Future Changes in the Ozone Quasi-Biennial Oscillation with Increasing GHGs and Ozone Recovery in CCMI Simulations

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ABSTRACT

The future quasi-biennial oscillation (QBO) in ozone in the equatorial stratosphere is examined by analyzing transient climate simulations due to increasing greenhouse gases (GHGs) and decreasing ozone-depleting substances under the auspices of the Chemistry–Climate Model Initiative. The future (1960–2100) and historical (1979–2010) simulations are conducted with the Meteorological Research Institute Earth System Model. Three climate periods, 1960–85 (past), 1990–2020 (present), and 2040–70 (future) are selected, corresponding to the periods before, during, and after ozone depletion. The future ozone QBO is characterized by increases in amplitude by 15%–30% at 5–10 hPa and decreases by 20%–30% at 40 hPa, compared with the past and present climates; the future and present ozone QBOs increase in amplitude by up to 60% at 70 hPa, compared with the past climate. The increased amplitude at 5–10 hPa suggests that the temperature-dependent photochemistry plays an important role in the enhanced future ozone QBO. The weakening of vertical shear in the zonal wind QBO is responsible for the decreased amplitude at 40 hPa in the future ozone QBO. An interesting finding is that the weakened zonal wind QBO in the lowermost tropical stratosphere is accompanied by amplified QBOs in ozone, vertical velocity, and temperature. Further study is needed to elucidate the causality of amplification about the ozone and temperature QBOs under climate change in conditions of zonal wind QBO weakening.

1. Introduction

Increases in greenhouse gases (GHGs), especially carbon dioxide (CO₂), have important implications for the future evolution of ozone because increased GHGs are expected to lead to decreasing temperature in the stratosphere, caused by strong CO₂ longwave emissions into space (e.g., Stolarski et al. 2012). This stratospheric cooling indirectly increases ozone because it reduces the ozone loss rate in the upper stratosphere, owing to the strong positive temperature dependence of the Chapman reactions and the NOₓ cycle on the ozone loss rate (World Meteorological Organization 2014; Meul et al. 2014). Also, future ozone changes are expected to be affected by an increase in mean meridional mass circulation (i.e., the Brewer–Dobson circulation) in response to climate change (e.g., Butchart et al. 2006; McLandress and Shepherd 2009). Diallo et al. (2012) indicated that the shallow and deep branches of the Brewer–Dobson circulation were modulated by variability in tropical upwelling and also that the shallow branch is modulated by the subtropical transport barriers. The projected future evolution of total column ozone in the tropics was particularly sensitive to changes in tropical
upwelling (World Meteorological Organization 2014). The large spread among ozone projections obtained from diagnostics of chemistry–climate model (CCM) simulations indicate that the confidence level remains low (e.g., Strahan et al. 2011).

The quasi-biennial oscillation (QBO), which dominates the variability of the equatorial stratosphere, is characterized by alternating zonal wind direction between westerlies and easterlies, which descend with a mean period of 28 months. The QBO was discovered by Ebdon (1960) and Reed et al. (1961), and the QBO in total column ozone over the sub tropics was soon reported in ground-based observations (e.g., Funk and Garnham 1962; Ramanathan 1963; Angell and Korshover 1964). Analyses of satellite observations by the Total Ozone Mapping Spectrometer (TOMS) revealed the influences of QBO on the total column ozone in a global scale (Bowman 1989; Lait et al. 1989). The ozone QBO in the equatorial region has a vertical structure with significant ozone anomalies in the lower stratosphere (20–30 hPa) and the middle stratosphere (5–10 hPa), along with a node near 15 hPa, analyzed by using ozone profiles from satellite observations of the Stratospheric Aerosol and Gas Experiment (SAGE) II (Chipperfield et al. 1994; Hasebe 1994; Randel and Wu 1996), the Halogen Occultation Experiment (HALOE) (Dunkerton 2001), and the Microwave Limb Sounder (MLS) aboard the Upper Atmosphere Research Satellite (Fadnavis and Beig 2008). A clear QBO signature in ozone at 15–55 km was obtained by lidar (Leblanc and McDermid 2001). In addition, the vertical fine structure of ozone QBO variability was obtained from measurements of ozonesondes (Logan et al. 2003) and from the Southern Hemisphere Additional Ozonesondes (SHADOZ) network (Witte et al. 2008; Lee et al. 2010; Thompson et al. 2011).

In a modeling study Dunkerton (1985) suggested that vertical advection of ozone associated with the QBO produces local heating at the equator, which leads to changes in zonal mean meridional circulation. Diabatic heating effects of the ozone QBO on temperature and phase relationships between the zonal wind and ozone QBOs were examined by using two-dimensional models (Hasebe 1994; Li et al. 1995; Huang 1996; Cordero and Nathan 2000; Tantii and Hasebe 2002) and three-dimensional models (Butchart et al. 2003; Shibata and Deushi 2005). Hasebe (1994) compared observational evidence from SAGE data with experimental results from a mechanistic model, and the observational evidence was that the phase of the ozone QBO at 15–40 hPa precedes that of the zonal wind by several months. In his experiments, if ozone is solely controlled by dynamics, the QBOs in ozone and zonal wind are essentially in phase and they lag with the QBO in temperature by a quarter cycle. However, when a diabatic heating anomaly of the ozone QBO, which causes some upwelling and offsets the downwelling that gave the ozone increase, becomes important, the phase of the ozone QBO shifts up to a quarter cycle ahead and approaches that of the temperature QBO. He hypothesized that because of this in-phase relationship between the ozone and temperature QBOs, the ozone heating works to reduce the magnitude of the induced meridional circulation associated with the dynamical QBO.

Contrary to Hasebe’s hypothesis, Huang (1996) insisted that the radiative (solar heating) feedback of the ozone QBO on the QBO-induced secondary circulation is negligible. Cordero and Nathan (2000) indicated that the ozone QBO feedbacks alone decrease the magnitude of the temperature QBO by up to 15% by examining effects of wave–ozone and zonal mean–ozone feedbacks. Butchart et al. (2003) found that a diabatic feedback of the simulated ozone QBO in the lower stratosphere gives a 10% increase in the mean period of the oscillation and a 35% increase in the amplitude of the temperature oscillation by using the Met Office Unified Model. Shibata and Deushi (2005) concluded that the ozone feedback produces a net decrease of radiative heating by one-third to one-half owing to a partial cancellation between solar heating and terrestrial heating and that this partial cancellation results in the phase relationship in which ozone QBO precedes zonal wind by about a quarter cycle below the middle stratosphere.

Climate change has a potential impact on the tropical circulation and a future recovery of ozone in the equatorial stratosphere. Kawatani and Hamilton (2013) indicated that a weakening trend of QBO amplitude in the lower stratosphere is consistent with a long-term trend of enhanced upwelling near the tropical tropopause, by analyzing radiosonde observations. Future QBO behaviors under climate change were investigated by Shibata and Deushi (2012) using transient simulations, and they found that the QBO amplitude trend is caused by the parameterized gravity wave forcing and vertical advection of zonal momentum. In this QBO amplitude trend, the vertical advection of zonal momentum is brought not only by the mean tropical upwelling but also by the QBO secondary circulation in the middle and upper stratosphere, while the vertical advection of zonal momentum is mainly brought by the mean tropical upwelling in the lower stratosphere because of the very weak vertical shear and the very small vertical shear trend of the background wind. Future changes in the ozone chemistry in the tropical lower stratosphere was investigated by Meul et al. (2014), who showed that an
increase in ozone between 30 and 50 hPa is attributed to changes in the ozone loss rate with slowing the ClO and NOx loss cycles.

