The Quasi-Biennial Oscillation in a Double CO₂ Climate

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Abstract

The effects of anticipated twenty-first-century global climate change on the stratospheric quasi-biennial oscillation (QBO) have been studied using a high-resolution version of the Model for Interdisciplinary Research on Climate (MIROC) atmospheric GCM. This version of the model is notable for being able to simulate a fairly realistic QBO for present-day conditions including only explicitly resolved nonstationary waves. A long control integration of the model was run with observed climatological sea surface temperatures (SSTs) appropriate for the late twentieth century, followed by another integration with increased atmospheric CO₂ concentration and SSTs incremented by the projected twenty-first-century warming in a multimodel ensemble of coupled ocean–atmosphere runs that were forced by the Special Report on Emissions Scenarios (SRES) A1B scenario of future atmospheric composition. In the experiment for late twenty-first-century conditions the QBO period becomes longer and QBO amplitude weaker than in the late twentieth-century simulation. The downward penetration of the QBO into the lowermost stratosphere is also curtailed in the late twenty-first-century run. These changes are driven by a significant (30%–40%) increase of the mean upwelling in the equatorial stratosphere, and the effect of this enhanced mean circulation overwhelms counteracting influences from strengthened wave fluxes in the warmer climate. The momentum fluxes associated with waves propagating upward into the equatorial stratosphere, and the effect of this enhanced mean circulation overwhelms counteracting influences from strengthened wave fluxes in the warmer climate. The momentum fluxes associated with waves propagating upward into the equatorial stratosphere, and the effect of this enhanced mean circulation overwhelms counteracting influences from strengthened wave fluxes in the warmer climate.

1. Introduction

The quasi-biennial oscillation (QBO) is a persistent, quasi-periodic, large-amplitude oscillation of the low-latitude stratospheric circulation [see Baldwin et al. (2001) for a review]. The QBO is most evident in the zonal mean zonal wind near the equator, which undergoes reversals from strong easterlies to strong westerlies through each QBO cycle, but has clear observable signals in temperature and meridional circulation as well as the concentration of ozone and other trace constituents in the tropical stratosphere. The generally accepted theory of the QBO is that it results largely from interactions of the stratospheric mean flow with vertically propagating internal waves that are generated in the tropical troposphere (Lindzen and Holton 1968; Plumb 1977) and so is an oscillation generated internally within the atmosphere.

There is evidence that the tropical QBO has significant remote dynamical effects on the circulation in the extratropical stratosphere (e.g., Holton and Tan 1980) and in the extratropical lower atmosphere even down to the surface (e.g., Ebdon 1975; Coughlin and Tung 2001; Thompson et al. 2002; Boer and Hamilton 2008; Marshall and Scaife 2009). These extratropical effects are strongly modulated by season, so it is possible that the interaction of annual and QBO cycles is responsible for some of the longer period (notably quasidecadal) variability in the atmosphere (e.g.,
Hamilton 2002; Anstey and Shepherd 2008). In the tropical stratosphere itself the QBO is strong enough that it may have a significant role in determining the mean chemical composition and hence mean climate.

In situ evidence for tropical stratospheric mean flow wind reversals date to near the beginning of the twentieth century, although systematic direct observations are available only from about 1950 (Hamilton 1998). Since 1950 we have evidence of a reasonably stable zonal wind QBO with a mean period of about 28 months (ranging between 22 and 34 months in individual cycles) and rather consistent vertical structure from cycle to cycle (e.g., Naujokat 1986; Baldwin et al. 2001). Several authors have tried to use the indirect evidence from surface pressure observations of the solar atmospheric tides to infer long-term behavior of the QBO (Hamilton and Garcia 1984; Teitelbaum et al. 1995; Brönnimann et al. 2007). The results of these studies provide at least some evidence for fairly dramatic long-term changes in QBO behavior.

There has been interest in the question of how the QBO might respond to changes in external forcing of the climate system, such as that due to solar insolation variations (Salby and Callaghan 2000; Hamilton 2002; Kuai et al. 2009) or the presence of enhanced stratospheric aerosols after major volcanic eruptions (Dunkerton 1983). Another obvious source of external forcing to the climate system is the increasing greenhouse gas concentrations that are believed to be largely responsible for observed global warming over the last century and are predicted to strongly influence climate in the future. How the QBO might respond to large increases in atmospheric carbon dioxide (CO₂) concentration has provoked speculation dating at least to the work of Fels (1985) and has been the subject of a more recent comprehensive modeling study by Giorgetta and Doege (2005, hereafter GD). Understanding and projecting global climate effects on the QBO would help in interpreting the observed variability of the QBO. If greenhouse warming is expected to strongly affect the QBO, then some evidence may be sought in the currently available record. In addition, a credible prediction of the climate change perturbation on the QBO would give some context in which to consider the large long-term changes in QBO behavior claimed in indirect observational studies such as Teitelbaum et al. (1995). In addition, of course, the QBO itself is an important aspect of the chemistry–climate system and thus a significant factor in accurate projections of climate change.

As noted above, the QBO is driven by the zonal momentum fluxes from vertically propagating waves generated in the troposphere. The basic dynamics of the QBO has been embodied in mechanistic models of various complexity (e.g., Lindzen and Holton 1968; Plumb 1977; Saravanan 1990; Takahashi and Boville 1992), which impose wave fluxes at the lower boundary of the stratosphere and then simulate the mean flow response. The results of such models show that the QBO amplitude, period, and vertical structure are largely controlled by wave forcing and the large-scale upwelling circulation. Larger wave amplitudes (in the horizontal phase speed range that interacts strongly with the mean flow) lead to faster mean flow accelerations and shorter QBO periods. The effects of wave driving are counteracted by mean upwelling, which acts to slow the QBO phase descent and lengthen the period. In fact, the upwelling can be so strong as to prevent a mean flow oscillation, and Saravanan (1990) suggests that the lower boundary of the QBO penetration is determined by the profile of mean upwelling.

