The Effects of Changes in Sea Surface Temperature and CO₂ Concentration on the Quasi-Biennial Oscillation

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ABSTRACT

The effects of sea surface temperature (SST) and CO₂ on future changes in the quasi-biennial oscillation (QBO) are investigated using a climate model that simulates the QBO without parameterized nonstationary gravity wave forcing. Idealized model experiments using the future SST with the present CO₂ (FS run) and the present SST with the future CO₂ (FC run) are conducted, as are experiments using the present SST with the present CO₂ (present run) and the future SST with the future CO₂ (future run). When compared with the present run, precipitation increases around the equatorial region in the FS run and decreases in the FC run, resulting in increased and decreased wave momentum fluxes, respectively. In the midlatitude lower stratosphere, westward (eastward) wave-forcing anomalies form in the FS (FC) run. In the middle stratosphere off the equator, westward wave-forcing anomalies form in both the FS and FC runs. Corresponding to these wave-forcing anomalies, the residual vertical velocity significantly increases in the lower stratosphere in the FS run but decreases to below 70 hPa in the FC run, whereas residual upward circulation anomalies form in both the FS and FC runs in the middle equatorial stratosphere. Consequently, the amplitude of the QBO becomes smaller in the lower stratosphere, and the period of the QBO becomes longer by about 1–3 months in the FS run. On the other hand, in the FC run, the QBO extends farther downward into the lowermost stratosphere, and the period becomes longer by 1 month.

1. Introduction

The quasi-biennial oscillation (QBO) is a quasi-periodic oscillation in the equatorial stratosphere. The QBO has significant remote effects ranging from the North Pole to the South Pole and from the lower stratosphere to the upper mesosphere (Baldwin et al. 2001). Projecting how the QBO will change in response to increased greenhouse warming or other external perturbations requires self-consistent, comprehensive models that can simulate the changes in wave fluxes and mean vertical motion. 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. Because of the difficulties in spontaneous simulation of the QBO in commonly used climate models, changes in the QBO in a future climate with increased CO₂ and presumably higher tropical SSTs were not sufficiently clarified.

Kawatani et al. (2011) investigated the effects of anticipated twenty-first-century global climate changes on the QBO using a high-resolution version of the Model for Interdisciplinary Research on Climate (MIROC) atmospheric general circulation model (AGCM), which simulates the QBO by model-resolved waves only. They ran a long control integration of the model with observed climatological sea surface temperatures (SSTs) appropriate for the late twentieth century and a CO₂ mixing

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ratio of 345 ppmv. They then ran another integration with doubled CO$_2$ concentration and increased SSTS, incremented by the projected twenty-first-century warming in a multimodel ensemble [phase 3 of the Coupled Model Intercomparison Project (CMIP3)] of coupled ocean–atmosphere runs, forced by the future atmospheric composition for scenario A1B of the Special Report on Emissions Scenarios. In the experiment for late-twenty-first-century conditions, the QBO period becomes longer and the QBO amplitude becomes weaker than those in the late twentieth-century simulation. The downward extension of the QBO into the lowermost stratosphere is also curtailed in the late-twenty-first-century run. These changes are driven by a significant increase in 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 strengthen overall by approximately 10%–15% in the warm simulation, but the increases are almost entirely in zonal phase speed ranges, with little effect on the stratospheric QBO.

Simple vertical one-dimensional, radiative–convective equilibrium model experiments (e.g., Takata and Noda 1997) have shown that the change in thermal vertical profile due to the increase in CO$_2$ is mainly determined by warmed SST in the troposphere and by CO$_2$-induced longwave cooling in the stratosphere. Separation of the roles of increased CO$_2$ and a warmer ocean surface is not possible in standard climate change simulations. Idealized experiments are needed. Sugi and Yoshimura (2004) investigated the separate effects of CO$_2$ and SST increase to understand the mechanisms of atmospheric temperature and precipitation changes. They conducted several idealized AGCM experiments using different SST but the same CO$_2$ concentration and using the same SST but different CO$_2$ concentration. Their results showed that increasing CO$_2$ reduces radiative cooling in the troposphere because of the overlapping effect between water vapor and CO$_2$ bands at 15 $\mu$m, leading to a reduction in tropical precipitation, whereas increasing SST increases the atmospheric temperature and water vapor, leading to increases in radiative cooling and tropical precipitation.