Yet, despite the importance of ozone chemistry and dynamics, no study has thus quantitatively examined the future evolution of the ozone QBO from the point of view of combined effects of climate change caused by increasing GHGs and decreasing ozone-depleting substances (ODSs). Therefore, to address this problem, we investigate the future evolution of the ozone QBO in the equatorial stratosphere, by analyzing Meteorological Research Institute (MRI) CCM transient simulations in the future (1960–2100) and historical (1979–2010) runs.

We aim to answer two key questions: 1) How do future changes in stratospheric cooling and tropical upwelling under ozone recovery affect the QBO in ozone in the equatorial stratosphere? And 2) how do these climate changes affect relative contributions of the chemistry and dynamics to the ozone QBO in the whole equatorial stratosphere? Answers to these questions will help us to clarify the roles of chemistry and dynamics in the QBO in ozone and other minor constituents such as methane and nitrogen oxides and may provide an additional insight into why ozone projections by CCM simulations have large spreads in tropical areas.

This paper is organized as follows. Section 2 describes experiments, model descriptions, and data processing. Section 3 shows the present climate for understanding the driving mechanism of ozone QBO. Section 4 gives the future change in the ozone QBO by comparing past, present, and future climates. The final section presents our conclusions.

2. Methods

a. Experiments and model descriptions

The rationale behind the future transient simulation (C2) is to analyze projections of tropospheric and stratospheric ozone under the auspices of the Chemistry–Climate Model Initiative (CCMI) (Eyring et al. 2013). The C2 transient integration includes all anthropogenic and natural forcings based on changes in trace gases and solar variability—but does not consider volcanic eruptions—over a period extending from the past into the future, that is, between 1960 and 2100. We also analyze the historical transient simulation for the period from 1979 to 2010 conducted in specified dynamics mode (C1SD); this simulation is nudged by using the JRA-55 meteorological reanalysis dataset (Kobayashi et al. 2015). The C1SD transient integration includes all anthropogenic and natural forcings due to changes in trace gases, the 11-yr solar cycle, volcanic eruptions, and prescribed SSTs.

The transient simulations (Deushi et al. 2014) are conducted with a new Meteorological Research Institute Earth System Model (MRI-ESM1) (Adachi et al. 2013). The original version of the MRI-ESM1 consisted of a core component of an MRI atmosphere–ocean coupled global climate model (GCM) and subcomponents of ocean, sea ice, aerosol, and ozone models (Yukimoto et al. 2012), using a coupler to interactively integrate those component models (Yoshimura and Yukimoto 2008). A current version of the MRI-ESM1 conducted for the CCMI simulations consists of an atmospheric GCM with a TL159 horizontal resolution, an ocean-ice model (Tsujino et al. 2010), the MRI Chemistry–Climate Model, version 2 (MRI-CCM2), with a T42 horizontal resolution (Deushi and Shibata 2011), and a global aerosol model with a TL95 horizontal resolution (Tanaka et al. 2003). All components of atmospheric models have 80 vertical layers up to 0.01 hPa.

The 11-yr solar cycle is constructed from the daily spectrally resolved solar irradiance data (Lean et al. 2005; from http://solarisheppa.geomar.de/ccmi). We reconstruct future projection to repeat a sequence of the last four solar cycles during a period from 1960 to 2010. The UV-B spectrum from 280 to 315 nm is calculated from the daily solar irradiance data for using an index of solar signal.

b. Data processing

The wavelet analysis is a useful tool for analyzing time series with many different time scales or changes in variance (Torrence and Compo 1998). By decomposing a time series in the time–frequency domain we can obtain a measure of the dominant time-dependent modes of variability. In this study we use the Morlet wavelet (appendix A) that is composed of a complex exponential (carrier) multiplied by a Gaussian window (envelope). To examine fluctuations in the QBO power over a range of scales, the scale-averaged wavelet power is estimated from the weighted sum of the wavelet power spectrum in a band of 20–40 months (appendix A). A Lanczos filter (e.g., Duchon 1979) is applied to monthly mean time series to derive bandpass data within the QBO period (20–40 months) and to derive low-passed data in a low-frequency range much longer than the QBO period (set here to be 60 months).

The uncertainty is used as a confidence interval of the variance and is derived from the standard deviation divided by the square root of a sample size. To estimate the sample size, we account for an effective time scale of sequential persistence by using detrended and deseasonalized daily data (appendix B). In this analysis, we define the effective sampling time scale as twice the exponential decay time during which the serial correlation
exponentially decreases to $1/e$ of its initial value. In the equatorial stratosphere, we estimate the effective sampling time scale (i.e., decorrelation time) for ozone at about 4 months using daily data and for the ozone QBO at 12 months using monthly data. The effective sample size is also used in Student’s $t$ test to determine whether two sets of data are significantly different.

Confidence intervals are closely related to statistical significance (appendix C). In some analyses of our study we will compare means of three independent samples. In that case, triangular comparisons using Student’s $t$ test are so complex that we will display three means associated with uncertainties that represent confidence intervals as twice standard errors denoted by error bars or thin lines, and we use these intervals as a measure of statistical significance. Two statistics are significantly different depending on whether or not their confidence intervals overlap (appendix C). If the two confidence intervals do not overlap, then the statistics will be significantly different at the confidence level of 95% when an effective sample number $N$ is about 20. If the confidence intervals overlap about 10%–20%, two statistics are likely to be significantly different; if the confidence intervals overlap about more than 30%, two statistics are unlikely to be significantly different.

c. Vertical phase profiles for ozone and dynamical QBOs

The wavelet transform is applied to zonally averaged monthly variables (i.e., ozone, zonal wind, the production rate of ozone, etc.) at $10^\circ$–$10^\circ$N to yield time series of QBO phases from $\tan^{-1}\{W_n(s)/R[W_n(s)]\}$ (appendix A). The phase coordinate is set relative to the 20-hPa equatorial zonal wind and a QBO cycle starts at the westerly-to-easterly transition at 20 hPa. To obtain climatological mean of QBO phases, the time series of phases are proportionally redistributed into 24 time intervals (segments), and each segment is averaged over $\sim$10 QBO cycles during $\sim$25 yr from 1961 to 1985.

3. Present climate

a. Ozone QBO

First, to build confidence in our model’s ability to simulate dynamical and ozone QBOs, we present the time evolution of equatorial zonal wind and ozone anomalies during 7 yr of the C1SD and C2 simulations, after deseasonalizing zonal wind and ozone volume-mixing ratio (Fig. 1). It is noted that the zonal wind QBO in C1SD is almost the same as that in the JRA-55 reanalysis (not shown). The zonal wind QBO in the free-running C2 simulation (Fig. 1c) is similar to that in the C1SD simulation in the equatorial stratosphere (Fig. 1a). In other words, the QBO in C2 adequately reproduces that in the JRA-55 reanalysis with lesser amplitude, except in the equatorial lower stratosphere where our model represents weak QBO signatures. Both C1SD and C2 QBOs propagate downward with a period of 2–3 yr. The zonal wind QBO in C1SD has a large amplitude of $20–25$ m s$^{-1}$ in the middle stratosphere to propagate down to the lowermost stratosphere, whereas the zonal wind QBO in C2 is overestimated in the upper stratosphere at 3–10 hPa with amplitude of $\sim 20$ m s$^{-1}$. Time series of the zonal wind at 10 and 70 hPa in the whole C2 simulation period are shown in Figs. 1e and 1f, respectively; these panels represent long-term QBO variability. The QBO amplitude at 70 hPa is reduced after 2040 whereas the trend of QBO amplitude at 10 hPa is not so clear. A long-term reduction of the QBO amplitude in the lower stratosphere is consistent with what Kawatani and Hamilton (2013) indicated.