Projecting how the QBO will change in response to increased greenhouse warming (or other external perturbations) requires comprehensive models that can self-consistently simulate the changes in wave fluxes and mean vertical motion. This problem is particularly challenging as most current comprehensive global climate models (GCMs) do not simulate a realistic QBO, and indeed typically display very little interannual variability in the tropical stratosphere. There is no evidence that any of the models employed in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) model intercomparison simulated a stratospheric QBO. However, a few GCMs have had some success in simulating stratospheric mean flow oscillations that are similar in many respects to the observed QBO. A number of models have accomplished this only by incorporation of a parameterization of momentum transport by nonstationary gravity waves (e.g., Scaife et al. 2000; Giorgetta et al. 2002, 2006; Shibata and Deushi 2005; Kulyamin et al. 2009). Only a few models have produced QBO-like oscillations with only explicitly resolved waves (Takahashi 1996, 1999; Hamilton et al. 1999, 2001; Watanabe and Takahashi 2005; Kawatani et al. 2005, 2009, 2010a,b). Such models typically have fine vertical resolution allowing an explicit representation of a small vertical wavelength of internal waves.

The earlier model study of QBO changes under global warming conditions by GD employed the middle atmosphere ECHAM5 (MAECHAM5) model at T42L90 resolution, which produced a stratospheric QBO by incorporating the nonstationary gravity wave parameterization of Hines (1997). GD performed a control integration that simulates a QBO with mean period of about 29 months. They then performed three runs with sea surface temperatures (SSTs) and atmospheric composition changed to represent a doubled CO₂ climate: in one experiment they left the parameterized gravity wave sources (imposed at 600 hPa in their model) unchanged from the control and in the other two experiments they increased the
assumed gravity wave source amplitudes by 10% and then 20% above the control value. The simulated QBO in each case has a shorter period in the doubled CO2 runs: 26, 22, and 17 months respectively in the 0%, 10%, and 20% changed source experiments (although GD note that the period change in the 0% run cannot be judged statistically significant with just the 20-yr simulation they analyzed). The period reduction appeared to result from both an increase in wave forcing and a decrease in the mean upwelling in the warmed climate simulations.

In the present paper we investigate the effects of greenhouse gas–induced climate change on the QBO using a model that simulates a fairly realistic stratospheric QBO without any parameterized nonstationary gravity wave effects. Our study thus avoids the somewhat arbitrary assumptions about climate change effects on gravity wave source strength needed in the earlier work of GD and is thus the first completely self-consistent simulation of QBO response to external perturbation. As shown below, we also find quite different results from those of GD.

This paper is arranged as follows. Section 2 describes the model and experimental design. Section 3 describes the climatological differences. Section 4 investigates the changes of the QBO and residual circulation. Section 5 examines the changes of wave activity. Section 6 summarizes the study and provides the concluding remarks.

2. Model description and experimental design

The model we used is based on the atmospheric component of version 3.2 of the Model for Interdisciplinary Research on Climate (MIROC), a coupled atmosphere–ocean GCM developed by the Center for Climate System Research/National Institute for Environmental Studies/Frontier Research Center for Global Change (CCSR/NIES/FRCGC; Hasumi and Emori 2004). The atmospheric GCM has been referred to in previous studies as the CCSR/NIES AGCM and the CCSR/NIES/FRCGC AGCM. The hydrostatic primitive equations on a sphere are used. The model has a horizontal resolution of T106 spectral truncation, which corresponds to a grid interval of approximately 120 km in the tropics (1.125°). Seventy-two vertical layers are used (L72), with the top boundary at 1.2 hPa (~47 km). The vertical resolution is set to 550 m from ~300 to 5 hPa.

The cumulus parameterization is based on the method of Arakawa and Schubert (1974). In the original Arakawa–Schubert scheme, convective precipitation characteristically becomes more frequent and weaker as the horizontal resolution of the GCM increases. To prevent this problem, a relative humidity limit method is incorporated into the cumulus convection scheme (Emori et al. 2001). If the ratio between the vertical integration of the specific humidity and that of the saturation specific humidity from the bottom to the top of a cloud is less than a critical value (here, 0.72), the cloud mass flux is set to zero [see Emori et al. (2001) for further details]. This approach suppresses overly frequent precipitation and improves the generation of organized convective precipitation. Convectively coupled and stratospheric equatorial trapped waves (EOWs) with relatively small equivalent depth (\(h_e \leq 90 \text{ m}\)) are well simulated using this method (Kawatani et al. 2009, 2010a).

The radiative transfer scheme is based on the two-stream discrete ordinate method and a correlated k-distribution method. The Mellor and Yamada (1982) level-2 closure scheme is used for eddy vertical diffusion parameterization. This parameterization mainly represents vertical mixing associated with gravity wave breaking due to both convective instability and shear instability in the GCM. A dry convective adjustment is applied to eliminate the convective instability that is not suppressed by the vertical diffusion parameterization. The \(v^4\) hyper-viscosity diffusion is used, and the e-folding time for the smallest resolved wave is 0.9 days. To reduce unrealistic reflection of vertically propagating waves from the model top, a sponge layer is defined at elevations above 5 hPa. The sponge layer consists of five levels in which the strength of the \(v^4\) horizontal diffusion is successively doubled (i.e., 2, 4, 8, 16, and 32 times) with respect to standard value. Much more detailed model descriptions are presented in Watanabe et al. (2008).

This experiment included mountain-induced gravity wave parameterization by McFarlane (1987) to obtain a realistic large-scale circulation at mid to high latitudes. However, nonstationary gravity wave parameterization is not included. Hence, the simulated QBO is driven by explicitly resolved waves in the model.

We ran a 90-yr control integration of the model with observed monthly mean SSTs and sea ice from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) climatology averaged from 1979 to 1998. An 90-yr future climate integration was then run using monthly mean SSTs and sea ice values incremented by the predicted changes from 1979–98 to 2080–99 in a multimodel ensemble [phase 3 of the Coupled Model Intercomparison Project (CMIP3)] of coupled ocean–atmosphere runs that were forced by the Special Report on Emissions Scenarios (SRES) A1B atmospheric composition scenario. Specifically, the results for SST and sea ice changes of Mizuta et al. (2008) were used. There are no interannual variations in the SSTs employed in either the control or future climate runs.