Kodama et al. (2007) investigated the separate roles of increased CO$_2$ and higher SST in changes in the stratospheric Brewer–Dobson circulation. They conducted two idealized AGCM experiments involving 2 $\times$ CO$_2$ with the present SST and 1 $\times$ CO$_2$ with the future SST. They found that both increased CO$_2$ and SST enhance the Brewer–Dobson circulation during winter, whereas CO$_2$ effects mainly enhance the circulation in the summer stratosphere.

Sigmond et al. (2004) used an AGCM to investigate the separate climate effects of doubling middle-atmospheric and tropospheric CO$_2$. In their study, SST changes due to the tropospheric or stratospheric CO$_2$ doubling were calculated approximately from radiative forcing. They reported that of the increased residual circulation, approximately two-thirds can be attributed to the tropospheric CO$_2$ doubling and about one-third to the middle atmospheric CO$_2$ doubling. The increase in the tropospheric Northern Hemisphere midlatitude westerlies can also be attributed mainly to the middle-atmospheric CO$_2$ doubling, demonstrating the importance of the middle-atmospheric CO$_2$ doubling in the tropospheric climate change.

The QBO is a phenomenon in the stratosphere but is driven by wave forcing propagating from the troposphere. The future QBO changes studied by Kawatani et al. (2011) included the effects of both increased CO$_2$ concentration and higher SST. It would be interesting to investigate how the changes in the QBO found by Kawatani et al. (2011) are individually affected by SST change and CO$_2$ increase. As we will show, studying these two effects separately also allows a clear demonstration of the role of mean upwelling in determining the period and vertical structure of the QBO, an issue that has been addressed earlier in mechanistic model studies (e.g., Saravanan 1990). The main purpose of this study is to clarify the individual effects of CO$_2$ and SST on changes in the QBO associated with differences in several physical quantities such as precipitation, mean zonal wind, residual mean circulation, and wave forcing. This paper thus combines and further extends the work of Kawatani et al. (2011) and is arranged as follows. Section 2 describes the model and experimental design. Section 3 describes the QBO changes. Section 4 investigates changes in basic fields. Section 5 examines changes in the wave forcing and residual circulation. Section 6 summarizes the study and provides concluding remarks.

2. Model description and experimental design

We used the same model as used by Kawatani et al. (2011). The model has a horizontal resolution of T106 spectral truncation. Seventy-two vertical layers (L72) with vertical resolution of 550 m from approximately 300 to 5 hPa are used to well represent wave–mean flow interaction. This experiment includes the mountain-induced gravity wave parameterization by McFarlane (1987) but does not include nonstationary gravity wave parameterization. Hence, the simulated QBO is driven by explicitly resolved waves in the model. Some spectral ranges of gravity waves relevant to the QBO are likely to
be missing from the model [see section 6 of Kawatani et al. (2011) for discussion of possible resolution dependence of the future QBO changes].

Kawatani et al. (2011) conducted two types of runs with 90-yr integration: 1) the control run (present climate run) with present SST and sea ice and present CO$_2$ and 2) future SST and sea ice and future CO$_2$ (future climate run). We conducted further two types of runs: 3) future SST and sea ice and present CO$_2$ (FS run) and 4) present SST and sea ice and future CO$_2$ (FC run). In this paper, data from all runs are analyzed, and “differences” are defined by subtracting the present run from each of the future, FS, and FC runs. More details of the model are provided in section 2 of Kawatani et al. (2011).