Prominent QBO signals in the C1SD simulation (Fig. 1b) exhibit downward propagation of the ozone, and the QBO in ozone displays a double-peak structure at the equator, with maxima at 30 and 7 hPa and a phase reversal at around 15 hPa. It is noted that the ozone QBO in C1SD is almost similar to that of observations [e.g., Fig. 3a of Hasebe (1994) derived from SAGE II observations]. The ozone anomalies above 10 hPa in C1SD have magnitudes of 0.3–0.5 ppmv (4%–10% variations from a climatological mean), and those below 10 hPa have magnitudes of around 0.5 ppmv (10% variations) (see also Fig. 9a for climatological mean profiles of C2 ozone). These results are in reasonable agreement with observational analyses from SAGE II (Randel and Wu 1996) and the Microwave Limb Sounder aboard the Upper Atmosphere Research Satellite (Fadnavis and Beig 2008). In the C2 simulation (Fig. 1d), the agreement in amplitude with C1SD is better at 5–10 hPa than lower down. The ozone QBO amplitude ranges between 0.3 and 0.4 ppmv (5%–8% from the climatological mean) at 5–10 hPa, whereas it reduces to 0.2 ppmv (3%) at 30 hPa. The nodal height of the minimum amplitude at 15 hPa is well captured in the model free run.

The power spectrum of the ozone mixing ratio in the frequency domain (periods from 1 to 360 months) in the C1SD and C2 simulations (Fig. 2) shows three distinct distributions: a broad spectral distribution with a peak at around 28 months, a bimodal distribution sharply concentrated at an annual frequency with two maxima at 5–10 and 20–30 hPa, and a narrow monomodal distribution concentrated above 20 hPa at a semiannual frequency with a maximum in the upper stratosphere. The first broad-spectral distribution represents the QBO in ozone. The power spectrum of the ozone QBO has a
primary maximum of amplitude near 7 hPa and a secondary maximum near 20–30 hPa, along with a pronounced minimum near 15 hPa. Quantitatively, the amplitude of the ozone QBO in C2 is reduced to 0.08 ppmv at 7–10 hPa and 0.04 ppmv at 20–30 hPa, that is, to 55% and 35%, respectively, of the C1SD amplitude. The amplitude of the lower ozone QBO peak is weaker than that of the upper peak. Because the power spectrum of the C2 zonal wind is weaker at lower altitudes than that in the C1SD simulation (not shown), the
The power spectrum of the ozone QBO in C2 has smaller amplitude in the equatorial stratosphere.

b. Phases of ozone QBO

Because the stratospheric ozone and atomic oxygen rapidly reach photochemical equilibrium, our model calculates the chemical budget and transport for the odd oxygen family $O_x = O_3 + O(3P) + O(1D)$ (Deushi and Shibata 2011). The local rate of change of odd oxygen family $[O_x]$ is given by

$$\frac{\partial [O_x]}{\partial t} = P_{\text{net}} = P_{\text{chem}} + P_{\text{trans}} + P_{\text{conv}},$$

where $P$ is the rate of production or destruction of the ozone. The net production rate of ozone $P_{\text{net}}$ can be divided here into three parts: that is, the production rates of ozone due to chemistry $P_{\text{chem}}$ and transport $P_{\text{trans}}$ and ozone mixing due to convection $P_{\text{conv}}$. The term $P_{\text{trans}}$ includes vertical as well as horizontal transport of ozone. The convection term is negligible in almost the equatorial stratosphere, except in the lowermost area above the tropopause where the mixing of ozone due to overshooting convective activity is large. Although the ratio of atomic oxygen to ozone increases exponentially with altitude, the abundance of atomic oxygen is three to four orders of magnitude smaller than that of odd oxygen $O_x$, even at the stratopause (~1 hPa). Thus, we regard the dominant form of odd oxygen below 1 hPa as $O_3$.

Figure 3 shows the time series over a period of about five QBO cycles of the ozone mixing ratios, $P_{\text{chem}}$ and $P_{\text{trans}}$ at 5, 9, 30, and 70 hPa for bandpass-filtered data within the QBO period between 20 and 40 months. The pressure levels of 9 and 70 hPa are chosen because relative ratios between two quantities $P_{\text{chem}}$ and $P_{\text{trans}}$ have maxima at these altitudes (as we will show vertical profiles later in Fig. 10c). The pressure level of 5 hPa is chosen as being a representative of ozone QBO in the upper stratosphere, and the level of 30 hPa is chosen where the secondary maximum of ozone QBO is located.

The governing equation of the ozone QBO [Eq. (1)] indicates that a (small) difference from the out-of-phase relationship between the two terms ($P_{\text{chem}}$ and $P_{\text{trans}}$) determines the tendency of ozone QBO and that a tendency term with relatively larger magnitude controls the phase and amplitude of the ozone QBO. The QBOs in $P_{\text{chem}}$ (blue lines) and $P_{\text{trans}}$ (red lines) are essentially in similar magnitude and are out of phase with each other (Figs. 3a and 3c). This means that the net rate of change of ozone, $P_{\text{net}} = P_{\text{chem}} + P_{\text{trans}}$ (recalling that $P_{\text{conv}}$ is negligible at these altitudes), has smaller
magnitude than that of $P_{\text{chem}}$ or $P_{\text{trans}}$. In fact, $P_{\text{net}}$ is preceded or delayed with either $P_{\text{chem}}$ or $P_{\text{trans}}$ by about a quarter cycle, and $P_{\text{net}}$ amplitude is about a tenth of magnitude of either $P_{\text{chem}}$ or $P_{\text{trans}}$ at 5 hPa and a third at 30 hPa, respectively. On the other hand, at the levels where the relative ratio of two quantities has an extremum, $P_{\text{net}}$ at 9 hPa (at 70 hPa) is rather in phase with $P_{\text{chem}}$ ($P_{\text{trans}}$), and $P_{\text{net}}$ amplitude is about the same order of magnitude as $P_{\text{chem}}$ ($P_{\text{trans}}$).

At 5 hPa (Fig. 3a), the QBO in $P_{\text{chem}}$ initiates the QBO in ozone and leads it by about a half cycle, and the QBO in $P_{\text{trans}}$ slightly follows the QBO in ozone. This means that $P_{\text{chem}}$ ($P_{\text{trans}}$) does (does not) play a leading role in the net tendency of ozone. In a region including this altitude, any dynamically induced deviations from photochemical equilibrium are readily canceled out by chemical reactions with the same magnitude as $P_{\text{trans}}$ with opposite sign, and thus the ozone is under chemical control.

At around 9 hPa, ozone upward transport becomes weak because of the mean ozone vertical gradient being small so that amplitude of $P_{\text{chem}}$ becomes larger than that of $P_{\text{trans}}$. In Fig. 3b, the QBO in $P_{\text{chem}}$ ($P_{\text{trans}}$) initiates (follows) the QBO in ozone, indicating that ozone chemistry controls the ozone QBO. At 30 hPa (Fig. 3c), the QBO in $P_{\text{chem}}$ is exactly out of phase with the QBO in ozone, and the QBO in $P_{\text{trans}}$ slightly initiates the QBO in ozone. This level may represent that chemical processes do not quickly respond to perturbations created by dynamics, so that the magnitude of $P_{\text{chem}}$ does not reach the quantity required for photochemical equilibrium of ozone against perturbations caused by dynamics; thus the ozone is driven by the dynamics.

At 70 hPa, $P_{\text{trans}}$ amplitude is larger than $P_{\text{chem}}$ during periods when dynamical QBO amplitude is large, but when dynamical QBO amplitude is small during such a period of 1986–89 the $P_{\text{trans}}$ amplitude becomes smaller than that of $P_{\text{chem}}$. Because QBO signals become weak at this level, overshooting convection activities in the tropics are likely to disturb the QBO. We compare the root-mean-square (RMS) amplitude of $P_{\text{trans}}$ and $P_{\text{chem}}$ averaged over the whole simulation period and find that the RMS of $P_{\text{trans}}$ at 70 hPa ($2.2 \times 10^{-15}$ ppv s$^{-1}$) is 2–3 times larger than that of $P_{\text{chem}}$ ($0.86 \times 10^{-15}$ ppv s$^{-1}$).
Similar results will be shown later (in Fig. 10c), where the relative ratio of $P_{\text{trans}}$ and $P_{\text{chem}}$ amplitude will be presented. In Fig. 3d, except the period with weak signals of the QBO dynamics, the QBO in $P_{\text{trans}}$ ($P_{\text{chem}}$) initiates (follows) the QBO in ozone and leads (lags) it by several months; these characteristics suggest that the QBO dynamics essentially controls the ozone QBO.