The CO2 concentration in the control run was taken to be 345 ppmv and this was doubled to 690 ppmv in our future climate run, roughly consistent with the CO2 changes...
projected between 1979–98 and 2080–99 in the SRES A1B scenario (hereafter, we refer to the control and double CO$_2$ runs as the “present” and “future” climate, respectively. “Differences” are defined as future minus present). The United Kingdom Universities’ Global Atmospheric Modelling Programme (UGAMP) monthly ozone climatology data are used (available online at http://badc.nerc.ac.uk/data/ugamp-o3-climatology/). The concentrations of other long-lived greenhouse gases such as methane and ozone were not changed in our future climate run. Hence, the influence of ozone change is not considered in the study. Garcia and Randel (2008) suggested that the behavior of polar ozone is not crucial to the change in the Brewer–Dobson circulation.

The initial condition consists of data from the T213L256 AGCM (Watanabe et al. 2008). The 1 January result of the T213L256 simulation after a 3-yr integration period was interpolated onto the T106L72 grid. The same data are used in present and future runs as the initial condition. The model is integrated for 90 yr. One-hourly mean data were archived for years 81–90 in order to facilitate the detailed analysis of transient waves described in section 5 below.

3. Climatological differences

In this section, the climatological differences between present and future climate simulations with statistical significance greater than 95% are discussed. The statistical significance is based on the two-sided Student’s $t$ test for sampling the 90 individual yearly mean data for each of the present and future runs.

Figure 1 shows the annual mean SST that is used as the boundary condition in the present and future climates. In the future climate, SST increases by about 1.9°C in the global mean and 2.1°C in the 30°S–30°N mean. Many areas of increase are located in the equatorial and midlatitude regions (Fig. 1c). In the eastern Pacific, the meridional gradient of SST difference is apparent; the warming is greater in the equatorial regions and smaller in the mid-latitudes of the Southern Hemisphere. Zonally inhomogeneous SST differences, as seen during El Niño or La Niña, are not apparent. The future sea ice used in this simulation is decreased in both Arctic and Antarctic regions (not shown).

We examined differences in our simulated 2-m air temperatures (not shown). The global distribution of these 2-m temperature differences is characterized by larger warming over the continents than over the ocean and strongly enhanced warming at high northern latitudes. These features are also seen in the CMIP3 multimodel ensemble mean.

Figure 2 shows the annual mean climatological global distributions of precipitation in the present and future climates and the difference between them. The differences with statistical significance greater than 95% are colored in Fig. 2c. In the present climate, the model reasonably simulated the spatial distribution of precipitation, including such features as the separation between the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ), and the strong precipitation over Africa and South America (Fig. 2a). The model also closely simulated spectral signals of convectively coupled EQWs in the zonal wavenumber–frequency domain (not shown; see Kawatani et al. 2009, 2010a). The amount of zonal mean precipitation in the present climate is quantitatively similar to that of the real atmosphere in the equatorial
region, but the model overestimates the midlatitude precipitation. In the future climate, precipitation increases significantly over the equator, but it does not increase much around the ITCZ region (Figs. 2b,c). Characteristics, such as increasing (decreasing) precipitation in the equatorial (subtropical) region are also seen in the CMIP3 multimodel ensemble mean (not shown). In the future climate, zonal mean precipitation over the equator increases by ~20%.

Figures 3a and 3b show latitude–height cross sections of climatological annual mean zonal mean temperature and zonal wind in the present and future climate. The black and red contours correspond to present and future climate, respectively. The differences with statistical significance greater than 95% are colored. The dashed lines in Fig. 3b show the tropopause, which is defined as the $2.5 \times 10^{-4} \text{s}^{-2}$ line of the squared buoyancy frequency $N^2$. The tropopause shifts slightly higher in the future climate (Fig. 3b). A warming in the troposphere and cooling in the stratosphere are evident (Fig. 3a). At 300–50 hPa, the latitudinal gradient of temperature differences is large due to warming in the tropical upper troposphere and cooling in the extratropical lowermost stratosphere. The upper parts of the subtropical jets strengthen, consistent with thermal wind balance (Fig. 3b). The upward displacement of a zero wind line is obvious in the extratropics. The increase of the westerlies (eastward wind) in the lower stratosphere should enhance the upward propagation of westward-propagating gravity waves, as well as orographic gravity waves. We also note that the zero-wind lines shift equatorward from the upper troposphere to the stratosphere, which causes more extratropical Rossby waves to propagate into the equatorial region. These points are discussed in the next section. Figure 3c is the same as Figs. 3a and 3b, but for the square of the buoyancy frequency $N^2$. Because of the temperature changes, the static stability increases in the equatorial troposphere, whereas it decreases in the equatorial stratosphere. The differences are obvious in the equatorial lower stratosphere around 100–50 hPa. Other notable differences are seen in the midlatitude tropopause region.

4. Changes in the QBO and residual circulation

a. QBO in the present and future climates

First, we analyze the QBO variations over the 90-yr simulation period. Figure 4 shows a time–height cross section of the monthly mean, zonal mean zonal wind over
the equator in the present and future climates. The red and blue colors correspond to westerlies and easterlies, respectively. In the present climate, an obvious QBO-like oscillation with a period of approximately 2 yr is apparent (Fig. 4a). The period of the simulated QBO varies little from cycle to cycle, suggesting that the simulated oscillation is phase locked to the annual cycle. The maximum speed of the easterly is approximately 22.5 m s\(^{-1}\), and that of the westerly is 15 m s\(^{-1}\). In contrast, Naujokat (1986) reported 23.5 and 20 m s\(^{-1}\) for maximum speed of the easterly and westerly winds, respectively. So the simulated oscillation has somewhat weaker amplitude but the same east–west phase asymmetry as in observations. The QBO wind variations extend down to approximately 60–80 hPa, but the amplitude in the lower stratosphere is smaller than that in the real atmosphere. The downward propagation of easterly shear zones (\(\partial u/\partial z < 0\)), which agrees with observations.