3. Changes in the QBO

Figure 1 shows a time–height cross section of the monthly and zonal-mean zonal wind over the equator in the present, future, FS, and FC runs for 45 yr (results for the last 45 yr are shown here). In the present climate, an obvious QBO-like oscillation with a period of approximately 2 yr is apparent (Fig. 1a). The simulated period is slightly shorter than that in the real atmosphere ($\sim$28 months on average). The maximum speeds of the easterlies are approximately $-25$ m s$^{-1}$, and those of the westerlies are $15$ m s$^{-1}$. In contrast, Naujokat (1986) reported maximum speeds of $-35$ and $20$ m s$^{-1}$ for easterly and westerly winds, respectively. Thus, the
simulated oscillation has somewhat weaker amplitude but the same east–west phase asymmetry as seen in the observations. The QBO wind variations extend down to approximately 60–80 hPa despite the smaller amplitude in the lower stratosphere compared with that in the real atmosphere. This is because gravity waves with horizontal wavenumbers larger than 107, which are unresolved by T106 resolution, are important for the QBO in the lower stratosphere (Kawatani et al. 2010a). The downward propagation of the westerly shear zones of the zonal wind ($\partial \bar{u} / \partial z > 0$, where $z$ is the altitude) is faster than the downward propagation of the easterly shear zones ($\partial \bar{u} / \partial z < 0$), which agrees with observations.

In the FS run, the 0 m s$^{-1}$ lines of the westerly phase of the QBO extend down to approximately 50 hPa (Fig. 1c), whereas they extend down to approximately 70 hPa in the present run. The amplitude of the QBO becomes smaller, especially in the lower stratosphere. On the other hand, in the FC run (Fig. 1d), the westerly phases of the QBO extend down to approximately 70 hPa and sometimes to approximately 100 hPa (e.g., years 59, 72, and 74). Zonal wind variation associated with the QBO component around 70–100 hPa is slightly larger than that in the present run (not shown). In the future climate, the 0 m s$^{-1}$ lines of the westerly phase of the QBO extend down to approximately 50 hPa and sometimes to 20–30 hPa (Fig. 1b). The QBO in the future climate becomes more irregular.

Figure 2 shows the frequency power spectra of the zonal-mean zonal wind at the equator as a function of height in the (a) present, (b) future, (c) FS, and (d) FC runs. The shaded contours are 1, 2, 4, 8, 16, 32, 64, and 128 m$^2$ s$^{-2}$ month$^{-1}$.
4. Changes in the basic fields

In this section, the climatological differences with a statistical significance of greater than 95% are discussed. The statistical significance is based on a two-sided Student’s t test of the 90 individual-year mean data for each of the present, future, FS, and FC runs. First, we investigated whether the total effects of the future could...
be separated into FS and FC effects. To verify the adequacy of the separation, we investigated the nonlinearity defined below (cf. Kodama et al. 2007):

Nonlinearity = future anomalies – (FS anomalies + FC anomalies).

The nonlinearities of physical quantities such as the zonal-mean temperature, zonal-mean zonal wind, precipitation, and 2-m temperature were negligible compared with the changes. Therefore, we could assess changes due to the future effects by separating the FS and FC effects. In the following, we mainly present results of the FS and FC runs.

Figures 3a–f show the global distribution of differences in the 2-m temperature and precipitation. Differences with statistical significance greater than 95% are colored. In the present climate, the model reasonably simulates the spatial distribution of basic fields, including such features as the separation between the intertropical convergence zone (ITCZ) and South Pacific convergence zone (SPCZ) and the strong precipitation over Africa and South America in the precipitation field [not shown; see Fig. 2a of Kawatani et al. (2011)].

In the FS run, the global distribution of 2-m temperature differences is characterized by larger warming over the continents than over the ocean and strongly enhanced warming at high latitudes especially in the Northern Hemisphere (Fig. 3c). These features are due to the differences in surface heat balance between ocean and land and to asymmetric changes in SST and sea ice between the Northern and Southern Hemispheres. Precipitation increases significantly over the equator, whereas it generally decreases in the subtropical zones (i.e., descending branches of the Hadley circulation) in both hemispheres (Fig. 3d). These responses are commonly found in global warming experiments by CMIP3 models (Solomon et al. 2007).