Figure 4 depicts the QBO phases of ozone, zonal wind, temperature, vertical wind (the log-pressure velocity), $P_{\text{chem}}$, and $P_{\text{trans}}$, averaged over all QBO cycles during the past climate (CL1; 1960–85). The temperature QBO at the tropics is in thermal wind balance with the mean zonal wind, expressed for the equatorial $\beta$ plane [Andrews et al. 1987, their Eq. (8.2.2); Baldwin et al. 2001] as follows:

$$\frac{\partial u}{\partial z} = -\frac{R}{H\beta} \frac{\partial^2 T}{\partial y^2} \sim \frac{R}{H\beta L^2} \delta T,$$

where $u$ is the zonal wind, $T$ is the temperature, $z$ is the log-pressure height, $y$ is the distance north of the equator, $R$ is the gas constant for dry air, $H \sim 7\text{ km}$ is the scale height, $\beta$ is the meridional derivative of the Coriolis parameter, and $L$ is the meridional scale. Through the thermal wind balance, the vertical derivative of zonal wind with respect to height is in phase with the temperature anomaly; thus, the temperature QBO precedes the zonal wind QBO by a quarter cycle. The temperature QBO precedes both the zonal wind and ozone QBOs by a quarter cycle. The vertical wind QBO is essentially in phase with the zonal wind, with a slight delay in the lowermost stratosphere; the temperature QBO precedes both the zonal wind and ozone QBOs by a quarter cycle. The vertical wind QBO determines the ozone QBO by means of dynamical transport. The thermal wind balance [Eq. (2)] describes that the vertical wind shear of the zonal-mean zonal wind in the equatorial stratosphere is proportional to the meridional curvature of the zonal mean temperature, where the westerly (easterly) shear zone of the QBO produces warm (cold) air masses (e.g., Reed 1964). The temperature anomalies would be damped by infrared cooling (heating), so maintaining this balance requires adiabatic heating (cooling) by a secondary meridional circulation with sinking (rising) at the equator in the westerly (easterly) shear zones. Through vertical advection, the vertical motion associated with the westerly (easterly) shear zones of the QBO produces positive (negative) ozone anomalies owing to the positive vertical gradient of the ozone volume mixing ratio in this region (Hasebe 1994; Cordero and Nathan 2000).

At 20–50 hPa, ozone is still controlled by dynamics. The downward (upward) motion associated with the QBO is induced to maintain the positive (negative) temperature anomalies compensating for radiative damping. However, a feedback of ozone QBO produces diabatic heating (cooling) through absorption of shortwave radiation. For a positive QBO temperature anomaly, for example, the upward transport induced by
ozone heating offsets the downward transport that created the positive ozone anomaly in the first place, thereby reducing the overall downward transport of ozone. This feedback plays a significant role in determining the phase of the ozone QBO. Because the maximum of ozone occurs earlier in time than when ozone heating would not occur, the phase of the ozone QBO is shifted ahead by up to a quarter cycle owing to this feedback; thus, it precedes the phase of the zonal wind QBO by several months (Hasebe 1994). The $P_{\text{trans}}$ and vertical wind QBOs are expected to be out of phase (by half a QBO cycle) in a region at 20–50 hPa.

From 20 to 13 hPa, the photochemical effect gradually gains in importance. This layer is characterized by a transition of the ozone QBO with the abrupt phase change at around 15 hPa; above 13 hPa the ozone QBO phase is almost opposite to what it is, relative to the zonal wind QBO, below 20 hPa. There is little phase difference between the $P_{\text{chem}}$ QBO and the temperature QBO. Here, we can interpret the phase jump at 15 hPa in the ozone QBO as follows. A slight phase shift from the out-of-phase relationship between the two terms with similar magnitude controls the phase of the ozone QBO. Let $P_{\text{trans}}$ be denoted as $\cos(t)$ and $P_{\text{chem}}$ as $\cos(t - \pi - \delta)$; $P_{\text{chem}}$ lags $P_{\text{trans}}$ by $\pi + \delta$. The sum will be $\sim A\cos(t - \pi/2)$, where $A$ is the amplitude = $2\cos(\pi/2 + \delta/2)$. We see from Figs. 3 and 4 that $\delta$ is a small positive at 20–30 hPa leading to a negative value of $A$, while at $\sim$10 hPa, $\delta$ is a small negative resulting in a positive value of $A$. The change of sign in A, appearing as the phase reversal (and a small amplitude) at around 15 hPa, results from a slight phase shift between $P_{\text{trans}}$ and $P_{\text{chem}}$ (change of sign in delta). Therefore, we can understand that the main cause of the phase jump at 15 hPa in ozone QBO will be not the relative magnitude of $P_{\text{trans}}$ and $P_{\text{chem}}$ but the slight phase shift between the two.

At 13–5 hPa, where the ozone QBO is dominated by temperature-dependent photochemistry, ozone variation is 9–12 months ahead of the temperature variation. The phase of the ozone QBO is almost opposite to that of the zonal wind, and it has a phase delay of a few months relative to the vertical wind QBO. Below 10 hPa, the temperature and $P_{\text{chem}}$ QBOs are out of phase, meaning that a negative feedback of temperature on ozone results in an increase in the ozone mixing ratio when QBO-induced upward motion causes adiabatic cooling.

This negative feedback may be rather complicated. In the easterly shear zone, for example, negative temperature anomalies exist satisfying the thermal wind relationship. The cold air masses, subject to radiative damping, are maintained by adiabatic cooling associated with the upward motion. The upward motion will drive a decrease in ozone due to the positive ozone gradient (as we will show the ozone profile later in Fig. 9a). However, the ozone advection is not effective since the vertical ozone gradient is small at this altitude. Instead, the ozone perturbation will be dominated by the temperature-dependent ozone chemistry (equilibrium ozone is high in cold environment), leading to the positive ozone anomaly in the easterly shear (upward motion) zone (Fig. 4). This may cause positive heating anomalies due to shortwave absorption, which, in addition to longwave heating, must be balanced by stronger upward motion. Thus, the secondary circulation will be intensified by the ozone QBO in the region in which ozone QBO is driven by chemistry. This stronger upward motion may drive the ozone decrease, partially compensating for the positive ozone anomaly due to chemistry.

The phase relationships between vertical wind and $P_{\text{trans}}$ and between temperature and $P_{\text{chem}}$ gradually shift from being out of phase at 10 hPa to in phase at 5 hPa. Above 5 hPa ($\sim$38 km), the ozone QBO lags the zonal wind QBO by a quarter cycle but in phase with the QBOs in the vertical velocity and $P_{\text{trans}}$. The ozone QBO is almost out of phase with the QBOs in the temperature and $P_{\text{chem}}$. It is clear enough why $P_{\text{trans}}$ shifts from being out of phase with vertical wind below 10 hPa to in phase above 5 hPa as a consequence of the sign change in the vertical gradient of mean ozone below and above this level. It looks as if the gradual phase shift of $P_{\text{chem}}$ with height above 15 hPa might be driven by $P_{\text{trans}}$ but dynamically induced deviations from photochemical equilibrium are accompanied with nitrogen as well as ozone, implying that the dynamical transport itself contributes to the ozone chemistry.