The QBO in the future climate differs from that in the present climate (Fig. 4b). It does not extend as far down into the lower stratosphere. In the future climate, the 0 m s\(^{-1}\) lines of the westerly phase of the QBO extend down to \(\sim 50\) hPa and sometimes to 20–30 hPa, but they extend down to \(\sim 70\) hPa in the present climate. The \(-5\) m s\(^{-1}\) lines of the easterly phase of the QBO extend down to \(\sim 80\) hPa in both the present and future climates, but the level of \(-10\) m s\(^{-1}\) lines in the future climate are higher than those in the present climate. Consequently, the amplitude of the QBO in the future climate becomes much smaller, especially in the lower stratosphere. The periods of the QBO in the future climate become longer and more irregular. It might be expected that stronger equatorial precipitation in the future (Fig. 2c) would generate more waves, and then the periods of the QBO would become shorter (cf. GD). However, these results are in the opposite direction.

Figure 5 shows the frequency power spectra of zonal mean zonal wind at the equator as a function of height. The fast Fourier transform (FFT) method is used to calculate the spectra. The most dominant period of the QBO in the present climate is 24 months. There are other peaks at 22 and 27 months, with much smaller variance. The relatively strong variances extend to \(\sim 70\) hPa, with a maximum at \(\sim 10\) hPa. The future QBO has a broad range of periods around 22–33 months and large variance around at 25–29 months. The dominant periods of the QBO in the future climate become longer by about 1–5 months. The QBO variances in the future become much weaker below 50 hPa.

b. Change of the residual circulation

Equatorial mean residual upwelling acts to retard the downward propagation of the QBO wind reversals (e.g., Saravanan 1990; Dunkerton 1991). How the upward residual circulation changes in the future climate is therefore also an important factor in the future QBO. Several studies have shown a rather robust tendency for GCMs to simulate stronger Brewer–Dobson circulations, including increased equatorial upwelling, in global warming climates (Butchart and Scaife 2001; Butchart et al. 2006; Garcia and Randel 2008; McLandress and Shepherd 2009; Okamoto et al. 2010, manuscript submitted to J. Geophys. Res.) In this section, we investigate how and why the simulated residual circulation in our model changes in the future.

The Eliassen–Palm flux (EP flux), which is widely used to analyze wave propagation and zonal wave forcing in the meridional plane of the zonal mean zonal wind, is defined as follows in spherical and log-pressure coordinates (Andrews et al. 1987):
$$F^{(\phi)} = \rho_0 a \cos \phi \left( \frac{\overline{v' \theta'}}{\overline{\theta'}} - \overline{u' v'} \right),$$

$$F^{(z)} = \rho_0 a \cos \phi \left[ \left( [f - (a \cos \phi)^{-1}(\overline{u \cos \phi})_\phi] \right) \right. \times \left. \frac{\overline{v' \theta'/\overline{\theta'}}}{\overline{\theta'}} - \overline{u' w'} \right),$$

$$\mathbf{V} \cdot \mathbf{F} = (a \cos \phi)^{-1} \frac{\partial [F^{(\phi)} \cos \phi]}{\partial \phi} + \frac{\partial F^{(z)}}{\partial z},$$

with the zonally averaged momentum equation expressed as

$$\overline{\pi}_t = \overline{\pi} [f - (a \cos \phi)^{-1}(\overline{\pi \cos \phi})_\phi] - \overline{\pi} \overline{\pi}_z$$

$$+ (\rho_0 a \cos \phi)^{-1} \mathbf{V} \cdot \mathbf{F} + \mathbf{X}.$$  

In the above equations, $\rho_0$, $a$, $\phi$, $z$, $u$, $v$, $w$, $\theta$, and $f$ are respectively the log-pressure height-dependent density, the mean radius of the earth, latitude, log-pressure height, zonal wind, meridional wind, vertical wind, potential temperature, and the Coriolis parameter ($f = 2\Omega \sin \phi$, where $\Omega$ is the rotation rate of the earth). The subscripts $\phi$, $z$, and $t$ denote the meridional, vertical, and time derivatives, respectively. The mean residual circulations of the meridional and vertical components are expressed by $\overline{\pi}^*$ and $\overline{w}^*$. Eastward and westward wave forcing correspond to the EP flux divergence and convergence (i.e., $\mathbf{V} \cdot \mathbf{F} > 0$ and $\mathbf{V} \cdot \mathbf{F} < 0$), respectively. To quantitatively investigate the mean ascent in the equatorial lower stratosphere, the residual vertical velocity in the transformed Eulerian mean (TEM) formation is calculated as follows (Andrews et al. 1987):

$$\overline{\pi}^* = \overline{\pi} + (a \cos \phi)^{-1} \left( \cos \phi \overline{\theta'/\theta'} \right)_\phi.$$  

FIG. 4. Time–height cross section of the monthly mean, zonal mean zonal wind over the equator in (a) present and (b) future climates. The color intervals are 5 m s$^{-1}$. 
in the equatorial lower stratosphere, where heating anomalies are positive.

Figure 7 shows the wave forcing anomalies due to all resolved waves, resolved waves with $1 \leq s \leq 11$ and with $12 \leq s \leq 106$, and zonal forcing due to mountain gravity wave parameterization. At around 100–50 hPa, resolved westward wave forcing anomalies are obvious in $10^\circ$–$80^\circ$N and $5^\circ$–$50^\circ$S (Fig. 7a). These anomalies are mainly due to waves with $1 \leq s \leq 11$ (not shown), suggesting that more Rossby waves propagate equatorward because of increasing westerly anomalies (solid lines in the figure) in the future climate. Westward forcing anomalies due to the mountain gravity wave parameterization occur in the midlatitude around 30–50 hPa in both the Northern and Southern Hemispheres, whereas eastward forcing anomalies are seen below the westward forcing anomalies (Fig. 7d). Shifting the zero line of the zonal wind anomaly causes the parameterized mountain gravity wave to propagate upward more, and then westward wave forcing anomalies are formed at a higher altitude whereas eastward wave forcing anomalies are formed at a lower altitude.