In the FC run, 2-m temperature slightly increases over the land because of the CO2-induced greenhouse effect but results in nearly no changes over the ocean because SSTs are prescribed at the present value (Fig. 3e). Precipitation significantly decreases over the ocean but generally increases over land (Fig. 3f). This response may be partly attributable to the increase in water vapor transport from ocean to land by the continental-scale land–sea breeze induced by the differential surface warming. Additionally, the precipitation decrease over the ocean can be explained by the change in heat balance in the troposphere (e.g., Sugi and Yoshimura 2004). When the concentration of CO2 is increased, longwave radiative cooling due to CO2 plus H2O is reduced in the troposphere because of the overlapping effect between water vapor and CO2 bands at 15 μm. Convective heating becomes smaller to balance the radiative cooling. As a result, precipitation decreases. Held and Soden (2006) analyzed models employed in the IPCC AR4 and showed that the global mean precipitation increased by 2% K\(^{-1}\), which is much weaker than Clausius–Clapeyron scaling (i.e., 7.5% K\(^{-1}\)). Figure 2d of Held and Soden (2006) implies that precipitation decreases under increasing CO2 without temperature changes (i.e., nearly equivalent to the FC run over the ocean).

Figures 3g and 3h show latitudinal variations in the zonal-mean 2-m temperature and precipitation anomalies. In the future run, the hemispheric asymmetry is evident in 2-m temperature at high latitudes. The global mean 2-m temperature anomaly is about +2.5 K, which is in the range of projections by the CMIP3 models. Zonal-mean precipitation at 10°S–10°N increases by 0.54 mm day\(^{-1}\), similar to the 0.43 mm day\(^{-1}\) in the CMIP3 multimodel mean. In the FS run, zonal-mean precipitation at 10°S–10°N average increases, whereas that in the FC run slightly decreases. Zonal-mean 2-m temperature and precipitation changes in the future are attributable to the SST effect rather than the CO2 effect in most latitudes.

Figure 4 shows the latitude–height cross sections of the climatological annual and zonal-mean temperature, zonal wind, and squared buoyancy frequency \(N^2\) anomalies in the future, FS, and FC runs. The differences with a statistical significance greater than 95% are colored. The black solid contour lines correspond to the present run, and the green dashed lines correspond to the future, FS, and FC runs. The tropopause, which is defined as the 2.5 \(\times\) 10\(^{-4}\) s\(^{-1}\) line of \(N^2\), is shown in the right panels.

In the future run, warming in the troposphere and cooling in the stratosphere are evident (Fig. 4a). Warming in the troposphere is mainly due to the change in SST (Fig. 4d), whereas cooling in the stratosphere is due to the increased CO2 (Fig. 4g). In the FS run, large warming is seen in the middle to upper troposphere, and small cooling exists in the tropical lower stratosphere. A “hornlike” warming structure extends from the subtropical troposphere to the polar stratosphere; this structure was also seen in a model experiment performed by Kodama et al. (2007). The hornlike structure is closely related to changes in residual vertical velocity, as described in the next section. The latitudinal gradient of the temperature differences is large at 50–300 hPa, and the upper parts of the subtropical jets strengthen, which is consistent with the thermal wind balance (Fig. 4e). The upward displacement of a zero-wind line is obvious in the extratropics, and the zero-wind lines shift equatorward from the upper troposphere to the
stratosphere (see black solid lines for the present run and green dashed lines for the FS run).

In the FC run, the maximum-warming anomaly is found around the tropical tropopause region where the temperature is the coldest so that CO$_2$ longwave radiation heats by absorbing longwave radiation (Fig. 4g). This warming anomaly extends down to the middle to lower troposphere around 60°N and 50°S. Corresponding to this warm anomaly, the zonal wind anomalies are easterly from 40°S to 40°N and westerly around 55°S and 65°N (Fig. 4h). On the other hand, changes in zero-wind lines in the upper-troposphere and lower-stratosphere (UTLS) regions are not clear.

Warm anomalies extending down to the troposphere in the midlatitudes were also simulated under the same condition as the present SST with future CO$_2$ by Kodama et al. (2007). Sigmond et al. (2004) conducted an experiment with 2×CO$_2$ in the stratosphere and 1×CO$_2$ in the troposphere with the present SST (i.e., the difference between their and our simulations is the CO$_2$
concentration in the troposphere). In that experiment, the tropopause warms, and the warming areas extend downward to the midlatitude troposphere, resulting in easterly anomalies around 30°N and westerly anomalies around 60°N in the December–February (DJF) mean.