Identification of the dominant contributors to the ozone QBO in the upper stratosphere has been a matter of controversy. Some studies emphasized the importance of the temperature-dependent photochemistry (e.g., Ling and London 1986), the nitrogen loss cycle (Chipperfield et al. 1994; Tian et al. 2006), ozone QBO signals without QBO-induced NO$_x$ transport (Butchart et al. 2003), or a different regime relationship between O$_3$ and NO$_2$ above 5 hPa (Hauchecorne et al. 2010). Therefore, we will only see in a later section how the photochemical loss rate varies with height, and addressing these topics about detailed chemical budgets about ozone and nitrogen in the upper stratosphere is beyond the scope of the current study.

4. Future climate

a. Climate change in ozone

Next, we examine the evolution of profile and total column ozone anomalies relative to the 141-yr long-term
mean at the equator (10°S–10°N) in the C2 simulation, after removing variabilities shorter than 60 months (Figs. 5b and 5c), accompanied with a time series of solar irradiance (280–315 nm) (Fig. 5a). Although the QBO is the dominant component of equatorial ozone variability in the stratosphere, we filter out variations with shorter periods including QBO so that we could more easily detect the long-term trend and decadal variability. The profile ozone anomalies show multi-decadal variability being vertically elongated, which is just synchronized with 11-yr-like solar signals. It is found from Fig. 5b that the ozone mixing ratio decreases in the lower stratosphere below 20 hPa in a long-term range because an increase in the tropical upwelling owing to global warming brings more ozone-poor tropospheric air into the lower stratosphere (e.g., Akiyoshi et al. 2010).

We identify three periods from multi-decadal variability: a period before 1985, characterized by positive ozone anomalies; a period between around 1990 and 2030, characterized by negative anomalies; and a period after around 2040, also characterized by positive anomalies. The middle period roughly corresponds to the time interval when ozone depletion is occurring in the equatorial stratosphere. Therefore, we select three periods, 1960–85, 1990–2020, and 2040–70, as being representative of typical ozone climates, and we refer to these three periods as the past (CL1), present (CL2), and future (CL3) climates, respectively. The total column ozone (Fig. 5c), which is close to the 141-yr mean, also shows such decadal variability, accompanied with the 11-yr solar signal. Column ozone at the equator roughly is characterized by positive, negative, and positive anomalies from the 141-yr
mean, respectively, during 1960–85, 1990–2030, and 2040 onward.

In CL1, before ozone severely depleted in the equatorial stratosphere, positive ozone anomalies are present throughout the stratosphere with a maximum of 0.4 ppmv at 5 hPa. In CL2, severe ozone depletion occurs owing to ODSs originating from halogens through gas-phase reactions, and there are negative ozone anomalies in the upper to middle stratosphere with a minimum of −0.2 ppmv at 5 hPa.

In CL3, positive ozone anomalies are present in the upper to middle stratosphere, with a maximum of 0.3 ppmv at 7 hPa. It is interesting to see that the ozone recovery is prominent at 5–10 hPa, and this largely accounts for positive column ozone anomalies at the equator. In the future climate, the recovery of ozone in the middle and upper equatorial stratosphere would be expected to occur for two reasons (World Meteorological Organization 2014): 1) a decrease in the ODSs and 2) an increase in the negative temperature-dependent photochemistry under climate change. An increase in CO2 causes stratospheric cooling due to the emission of longwave radiation into space. Strong cooling, which occurs in the middle and upper stratosphere, causes the ozone-depletion rate to slow down in that region, which results in an increase in ozone abundances.

Figure 6a shows the latitude–height section of the local QBO wavelet power amplitude of the ozone mixing ratio for the past climate. The QBO amplitude is derived from the bandpass wavelet power amplitude within the QBO period from 20 to 40 months (appendix A). This figure exhibits a double peak structure with amplitude maxima at 7–9 and 25 hPa, along with a node at 15 hPa, in the equatorial stratosphere. The primary maximum of amplitude at 7–9 hPa is about 0.5 ppmv, and the node amplitude is less than 0.1 ppmv. The variation of the ozone QBO widens into the subtropical stratosphere at the 10-hPa level with amplitude of 0.1–0.15 ppmv, about one-third the amplitude at the equator. The subtropical ozone QBO is produced by the secondary meridional circulation induced by the QBO with advection of ozone out of the tropics (Baldwin et al. 2001; Fadnavis and Beig 2008).

Figures 6b–d depict the differences of the ozone QBO amplitude of CL3 − CL1, CL3 − CL2, and CL2 − CL1, respectively. The mean amplitude is derived from the average over each climate period, and we compare the difference between two means. A statistical significance is derived from a Student’s t test where an effective sample size (decorrelation time) is estimated as 12 months from the bandpass data. The difference of ozone QBO amplitude between the future and past
climates (Fig. 6b) is characterized as a significant positive anomaly with a maximum of 0.1 ppmv at 5–10 hPa at the equator, positive anomalies at the subtropics in both hemispheres, and a negative anomaly in the equatorial lower stratosphere.

The climate difference of CL3 – CL2 corresponds to the difference between the future ozone recovery and the present ozone depletion. Significant positive anomalies (Fig. 6c) are found in a broad region at roughly 5–20 hPa from the equator to SH midlatitudes; a negative anomaly is present in the equatorial lower stratosphere. The climate difference of CL2 – CL1 corresponds to the difference between the present ozone depletion and the past pre-ozone depletion. Figure 6d shows positive anomalies at 5 hPa and the equator and in the northern subtropics at 7–20 hPa and 20º–40ºN. A positive 5-hPa anomaly at the equator is surrounded by negative anomalies, and negative anomalies are found in the midlatitude upper stratosphere in both hemispheres.

A possible mechanism to explain why the ozone QBO amplitude in the equatorial middle to upper stratosphere increases in the future climate may include amplification of the temperature QBO and/or strengthening of the residual mean meridional circulation. Although the mechanism is unclear at this moment, we will confirm later that the temperature QBO amplitude is enhanced in this region. The negative impact of temperature on photochemistry could contribute to a net increase in ozone. The subtropical positive anomalies can be attributed to enhanced advection of the net increased ozone out of the tropics.

Figure 7a shows the evolution of the profile of the ozone QBO power amplitude at the equator from 1960 to 2100, and Figs. 7b and 7c present the time series of the amplitude at 7 and 40 hPa, respectively. The profile is generally characterized by the double peak structure with maxima at 7 and 25 hPa, along with the node at 15 hPa. Figure 7a does not show much systematic variation in the ozone QBO (e.g., at 7 hPa) except in the lower stratosphere (e.g., at 40 hPa), where the ozone QBO amplitude is small especially after 2020. The node amplitude at around 15 hPa does not exhibit any significant trend from 1960 to 2040, but after that the node amplitude becomes stronger.

Figure 8 shows the evolution from 1960 to 2100 of the QBO amplitude of the ozone mixing ratio in the wavelet power spectrum from 20 to 40 months at 7, 15, 40, and 70 hPa at the equator. The ozone QBO power amplitude is largely confined to between 22 and 32 months, with a peak power at a period of 28 months, during most of the simulated time interval. The ozone QBO power amplitude exhibits a long-term decrease at 40 hPa and an increase at 15 hPa. At the end of the twenty-first century, the peak period of the ozone QBO power is somewhat shortened to 24 months. This would be caused by further acceleration of the Brewer–Dobson circulation
responsible for enhanced synoptic-scale waves owing to the elevated GHGs. The ozone QBO power also shows decadal variability, that is, sporadically strengthening and weakening amplitude. This variability may be relevant to the 11-yr-like solar signals, but details of the mechanisms between the ozone QBO and the solar signals would be investigated in a future study.

b. Future change in ozone QBO

In this subsection, we further examine the future change in ozone QBO by comparing the three climates. At first, Figs. 9a and 9b present ozone profiles of the climatological-mean ozone mixing ratio for the three climate periods in the equatorial stratosphere (10°S–10°N) and their relative ratios compared with the past climate, respectively. The abscissa in Fig. 9b (and Fig. 9d) is taken from the ratio minus 1 in percent. The ozone profiles for the climates (Fig. 9a) exhibit a maximum of 9 ppmv at 10 hPa, which is reduced by 10% compared to an observed value during a period from 1996 to 2010 (Tegtmeier et al. 2013). Above 10 hPa the ozone anomalies of CL2 and CL3 represent deviations of 4%–12% relative to CL1, and below 10 hPa the CL2 and CL3 deviations from CL1 show a range of 3%–15%.