Different wave forcing distributions between the present and future climate are associated with structural changes in the zonal mean zonal winds in the upper troposphere and lower stratosphere. The wave forcing differences are mainly due to large-scale waves ($1 \leq s \leq 11$) and parameterized mountain gravity waves. The results shown in Figs. 6 and 7 are qualitatively consistent with those of Garcia and Randel (2008) and McLandress and Shepherd (2009). Wave forcing anomalies for $12 \leq s \leq 106$ also cause westward forcing around $30^\circ$S and $30^\circ$N in the lower stratosphere, corresponding to background zonal wind changes (Fig. 7c). However, the differences in these wave forcings are much smaller than those for other wave forcing types, indicating that future $\pi^*$ change could be investigated using lower-resolution GCMs.

Figure 8a shows the climatological annual mean vertical profiles of $\pi^*$ in the present and future climates averaged from $20^\circ$S to $20^\circ$N. The secondary circulation associated with the QBO is ascent (descent) in the easterly (westerly) shear, which leads to interannual $\pi^*$ variation. We also plot profiles of $\pi^* \pm 1$ standard deviation. The standard deviation is calculated from monthly mean $\pi^*$ data (i.e., 1080 samples for 90 yr). Therefore, variations with periods $\approx 2$ months, such as variations with annual cycle and QBO cycle, are included in the standard deviation. In the present climate, the averaged $\pi^*$ is approximately $(2.4 \pm 0.6) \times 10^{-1}$ mm s$^{-1}$ near 70 hPa, decreases to $(1.7 \pm 0.4) \times 10^{-1}$ mm s$^{-1}$ at ~50 hPa, and then increases to $(3.0 \pm 0.8) \times 10^{-1}$ mm s$^{-1}$ at ~15 hPa.
Figure 8b shows the vertical profile of the ratio of the future $\bar{\omega}$ to the present $\bar{\omega}$; $\bar{\omega}$ is strengthened about 1.4 times around 70 hPa and 1.1 times around 15 hPa. It is strengthened much more in the lower stratosphere than in the middle to upper stratosphere, as shown in Fig. 6b.

Our model’s present-day equatorial residual upwelling velocity profile compares well with various observational estimates based on studying the propagation of tracer signals (the so-called tape recorder signal) (e.g., by Mote et al. 1996 and Schoeberl et al. 2008) and also based on
the gradient of the estimated age of air as discussed by Eyring et al. (2010; see top left panel of their Fig. 5.6). Eyring et al. (2010) also show simulated equatorial residual vertical velocities between 100 and 10 hPa for a number of GCMs. The present model $w^*$ profile is similar to those of these other models, although some of the GCMs have somewhat larger $w^*$ values overall.

The response of the tropical residual upwelling velocity at 70 hPa to global warming was compared in 10 GCMs by Butchart et al. (2006, see their Fig. 6). The models were run in different experiments (e.g., short time slice experiments, transient runs, equilibrium simulations) but in all cases the results indicate increased tropical upwelling with warming global climate. Quantitative comparisons with the present experiments are made difficult by the different experiments, but most of the model predictions are consistent with increases of equatorial $w^*$ of $\sim$(20%–40%) in response to doubling of atmospheric CO$_2$, which would put our model at the high end of the other models’ sensitivity in this respect.

5. Changes in wave activity

a. Zonal forcing changes

In this section, we analyze the changes in wave activity in more detail using hourly sampled data. The phases of the QBO in present and future climate are nearly the same at the end of year 80 (see Fig. 4). Here, we discuss the results from year 81 to year 90. In this section, statistic significances of differences are calculated by the two-sided Student’s $t$ test using 10 individual yearly mean samples for each of the present and future runs.

Figures 9a and 9b show the time–height cross sections of monthly mean zonal mean zonal wind and its vertical shear in the 10°S–10°N average field in the present and future climate. The vertical shears of the zonal mean zonal wind in the future climate are weaker than those in the present climate, especially in the lower stratosphere, where $w^*$ is much strengthened (Fig. 8). However, the vertical shears do not change much in the upper stratosphere. Figures 9c and 9d show the EP flux divergence due to the resolved waves. The zonal forcing due to the mountain gravity wave parameterization is negligible compared to that due to the resolved waves in the 10°S–10°N mean field (not shown), indicating that it does not play a role in driving the QBO. In the mid- to lower stratosphere, the absolute values of EP flux divergence in the future climate are smaller than those in the present climate. Smaller EP flux divergences imply a longer period for the QBO. The differences of EP flux divergence are smaller at high altitudes, where the differences in $w^*$ are also small (Fig. 8), as those of the vertical shear of zonal wind are.
Figures 9e and 9f show the time variation of the tendency of zonal mean zonal wind \( \overline{u} \); the left side of Eq. (4)], EP flux divergence due to the resolved waves, and forcing due to the residual circulation [the first plus the second term of the right side of Eq. (4)] at 30 hPa averaged from 10\(^\circ\)S to 10\(^\circ\)N. Generally, forcing due to the residual circulation (blue lines) is opposite to the resolved wave forcing (red lines), and its absolute value is smaller than that of resolved wave forcing. Both the absolute values of resolved wave forcing and those of the forcing due to residual circulation in the future climate are smaller than those in the present climate. The weaker vertical shear of zonal wind \( \overline{u} \) in the future climate reduces forcing associated with \( \overline{w} \), \( \overline{u} \), and \( \overline{w} \) [see second term on the right side of Eq. (4)] despite strengthened \( \overline{w} \). The absolute value of forcing associated with \( \overline{w} \) in both the present and future climate is smaller than that of \( \overline{w} \), \( \overline{u} \) near the equator. The tendency of zonal wind (black lines) in the future climate is also smaller than that in the present climate, which corresponds to a longer period of the QBO in the future climate.

b. Change in the wave momentum flux

The smaller EP flux divergence around the QBO in the future climate (Fig. 9) does not necessarily result from a decrease in wave momentum flux, since the EP flux divergence depends not only on wave momentum fluxes but also on the background zonal wind shear of the QBO. In the future climate, the smaller vertical shear of the zonal wind captures fewer gravity waves per unit altitude (note that wave dissipation occurs when \( C_s \sim \overline{w} \)). Hence, the absolute value of EP flux divergence per unit altitude would become smaller even if total amount of wave momentum flux is unchanged. The changes in wave momentum flux in the future climate are investigated in this subsection.

