oriya, H. Onoda, K. Onogi, H. Kamahori, C. Kobayashi, H. Endo, K. Miyaoka, and K. Takahashi, 2015: The JRA-55 Reanalysis: General specifications and basic characteristics. *J. Meteor. Soc. Japan*, 93, 5–48.

Kodera, K., and Y. Kuroda, 2002: Dynamical response to the solar cycle. *J. Geophys. Res.*, 107, 4749, doi:10.1029/2002JD002224.

Kuroda, Y., and K. Kodera, 2005: Solar cycle modulation of the Southern Annular Mode. *Geophys. Res. Lett.*, 32, L13802, doi:10.1029/2005GL022516.

Labitzke, K., 1977: Interannual variability of the winter stratosphere in the Northern Hemisphere. *Mon. Wea. Rev.*, 105, 762–770.

Labitzke, K., 1982: On the interannual variability of the middle stratosphere during the northern winters. *J. Meteor. Soc. Japan*, 60, 124–139.

Lu, H., T. J. Bracegirdle, T. Phillips, A. Bushell, and L. Gray, 2013: Mechanisms for the Holton-Tan relationship and its decadal variation. *J. Geophys. Res.*, 119, 2811–2830.

McLandress, C., T. G. Shepherd, J. F. Scinocca, D. A. Plummer, M. Sigmund, A. I. Jonsson, and M. C. Reader, 2011: Separating the dynamical effects of climate change and ozone depletion. Part II: Southern Hemisphere troposphere. *J. Climate*, 24, 1850–1868.

Mitchell, D. M., L. J. Gray, M. Fujiwara, T. Hibino, J. A. Anstey, W. Ebisuzaki, Y. Harada, C. Long, S. Misios,
P. A. Stott, and D. Tan, 2015: Signatures of naturally induced variability in the atmosphere using multiple reanalysis datasets. *Quart. J. Roy. Meteor. Soc.*, 141, 2011–2031.

Morgenstern, O., M. I. Hegglin, E. Rozanov, F. M. O’Connor, N. L. Abraham, H. Akiyoshi, A. T. Archibald, S. Bekki, N. Butchart, M. P. Chipperfield, M. Deushi, S. S. Dhomse, R. R. Garcia, S. C. Hardiman, L. W. Horowitz, P. Jöckel, B. Josse, D. Kinnison, M. Lin, E. Mancini, M. E. Manyin, M. Marchand, V. Marécal, M. Michou, L. D. Oman, G. Pitari, D. A. Plummer, L. E. Revell, D. Saint-Martin, R. Schofield, A. Stenke, K. Stone, K. Sudo, T. Y. Tanaka, S. Tilmes, Y. Yamashita, K. Yoshida, and G. Zeng, 2017: Review of the global models used within phase 1 of the Chemistry-Climate Model Initiative (CCMI). *Geosci. Model Dev.*, 10, 639–671.

Naito, Y., 2002: Planetary wave diagnostics on the QBO effects on the deceleration of the polar-night jet in the Southern Hemisphere. *J. Meteor. Soc. Japan*, 80, 985–995.

Naujokat, B., 1981: Long-term variations in the stratosphere of the Northern Hemisphere during the last two sun-spot cycles. *J. Geophys. Res.*, 86, 9811–9816.

Newman, P. A., and W. J. Randel, 1988: Coherent ozone-dynamical changes during the Southern Hemisphere spring, 1979–1986. *J. Geophys. Res.*, 93, 12585–12606.

Newman, P. A., J. S. Daniel, D. W. Waugh, and E. R. Nash, 2007: A new formulation of equivalent effective stratospheric chlorine (EESC). *Atmos. Chem. Phys.*, 7, 4537–4552.

Niwano, M., and M. Takahashi, 1998: The influence of the equatorial QBO on the Northern Hemisphere winter circulation of a GCM. *J. Meteor. Soc. Japan*, 76, 453–461.

Plumb, R. A., and R. C. Bell, 1982: A model of the quasi-biennial oscillation on an equatorial beta-plane. *Quart. J. Roy. Meteor. Soc.*, 108, 335–352.

Randel, W. J., and J. B. Cobb, 1994: Coherent variations of monthly mean total ozone and lower stratospheric temperature. *J. Geophys. Res.*, 99, 5433–5447.

Shiotani, M., and I. Hirota, 1985: Planetary wave-mean flow interaction in the stratosphere: A comparison between Northern and Southern Hemispheres. *Quart. J. Roy. Meteor. Soc.*, 111, 309–334.

Shiotani, M., N. Shimoda, and I. Hirota, 1993: Interannual variability of the stratospheric circulation in the Southern Hemisphere. *Quart. J. Roy. Meteor. Soc.*, 119, 531–546.

Son, S.-W., L. M. Polvani, D. W. Waugh, H. Akiyoshi, R. Garcia, D. Kinnison, S. Pawson, E. Rozanov, T. G. Shepherd, and K. Shibata, 2008: The impact of stratospheric ozone recovery on the Southern Hemisphere westerly jet. *Science*, 320, 1486–1489.

SPARC CCMVal, 2010: SPARC report on the evaluation of chemistry-climate models. SPARC Rep. 5, WCRP-132, WMO/TD-1526, Eyring, V., T. Shepherd, and D. Waugh (eds.), SPARC, World Climate Research Programme, 434 pp.

Taguchi, M., 2010: Observed connection of the stratospheric quasi-biennial oscillation with El Niño–Southern Oscillation in radiosonde data. *J. Geophys. Res.*, 115, D18120, doi:10.1029/2010JD014325.

Thompson, D. W. J., and S. Solomon, 2002: Interpretation of recent Southern Hemisphere climate change. *Science*, 296, 895–899.

Thompson, D. W. J., S. Solomon, P. J. Kushner, M. H. England, K. M. Grise, and D. J. Karoly, 2011: Signatures of the Antarctic ozone hole in Southern Hemisphere surface climate change. *Nat. Geosci.*, 4, doi:10.1038/ngeo1296.

Wallace, J. M., R. L. Panetta, and J. Estberg, 1993: Representation of the equatorial stratospheric quasi-biennial oscillation in EOF phase space. *J. Atmos. Sci.*, 50, 1751–1762.

World Meteorological Organization, 2011: *Scientific Assessment of Ozone Depletion: 2010*. Global Ozone Research and Monitoring Project–Report No. 52, WMO, Switzerland, 516 pp.

Yamashita, Y., K. Sakamoto, H. Akiyoshi, M. Takahashi, T. Nagashima, and L. B. Zhou, 2010: Ozone and temperature response of a chemistry climate model to the solar cycle and sea surface temperature. *J. Geophys. Res.*, 115, D00M05, doi:10.1029/2009JD013436.

Yamashita, Y., H. Akiyoshi, and M. Takahashi, 2011: Dynamical response in the Northern Hemisphere midlatitude and high-latitude winter to the QBO simulated by CCSR/NIES CCM. *J. Geophys. Res.*, 116, D06118, doi:10.1029/2010JD015016.