At 15–20 hPa, in contrast, the CL2 and CL3 ozone deviations from CL1 are very small (Fig. 9b).

The abundance of ozone in the tropical lower stratosphere is primarily determined by the vertical advection of ozone. The future climate is characterized by an increase in mean tropical upwelling driven by intensified wave drag due to an increase in the long-lived GHGs (e.g., Butchart 2014). The faster the tropical upwelling is, the more ozone-poor the air is; thus, the ozone abundances are expected to decrease in the tropical lower stratosphere. In the equatorial middle to upper stratosphere (above 15 hPa), a significant decrease in ozone of up to 7% is found in Fig. 9b in the present climate compared with the past climate. In contrast, the ozone mixing ratio significantly increases by 2%–5% at all altitudes above 15 hPa in the future climate. The impact of stratospheric cooling induced by increasing GHGs is expected to be a net increase in the ozone mixing ratio in the equatorial middle to upper stratosphere.

Figures 9c and 9d depict vertical profiles of the mean ozone QBO amplitude for the three climates averaged over 10°S–10°N and their relative ratios of amplitude in percent compared with the past climate. The QBOs in
The residual vertical velocity, zonal wind, and temperature are also shown in Figs. 9e and 9f. The vertical profiles of ozone QBO amplitude in the equatorial stratosphere represent a maximum of 0.3–0.35 ppmv at 7–9 hPa and a maximum of 0.18 ppmv at 25 hPa, along with a minimum of 0.1 ppmv at 15 hPa. Compared with the past climate, the ozone QBO in the future climate shows a significant increase in amplitude in the upper stratosphere as well as at around 15 and 70 hPa and a significant decrease at around 40 hPa. The ozone QBO in the present climate depicts insignificant changes at levels of 7–50 hPa and a significant increase at around 70 hPa, compared with the past climate. Reduced amplitude at around 40 hPa in the future ozone QBO can be explained by decreased...
amplitude of residual vertical velocity associated with the zonal wind and temperature QBOs (Fig. 9f). The minimum amplitude near 15 hPa would be caused by the slight phase shift between $P_{\text{chem}}$ and $P_{\text{trans}}$ as discussed before. Compared with the present climate, the ozone QBO in the future climate shows a significant increase in amplitude at 7–10 hPa. This increase can be attributed to the recovery of the future ozone profile against the depletion of the present ozone. There is also an increase in ozone QBO amplitude in CL2 relative to CL1 during the time of ozone depletion (red line in Fig. 9d).

It is interesting to see that the future ozone QBO amplitude increases at 70 hPa, although zonal wind QBO reduces its amplitude (Fig. 9f). To interpret the change in ozone QBO amplitude in the lower stratosphere, it is important to consider the vertical gradient of the mean ozone mixing ratio and the QBO component of vertical velocity $w$ because the local rate of change of ozone with respect to time can be approximated as follows:

$$\frac{\partial [O_x]}{\partial t} = P_{\text{chem}}^Q + P_{\text{trans}}^Q,$$

and

$$P_{\text{trans}}^Q \sim w^Q \frac{\partial [O_x]}{\partial z} + w \frac{\partial [Q_x]}{\partial z}, \quad (3)$$

where the overbar denotes the climatological mean and the superscript $Q$ represents the QBO component. The term $\frac{\partial [O_x]}{\partial z}$ denotes the vertical gradient of the basic-state (climatological mean) ozone volume mixing ratio. As for an estimation about the vertical advection of zonal momentum made by Shibata and Deushi (2012), we estimate the vertical advection of ozone. The first term on the right-hand side of Eq. (3) represents the interaction between the mean ozone shear and the QBO secondary circulation, and the second term represents the contribution of the background vertical wind to the ozone QBO shear. Below the secondary maximum of the ozone QBO, the magnitude of each term on the right-hand side of Eq. (3) at 70 hPa is, on average

$$|w^Q| \sim 0.015 \text{ mm s}^{-1}, \quad w \sim 0.19 \text{ mm s}^{-1},$$

$$\frac{\partial [O_x]}{\partial z} \sim 0.38 \text{ ppmv km}^{-1},$$

$$\frac{\partial [Q_x]}{\partial z} \sim 0.011 \text{ ppmv km}^{-1},$$

where $|x|$ denotes the absolute value of $x$. Thus, the first and second terms on the right-hand side of Eq. (3) are $\sim 5.7 \times 10^{-15}$ and $2.1 \times 10^{-15} \text{ ppvs}^{-1}$, respectively. These estimations indicate that the QBO-induced vertical advection of the mean ozone mixing ratio is the major tendency term of the QBO component of the vertical ozone transport. We further estimate future changes of these terms in Eq. (3). Because the amplitude of the ozone QBO increases in the layer at 50–100 hPa as Fig. 9c (also Fig. 9d) indicates, the vertical shear of the ozone QBO should increase. It is found that the second term in Eq. (3) gains in importance in future and that its relative ratios to the QBO component of the vertical ozone transport are 11%, 21%, and 23%, respectively, for CL1, CL2, and CL3.

From Fig. 9b the vertical gradient of mean ozone mixing ratio at 50–100 hPa shows only a subtle change between the three climate periods, but it seems to decrease in the future (i.e., CL1 > CL2 > CL3). The amplitude of the vertical velocity QBO in CL2 and CL3 is larger than that in CL1 (Fig. 9e). Therefore, increased amplitude of the ozone QBO at around 70 hPa can be attributed to the change in the vertical velocity QBO.

Figures 10a and 10b present vertical profiles of $P_{\text{chem}}$ and $P_{\text{trans}}$, QBO amplitude for the three climates. Because the $P_{\text{chem}}$, QBO is almost offset against $P_{\text{trans}}$, QBO amplitude profiles have a similar shape to those for $P_{\text{trans}}$. The $P_{\text{chem}}$ amplitude for the future climate increases by 40% at 1–20 hPa, compared with the present climate (Fig. 10a). The profile of CL2 can be larger or smaller at different altitudes relative to that of CL1. Figure 10c represents the QBO amplitude of the relative ratio of $[P_{\text{trans}}]$ to $[P_{\text{chem}}]$ for the three climates. In the lower stratosphere $P_{\text{trans}}$ has a dominant component of the net rate of change of ozone and its amplitude is twice larger than that of $P_{\text{chem}}$. At 20 hPa, $P_{\text{trans}}$ QBO amplitude is almost the same as that of $P_{\text{chem}}$. At 10–20 hPa, $|P_{\text{chem}}|$ is larger than $|P_{\text{trans}}|$, suggesting that $P_{\text{chem}}$ drives the net rate of change of ozone. The relative ratio of amplitude between $P_{\text{chem}}$ and $P_{\text{trans}}$ shows a sharp peak with large $P_{\text{chem}}$ at around 9 hPa, where the vertical gradient of the mean ozone mixing ratio becomes small and vertical advection of ozone is very weak. Above 5 hPa, the amplitude of the QBO in $P_{\text{chem}}$ again approaches half the relative ratio.