Before investigating the wave momentum flux, we examine the differences in moist heating (i.e., cumulus convective heating plus large-scale condensation heating) because moist heating is the strongest wave source in the equatorial region. Figure 10 shows vertical profiles of 10-yr mean moisture heating and standard deviation of moisture heating due to 12\( \overline{w} \) between 20\(^\circ\)S and 20\(^\circ\)N. The standard deviation is calculated from hourly data, and therefore variations with periods larger than 2 h are included. Mean heating in the future is larger than that in the present above 400 hPa. The mean specific humidity in the future is also larger than that in the present (not shown), which would have been related to the increased moist heating. The standard deviations of moist heating due to both 12\( \overline{w} \) in the future are also larger than those in the present climate, which implies that more waves are generated in the future climate in both zonal wavenumber ranges.

Next, the change of wave momentum flux is investigated. Here, we discuss the vertical flux of zonal momentum \( u \)\( \overline{w} \) associated with waves due to 12\( \overline{w} \). Most of the EP flux divergence due to 12\( \overline{w} \) consists of the vertical divergence of \( u \)\( \overline{w} \) [second term of the right side of Eq. (2)]. Waves with 12\( \overline{w} \) are regarded as internal inertia–gravity waves, which play a significant role in driving the QBO (Kawatani et al. 2010a,b). We also discuss vertical group velocity \( C_{gz} \)
changes of these gravity waves. The net $\overline{u'w'}$ is composed of positive $(\overline{u'w'})_+$ and negative $(\overline{u'w'})_-$ momentum fluxes; $|\overline{u'w'}|$ is the sum of the absolute values of $\overline{u'w'}$,

$$\overline{u'w'} = (\overline{u'w'})_+ + (\overline{u'w'})_-,$$

and

$$|\overline{u'w'}| = |(\overline{u'w'})_+| + |(\overline{u'w'})_-|.$$  \hspace{1cm} (6)

To investigate how the wave momentum changes, $|\overline{u'w'}|$ rather than $\overline{u'w'}$ should be investigated because a large cancellation between positive and negative momentum fluxes reduces the net value of $\overline{u'w'}$ (Sato and Dunkerton 1997).

Figures 11a and 11b show latitude–height cross sections of the ratio of future to present climate for the absolute value of wave momentum flux $|\overline{u'w'}|$ due to waves with $12 \leq s \leq 106$, and vertical profiles of the $|\overline{u'w'}|$ of present and future climate between 10°S and 10°N. The regions with differences with statistical significance greater than 95% are colored in Fig. 11a. The wave momentum flux $|\overline{u'w'}|$ in the future climate is larger above $\sim 170$ hPa in the equatorial region, whereas $|\overline{u'w'}|$ decreases below $\sim 170$ hPa. These differences result from the upward shift of the vertical profile of $|\overline{u'w'}|$ in the future compared with the present climate (Fig. 11b). At 150–60 hPa in the equatorial region, $|\overline{u'w'}|$ increases by 10%–15% in the future climate, which would be related to strengthened variances of moisture heating with $12 \leq s \leq 106$ in the upper troposphere (Fig. 10). Note that $|\overline{u'w'}|$ associated with waves due to $1 \leq s \leq 11$ also increased significantly by 10%–15% in the equatorial region at 100–60 hPa (not shown).

Figures 11c and 11d are the same as Figs. 11a and 11b, respectively, but for $C_{gz}$. For internal gravity waves, $C_{gz}$ can be estimated from the total energy and vertical energy flux as follows (Gill 1982):

$$C_{gz} = \Phi w'/E,$$  \hspace{1cm} (8)

where $\Phi$ and $E$ are the geopotential and total energy (i.e., the potential energy plus the kinetic energy), respectively.
The $C_{gz}$ anomaly is negative below and positive above \(\sim 170\) hPa in the equatorial region (Fig. 11c). The level of the change of $C_{gz}$ sign is nearly equal to that of $j u^9$.

The maximum $C_{gz}$ in the upper troposphere shifts upward from $\sim 155$ hPa in the present to \(\sim 145\) hPa in the future.

In the future climate, static stability in the equatorial upper troposphere–lower stratosphere (UTLS) region decreases (Fig. 3c), which changes the properties of vertically propagating gravity waves. The $C_{gz}$ of gravity waves is $C_{gz} = \frac{N k}{m^2} = \frac{\omega^2}{N k}$, (9) where $k$, $m$, and $\omega$ are the zonal wavenumber, vertical wavenumber, and intrinsic frequency, respectively. For zonally propagating waves, the intrinsic frequency $\omega$ is $\omega = \omega - k \Pi$. (10)

Equations (9) and (10) indicate that if the background zonal wind $\Pi$ does not change with height, $C_{gz}$ becomes large for small values of $N$. Zonal wind anomalies are small in the equatorial UTLS region (see black contours in Figs. 11a,c). Here we are interested in the region above 80 hPa, where the differences in the QBO between present and future climates occur. The differences in moist heating are negligible above 80 hPa (Fig. 10). The positive $C_{gz}$ anomaly corresponds to a negative anomaly of $N^2$ (Fig. 3c). These results indicate that gravity waves propagate faster vertically in the future climate than in the present climate.

c. Changes in the wave fluxes as a function of the zonal phase velocity

In the equatorial lowermost stratosphere at $\sim (70$–80) hPa, the ratios of future to present climate for the absolute value of the vertical component of the EP flux $F^{(z)}$ for both $1 \leq s \leq 11$ and $12 \leq s \leq 106$ increase by up to 10%–15% similar to that of $[\overline{u w}]$ (not shown). Here we need to investigate whether these increased wave momentum fluxes result in increased forcing of the QBO.