A ratio of 0.5 means that $P_{\text{trans}}$ and $P_{\text{chem}}$ have the same magnitude. However, this does not mean that $P_{\text{trans}}$ plays a significant role in variation of the ozone QBO. Rather, any dynamically induced deviations from photochemical equilibrium are readily canceled out by chemical reactions with the same magnitude as $P_{\text{trans}}$ with opposite sign, and thus the ozone is under chemical control. In this context, it is interesting to see the profile in Fig. 10a again. The gradual growth up to $\sim 18$ hPa may reflect the activation of chemical processes due to the increase of photolysis rate. The deviations from such a smooth growth above this level may represent that chemical processes become fast enough to respond to perturbations created by dynamics, and thus the magnitude of $P_{\text{chem}}$ is controlled by the requirement of maintaining photochemical equilibrium amount of ozone against perturbations caused by dynamics.
It is also interesting to see that the change in the magnitude of the QBO in those production rates of ozone for the different climates is especially large at around 7–10 and 70 hPa. The magnitude of the QBO in $P_{\text{trans}}$ in the present and future climates is larger than that in the past climate. Because the vertical gradient of the ozone is small at around the altitude of the ozone maximum, the transport term is small in magnitude at 7–10 hPa. This means that changes in tropical upwelling can cause a large influence on the production rates of ozone. A mechanism about the 70-hPa peak will be discussed later.

Figures 10d and 10e show the ratio of amplitude relative to the past climate (CL1) in percent in the vertical profiles of $P_{\text{chem}}$ and $P_{\text{trans}}$ for the three climates. The amplitude of the QBO in $P_{\text{trans}}$ (Fig. 10e) depicts a significant increase in the future climate at around 10 hPa as well as in the uppermost stratosphere and does a significant decrease at around 40 hPa in the equatorial lower stratosphere, compared with the past climate (CL3/CL1). A similar change near 70 hPa is also present in CL3/CL1 (albeit more weakly), consistent with Fig. 9e, which shows the increased vertical velocity for both the CL3 and CL2 climates. A significant increase in $P_{\text{chem}}$ amplitude occurs at 10 and 1–2 hPa. However, the ozone QBO (Fig. 9d) shows almost no change in amplitude at 10 hPa. Thus, the change in $P_{\text{chem}}$ nearly offsets that in $P_{\text{trans}}$ at this level.

Again, the interesting question is that the magnitude of $P_{\text{trans}}$ at 70 hPa is strengthening in the future climate whereas the zonal wind QBO is weakening in the lower stratosphere (Fig. 9f). The thermal wind balance would require the weakening of temperature QBO proportional to the weakening of the vertical wind shear of the zonal wind QBO. The temperature QBO in Fig. 9f, however, shows larger amplitude at 50–100 hPa in CL2 and CL3, compared with CL1. To explain these results we think again about the thermal wind balance equation [Eq. (2)]. This equation describes that the vertical shear of the zonal wind is proportional to the meridional gradient of zonal mean temperature, assuming equatorial symmetry and using L'Hôpital's rule (Andrews et al. 1987). This means that the zonal wind shear is proportional to the meridional curvature of zonal mean temperature rather than the gradient (i.e., second derivative). If the temperature QBO is amplified and the meridional...
curvature of the temperature at the equator is more flattened, the vertical wind shear of the zonal wind QBO would become small. We roughly estimate characteristic lengths of meridional curvature of the temperature QBO at 70 hPa by the half-width of the equator peak if a profile is close to a Gaussian curve or by the distance from a peak to a minimum if a profile has a minimum before it drops to half of its peak. In fact, we find that those lengths are roughly 900, 1200, 1400 km, respectively, for CL1, CL2, and CL3. Thus, the amplification of the temperature, vertical wind, and ozone QBOs at 70 hPa could be attributed to the meridional curvature of the temperature QBO being small (i.e., to the flattening of the meridional curvature at the equator in the future and present climates). A further study is needed for causality of amplification of ozone and temperature QBOs in conditions of the zonal wind QBO weakening in the lower stratosphere in future climate, but this is beyond scope of this paper.

In this study, we have shown that the ozone QBO amplitude in CL3 decreases at 20–50 hPa and increases at 5–9 hPa, compared with CL1 (Fig. 9d). As already shown in Fig. 1, our model presented weak ozone QBO amplitude at 20–50 hPa in C2 relative to CI0. This result implies that the future change in ozone QBO amplitude in this region might be amplified if the model would more realistically simulate the dynamical QBO in the lower stratosphere. Also, discrimination of the shallow and deep branches of Brewer–Dobson circulations will be important because the response to global warming could be different between the two (Diallo et al. 2012). A better representation of dynamical processes, especially in the equatorial lower stratosphere, will be required for improving the interannual variability of ozone and other trace gases in climate models in a future work.

5. Conclusions

We have investigated climate change in the ozone QBO in the equatorial stratosphere from 1960 to 2100, by analyzing the future transient simulation. The analyses of the vertical phase profiles have revealed that the phase lines of dynamical (zonal wind, temperature, and vertical velocity) QBOs have a rather simple vertical structure characterized by almost steady downward propagation, whereas those of the ozone and the local rates of change of the ozone QBO due to chemistry ($P_{\text{chem}}$) and due to transport ($P_{\text{trans}}$) exhibit a more complex vertical structure (Fig. 4). The ozone QBO is nearly in phase with the temperature QBO in the lower stratosphere below 20 hPa, while both the ozone and temperature QBOs are nearly out of phase with each other above 15 hPa. We can interpret the phase jump of the ozone QBO at 15 hPa as a slight phase shift between $P_{\text{trans}}$ and $P_{\text{chem}}$ QBOs and not as the relative magnitude between the two since the slight phase shift from the out-of-phase relationship between the two terms with similar magnitude controls the phase of the ozone QBO. The term $P_{\text{chem}}$ shows a gradual phase shift with the temperature QBO from out of phase below 10 hPa to in phase above 5 hPa, and $P_{\text{chem}}$ and $P_{\text{trans}}$ are essentially out of phase with each other in almost the entire equatorial stratosphere.

We have shown that the climate change and ozone recovery will impact the complicated evolution of the ozone QBO in the tropical stratospheric circulation. Compared with the present climate (1990–2020), the ozone QBO in the future climate (2040–70) shows a significant increase in amplitude by 15% at 7–10 hPa and a decrease by 20% at 40 hPa. The increase at 7–10 hPa can be attributed to the future recovery ozone against the depletion of the present ozone (Figs. 6c and 9c). The reduced amplitude at around 40 hPa is caused by the decrease in residual vertical velocity amplitude, which is brought by the weakening vertical shear of zonal wind QBO at this level.

Compared with the past climate (1960–85), the ozone QBO in the future climate is characterized by the significant increase by 25% in amplitude at 5–10 hPa, the significant decrease by 20% at 40 hPa, and the increase by up to 60% at 70 hPa (Figs. 6b and 9c). The change in ozone QBO amplitude at 5–10 hPa suggests that temperature-dependent photochemistry plays an important role in the enhanced future ozone QBO there. The decreased amplitude at 40 hPa is caused by the weakening vertical shear of the future zonal wind QBO.

The interesting finding of this study is that while the zonal wind QBO is weakening in the lower stratosphere in the future (Fig. 9f), the QBOs in ozone, temperature, and vertical velocity are strengthened in this region (Figs. 9e, f). The thermal wind balance equation [Eq. (2)] describes that the zonal wind shear is proportional to the meridional curvature of zonal mean temperature rather than the gradient (i.e., second derivative). If the temperature QBO is amplified while its meridional curvature of the temperature is more flattened, the vertical shear of the zonal wind QBO would be small. We roughly estimate characteristic lengths of meridional curvature of the temperature QBO at 70 hPa and find that they are roughly 900, 1200, and 1400 km, respectively, for CL1, CL2, and CL3. Therefore, the increase in the temperature, vertical wind, and ozone QBO amplitude at 70 hPa could be attributed to smaller meridional curvature—that is, more flattening of the meridional curvature of the dynamical QBO in the equatorial lowest stratosphere in the future and present climates. Further studies are needed because the robustness of future predictions of the dynamical QBO is still uncertain, especially in the lower stratosphere.
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APPENDIX A

Wavelet Transform

The wavelet transform, which can analyze time series at many different frequencies, is described in detail by Torrence and Compo (1998). A wavelet function \( \psi(\eta) \) that depends on a nondimensional time parameter \( \eta \) has zero mean and is localized in both time and frequency. The basic wavelet we have chosen here is the Morlet wavelet, which consists of a plane wave modulated by a Gaussian:

\[
\psi_\delta(\eta) = \pi^{-1/4} e^{i\omega_0 \eta} e^{-\eta^2/\delta^2}, \tag{A1}
\]

where \( \omega_0 \) is the nondimensional frequency, taken as 6. The continuous wavelet transform \( W_n \) of a discrete time sequence \( x_n \) is defined as the convolution of \( x_n \) with a scaled and transformed version of \( \psi_\delta(\eta) \):

\[
W_n(s) = n \sum_{n'=0}^{N-1} x_{n'} \psi_{s}(n-n'\delta t/s), \tag{A2}
\]

where the asterisk denotes the complex conjugate, \( s \) indicates the wavelet scale (period), and \( N \) is the number of points in the time series. By varying the wavelet scale \( s \) and translating along the localized time index \( n \), one obtains the amplitude versus scale relationship and the variation of amplitude with time. Because the wavelet function \( \psi(\eta) \) is complex, the wavelet transform \( W_n(s) \) is also complex. The transform can then be divided in the real part \( \Re[W_n(s)] \) and imaginary part \( \Im[W_n(s)] \) or amplitude \( |W_n(s)| \), phase \( \tan^{-1}(\Im[W_n(s)]/\Re[W_n(s)]) \), and wavelet power spectrum \( |W_n(s)|^2 \).

Once a wavelet function is chosen, it is necessary to choose a set of scales \( s \) to use in the wavelet transform. It is convenient to write the scales as fractional powers of 2:

\[
s_j = s_0 2^{j\delta j}, \quad j = 0, 1, \ldots, J, \text{ and} \tag{A3}
\]

\[
J = \delta j^{-1} \log_2(N\delta t/s_0), \tag{A4}
\]

where \( s_0 \) is the smallest resolvable scale and \( J \) determines the largest scale. The choice of a sufficiently small \( \delta j \) depends on the width in spectral space of the wavelet function. In this study, we take \( N = 1692, \delta t = 1 \) month, \( s_0 = \delta t, \delta j = 0.25, \) and \( J = 44 \), giving a total of 45 scales ranging from 1 month up to 170 yr. While dealing with finite-length time series, errors will occur at the beginning and end of the wavelet power spectrum, as the Fourier transform assumes that the data are cyclic. One solution to minimize the error is to pad the end of the time series with zeros before doing the wavelet transform and then remove them afterward.

To examine a fluctuation in power over a range of scales (a band), one can define the scale-averaged wavelet power as the weighted sum of the wavelet power spectrum over scales \( s_1 \) to \( s_2 \) as follows:

\[
\bar{W}_n = \frac{\delta \tilde{\eta}}{C} \sum_{j=1}^{s_2} |W_n(s_j)|^2. \tag{A5}
\]

The scale-averaged wavelet power can give a measure of the bandpass-averaged variance over time. In this study, we derive the bandpass data within the QBO period from \( j = 19 \) (s1 = 23 months) to \( j = 22 \) (s2 = 39 months).

APPENDIX B

Effective Sampling Size

The independence of the sequential daily and monthly data with a sample number of \( N \) that are subtracted from the climatological annual cycle can be estimated by using a lag correlation:

\[
\rho_\tau = \frac{1}{N - \tau} \sum_{t=1}^{N-\tau} X(t)X(t+\tau), \tag{B1}
\]

where \( \rho_\tau \) is the autocorrelation coefficient and \( \tau \) is the lag time. In this study, the effective sampling time \( T_e \) is defined as twice the shortest lag time for which \( \rho_\tau \) falls below \( 1/e \). Thus, the effective sample size or number \( N_e \) is set to \( N/T_e \). The effective sample size can be used to estimate the standard error \( \sigma_e \) as

\[
\sigma_e^2 = \frac{\sigma^2}{N_e} = \frac{1}{N N_e} \sum_{t=1}^{N} [X(t) - \bar{X}]^2, \tag{B2}
\]

where \( \sigma \) is the standard deviation of the mean and \( \bar{X} \) is the mean.

A statistical significance of the difference \( Z \) in variable \( X \) between two populations 1 and 2 is tested if the sampling distribution of any statistics of the independent random variable is normal or nearly normal. \( Z \) is defined as
If we take an effective sample number as \( N \) (with twice the standard error, so they are calculated as standard errors of the two. We assume that dependent samples have their means (\( s \)) different depending on whether or not their confidence interval will not contain 0. Two statistics are significantly 0 at a significance level of 0.05, then the 95% confidence interval is performed and a statistic is significantly different from 0. (B3)

\[
Z = \frac{X_1 - X_2}{\sqrt{\frac{\sigma_1^2}{N_1} + \frac{\sigma_2^2}{N_2}}}
\]

Confidence Intervals and Significance

Confidence intervals are closely related to statistical significance testing. A confidence interval is a range of values that contains a population mean and a margin of error (e.g., uncertainty). The significance level, which defines the distance from the null hypothesis values, is a probability of rejecting the null hypothesis when it is true. For example, if a test of corresponding hypothesis is performed and a statistic is significantly different from 0 at a significance level of 0.05, then the 95% confidence interval will not contain 0. Two statistics are significantly different depending on whether or not their confidence intervals overlap. If the confidence intervals do not overlap, then the statistics will be significantly different. If they overlap, it is not necessarily true that they are not significantly different.

We suppose to compare the means from two independent samples. We assume that the two independent samples have their means \( X_1, X_2 \) with standard deviations \( \sigma_1, \sigma_2 \) for (effective) sample numbers \( N_1, N_2 \). Here, we take confidence intervals with twice the standard error, so they are calculated as \( (X_1 \pm 2\sqrt{\frac{\sigma_1^2}{N_1}}, X_2 \pm 2\sqrt{\frac{\sigma_2^2}{N_2}}) \). Then, we consider a case in which the mean difference is just twice the standard errors of the two. We assume that \( \sigma_1 \sim \sigma_2 \) and \( N_1 \sim N_2 \), and then the \( t \) statistics for comparing the two means yields

\[
Z = \frac{X_1 - X_2}{\sqrt{\frac{\sigma_1^2}{N_1} + \frac{\sigma_2^2}{N_2}}} = 2\sqrt{\frac{\sigma_1^2}{N_1} + \frac{\sigma_2^2}{N_2}} \sim 2\sqrt{2} \sim 2.8.
\]

If we take an effective sample number as \( N_1 \sim 20 \), then a value of the \( t \) distribution (two tailed) is 2.09 at a significance level of \( \alpha = 0.05 \). This means that the null hypothesis that the means of two samples are the same should be rejected at this level. In other words, the random error associated with the estimated mean difference is quantified at the confidence interval of 95\% (i.e., \( 1 - \alpha \)). If the confidence intervals overlap, two statistics are significantly different or not depending on how much their confidence intervals overlap with each other. In the above example, if we take a mean difference as 1.5 standard errors of the two, the \( t \) statistics for comparing two means yields \( Z \sim 2.12 \). From this result we can (cannot) say that the mean difference is statistically significant when the confidence intervals overlap by less (more) than 25\%.

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