To investigate the zonal phase velocity distribution relative to the ground ($C_v$) of $F^{(z)}$, we calculate the zonal wavenumber–frequency distribution of $F^{(z)}$: $F^{(z)}(s, \omega) = \rho_0 a \cos \phi \text{Re} \{[f - (a \cos \phi)^{-1}(\Pi \cos \phi)]_s \}$ $\times \hat{v}(s, \omega) \hat{u}^*(s, \omega) \partial_z - \hat{u}(s, \omega) \hat{w}^*(s, \omega)$. (11)

In Eq. (11), the asterisk denotes a complex conjugate, and $\hat{u}$, $\hat{v}$, $\hat{w}$, and $\hat{\theta}$ are the Fourier coefficients of the zonal, meridional, and vertical winds, and the potential temperature (cf. Horinouchi et al. 2003), respectively.

We selected an altitude of 71–82 hPa for the analysis. The zonal mean zonal wind at 71–82 hPa and 10°S–10°N in the present climate is about $\sim 3$ m s$^{-1}$, and its difference between the present and future climates is negligible (see contour lines in Figs. 11a,c). The 71–82-hPa level is the lowest part of the equatorial stratosphere, and the background zonal wind is weak. In addition, this altitude range is not affected by the phase of the simulated QBO.
Note that the ground-based frequency $\omega$ and zonal wave-number $k$ are conserved in the vertical, assuming that the background flow does not change with time or longitude, respectively. The distribution of zonal wavenumber–frequency spectra of momentum fluxes would change only if a wave were to undergo critical-level filtering and/or dissipation (Ern et al. 2008 and references therein). Therefore, the space–time spectra of $F^{(c)}$ at 71–82 hPa is suitable for investigating how wave momentum fluxes relevant to the QBO change in the future climate. The
spectra are calculated for successive overlapping segments of data and then averaged. Here, 72 days with 12 days of overlap between each segment are calculated (the total number of segments is 60 for 10 yr).

Figures 12a and 12b show the zonal wavenumber and frequency spectra of $F^{(c)}$ at 71–82 hPa between 10°S and 10°N averaged over 10 yr. The solid lines depict $C_x$. Positive zonal wavenumbers correspond to positive $C_x$ (eastward propagation) and negative zonal wavenumbers correspond to negative $C_x$ (westward propagation). In both present and future climates, $F^{(c)}$ is mostly distributed over a wide range of $|C_x|$ for both positive and negative zonal wavenumbers. Kawatani et al. (2010a) showed the same figure as Fig. 12, but for the $F^{(c)}$ and EP flux divergence at altitudes where the phase of the QBO changes from easterly to westerly (westerly shear) and westerly to easterly (easterly shear). They showed that the eastward forcing in the westerly shear occurs in the $2 \leq C_x \leq 20$ m s$^{-1}$ range, whereas the westward forcing in the easterly shear occurs in the $-30 \leq C_x \leq -5$ m s$^{-1}$ range. The wave forcing with faster $C_x$ occurs at a higher level, such as the altitude of the stratopause semiannual oscillation (SSAO) (Kawatani et al. 2010b).

Figure 12c shows the ratio of $F^{(c)}$ between the present and future climates (i.e., the future divided by the present). As mentioned, $|F^{(c)}|$ including all spectral ranges of waves increased up to 10%–15%. Spectral analysis reveals that $F^{(c)}$ with $2 \leq C_x \leq 20$ m s$^{-1}$ does not increase, whereas $F^{(c)}$ with $C_x \geq 20$ m s$^{-1}$ increases significantly. Therefore, the $F^{(c)}$ relevant to the westerly phase of the QBO does not increase in the future. Concerning the $F^{(c)}$ associated with the westward waves, $F^{(c)}$ for $-15 \leq C_x \leq 0$ m s$^{-1}$ increases, especially for $-10 \leq C_x \leq 0$ m s$^{-1}$. However, $F^{(c)}$ for $-30 \leq C_x \leq -15$ m s$^{-1}$ does not increase much. So $F^{(c)}$ relevant to the weak easterly phase of the QBO increases in the future, whereas that relevant to the relatively strong easterly phase does not.

The equatorial $\Pi^8$ increases in the future climate, especially in the lower stratosphere (Fig. 8). The lowermost levels of the westerly phase of the QBO in the future climate are higher than those in the present climate, whereas the $-5$ m s$^{-1}$ lines of easterly phase of the QBO extend down to $\sim 80$ hPa in both the present and future climates, despite the higher $-10$ m s$^{-1}$ lines in the future climate (Figs. 4 and 9). These results are related to the stronger $F^{(c)}$ for $-10 \leq C_x \leq 0$ m s$^{-1}$ and the nearly identical $F^{(c)}$ for $-30 \leq C_x \leq -15$ m s$^{-1}$ and $2 \leq C_x \leq 20$ m s$^{-1}$ in the future climate as compared with the present climate.

To examine why the EP flux differences depend on the zonal phase velocity $C_x$, we plot the same spectra but for precipitation in the 10°S–10°N averaged field (Fig. 13). The mean precipitation increases in the equatorial region in the future climate (Fig. 2). Interestingly, the spectral differences of precipitation are similar to those of $F^{(c)}$ in the lower stratosphere. In the future climate, the precipitation becomes stronger and more sporadic with a shorter period. However, precipitation with relatively slow zonal phase velocity does not change significantly,
especially in the $2 \leq C_x \leq 20$ m s$^{-1}$ and $-30 \leq C_x \leq -10$ m s$^{-1}$ ranges. These characteristics are also seen in the spectra of moisture heating (not shown). Differences in the precipitation spectra cause the differences in $F^{(z)}$ in the lower stratosphere.

6. Summary and concluding remarks

Because of the difficulties in the spontaneous simulation of the QBO with commonly used climate models, the changes in the QBO in a future climate with increased CO$_2$ and presumably higher tropical SSTs have not been understood sufficiently. This is the first study to investigate how the QBO changes in a double CO$_2$ climate using a climate model that simulates the QBO by model-resolved waves only. The present climate is characterized by a uniform CO$_2$ mixing ratio of 345 ppmv and climatological SST and sea ice data. The assumed global warming climate is characterized by a doubled CO$_2$ mixing ratio of 690 ppmv. SST and sea ice conditions are those in the present climate plus differences derived from atmosphere–ocean coupled CMIP3 multimodel data.

In the future climate, the period of the QBO becomes longer by about 1–5 months, and the amplitude becomes smaller, especially in the lower stratosphere, despite the stronger mean precipitation in the equatorial region. The mechanisms of the future QBO changes are as follows:

1) A warming troposphere and cooling stratosphere cause stronger westerlies in the midlatitudes and an upward/equatorward shift of the 0 m s$^{-1}$ zonal wind line.
2) Westward wave forcing due to large-scale waves (mostly due to midlatitude Rossby waves) shifts equatorward. Forcing due to parameterized mountain gravity waves shifts upward at midlatitudes.
3) The residual vertical velocity increases 30%–40% in the equatorial lower stratosphere due to strengthened westward wave forcing at midlatitude stratosphere in both hemispheres. The vertical zonal wind shear becomes weaker in the equatorial region, especially in the lower stratosphere.
4) Increased equatorial moisture heating (convective heating plus large-scale condensation heating) variance generates more waves. Wave momentum fluxes in the equatorial lower stratosphere increase by 10%–15%.
5) However, the wave momentum fluxes with $2 \leq C_x \leq 20$ m s$^{-1}$ and $-30 \leq C_x \leq -15$ m s$^{-1}$, whose spectral ranges are relevant to the QBO forcing, do not increase much.
6) As a consequence, the QBO changes in the future climate are obvious in the equatorial lower stratosphere. The period, amplitude, and lowermost levels of the QBO in the future climate become respectively longer, smaller, and higher.

The projected response in our model is clear and we have explained how it arises within the context of the model dynamics. If we assume the projection for QBO period change is correct (say, ~3-month lengthening of the mean period over a century), then it may take several decades more of observations to yield a clear confirmation of the predicted QBO period changes, given the other
cycle-to-cycle variations that appear in the observed record (e.g., Hamilton 2002).

The results of the present study stand in contrast to those of GD, who found quite substantial reductions in QBO period in a doubled CO2 climate, particularly when the nonstationary gravity wave sources were assumed to increase in the warmer climate. Because GD had to make assumptions about the change in parameterized gravity wave sources, the conclusions of that study depend on that additional assumption. This study, on the other hand, removes this assumption. GD noted that the period decrease in their model was due to a weakening of the mean tropical upwelling in the stratosphere as well as to increased parameterized gravity wave fluxes. The GD results are extremely unusual among existing model studies in finding the tropical upwelling weakening in response to greenhouse gas–induced global warming.

S. Watanabe and Y. Kawatani (2011, unpublished manuscript) investigated possible future changes in QBO behavior using the T42L80 MIROC Earth System Model (ESM) with the Hines gravity wave parameterization. In their experiments with unchanged parameterized gravity wave sources, the period of the QBO in an increased CO2 climate become longer because of enhanced mean tropical upwelling.

Kawatani et al. (2010a) used a higher-resolution AGCM (T213 with 300-m vertical resolution), integrated for 3 yr, and obtained a QBO with more realistic amplitude than that in the present T106 AGCM. The T106 model with 550-m vertical resolution must miss some spectral ranges of gravity waves relevant to the QBO. However, we do not have enough observations to speculate how much of the wave momentum flux spectrum is missing at T106 or any other horizontal truncation. No observational instrument can cover the entire gravity wave spectral ranges of interest (e.g., Alexander et al. 2010 and references therein).

Certainly the results presented here will be model dependent, but we can have confidence that some aspects of our findings should be robust. How the QBO changes in response to climate forcing will depend on the changes in equatorial mean upwelling and in the wave fluxes entering the equatorial lower stratosphere. In our model the strengthening of the tropical residual mean upwelling is driven mainly by changes in the subtropical EP flux divergence due to large-scale waves (1 ≤ \( s \) ≤ 11) and parameterized topographic waves whose effects may not depend strongly on model resolution or other details, and these aspects may be expected to be reasonably simulated in a global GCM. We also found that the fluxes associated with waves with the range of phase speeds that have the strongest influence on the QBO do not change much in the future climate (Fig. 12c). If this holds at smaller wavelengths, then the results for the change in QBO behavior may be reasonably independent of model horizontal resolution.

The wave flux change in vertically propagating waves entering the equatorial stratosphere will depend to some extent on (still uncertain) convection parameterizations and on model resolution. The flux changes were quite large for the very high phase speed waves (Fig. 12c) that correspond to a precipitation field in the warm climate that is heavier and more sporadic than in the present climate (Fig. 13c). Whether this aspect of the climate projection can be considered robust is a subject for further investigation, although at least the tendency for more short-term rainfall extremes in a warmer climate is a feature seen in other GCM studies and may result from a robust mechanism (e.g., O’Gorman and Schneider 2009a,b and references therein).

The large increase in high-frequency wave flux predicted by the present model in the warm climate should have important impacts on the mean circulation and tides in the mesosphere and lower thermosphere. This is an aspect that we hope to examine with a version of the model that has an appropriate deep vertical domain. Future progress may be made through higher-resolution global simulations, enhanced observations of vertically propagating waves, and development of a better basic understanding of what controls the space–time variability of latent heat release in the troposphere. The recent work of O’Gorman and Schneider (2009a,b) may open a promising line of research in terms of understanding the large-scale controls over precipitation variability.

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