The stability changes associated with temperature anomalies. In the FS run, the stability increases in the troposphere by a maximum of 20% at 200–300 hPa, decreases in the UTLS region by a minimum of 25% at 100 hPa, and increases by 5% around 55 hPa in the equatorial region (Fig. 4f). In the FC run, the stability increases around the tropopause region by a maximum of 15% at 110 hPa and decreases in the stratosphere by a minimum of 12% around 60 hPa (Fig. 4i). The altitude of the tropical tropopause becomes higher in the FS run but lower in the FC run (solid black line for the present run and green dashed lines for the FS and FC runs).

The results shown in this section reveal very different changes in the basic field between the FS and FC runs. Here, changes important for the next section are summarized: 1) zonal-mean precipitation in the equatorial region increases in the FS run but slightly decreases in the FC run; 2) in the FS run, zero-wind lines of the zonal wind shift upward in the extratropics and equatorward from the upper troposphere to the stratosphere, whereas they do not change much in the FC run; and 3) the zonal wind anomalies around 40°S–40°N in the UTLS region are westerly in the FS run but easterly in the FC run.

5. Changes in wave forcing and the residual circulation

a. Changes in residual circulation

Several studies have predicted that Brewer–Dobson circulations, including equatorial mean residual upwelling, will be stronger in the future climate (Butchart and Scaife 2001; Butchart et al. 2006; Garcia and Randel 2008; McLandress and Shepherd 2009; Okamoto et al. 2011). Kawatani et al. (2011) showed that future changes in the QBO are strongly related to strengthened equatorial upwelling. In this section, we discuss how and why the simulated residual circulation changes in the FS and FC runs.

Changes in tropical upwelling are related to wave forcing off the equator. Zhou et al. (2006) used a simple model based on the balanced transformed Eulerian mean equations to investigate the factors controlling tropical upwelling. They indicated that wave-driven circulation plays the crucial role in determining the annual mean tropical upwelling. Westward wave forcing off the equator strengthens tropical upwelling more effectively if it penetrates near to the equator. The position and strength of wave-forcing anomalies are the key factors that determine changes in the tropical upwelling.

To investigate changes in wave forcing and residual circulation quantitatively, the Eliassen–Palm flux (EP flux) and the mean residual circulation in the transformed Eulerian mean (TEM) formation are calculated as follows (Andrews et al. 1987):

\[ F^{(\phi)} = \rho_0 a \cos \phi \left( u \frac{\partial \bar{\theta}}{\partial z} - u \bar{\theta} \right), \]

\[ F^{(z)} = \rho_0 a \cos \phi \left[ f - (a \cos \phi)^{-1} (\bar{\bar{u}} \cos \phi \bar{\phi}) \right] \times \bar{\theta} / \bar{\theta}, \]

\[ \nabla \cdot F = (a \cos \phi)^{-1} \partial / \partial \phi (F^{(\phi)} \cos \phi) + \partial F^{(z)}/\partial z, \]

with the zonally averaged momentum equation expressed as

\[ \Pi = \bar{\bar{v}} [ f - (a \cos \phi)^{-1} (\bar{\bar{u}} \cos \phi \bar{\phi}) ] - \bar{\bar{v}} \Pi_z \]

\[ + (\rho_0 a \cos \phi)^{-1} \nabla \cdot F + \bar{X}. \]

In the above equations, \( \rho_0, a, \phi, z, u, v, w, \theta, \) and \( f \) are 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 Coriolis parameter (\( f = 2\Omega \sin \phi \), where \( \Omega \) is the rotation rate of the Earth), respectively. The subscripts \( \phi, z, \) and \( r \) denote the meridional, vertical, and time derivatives, respectively. The mean residual circulations of the meridional and vertical components are expressed by \( \Pi^r \) and \( \bar{\bar{v}}^r \). Eastward and westward wave forcing correspond to the EP-flux divergence and convergence (i.e., \( \nabla \cdot F > 0 \) and \( \nabla \cdot F < 0 \), respectively. Here \( \bar{X} \) means unresolved forcing in the model. Most of \( \bar{X} \) corresponds to forcing due to parameterized mountain gravity waves (not shown). The residual vertical velocity \( \bar{\bar{v}}^r \) is calculated as follows:

\[ \bar{\bar{v}}^r = \bar{v} + (a \cos \phi)^{-1} (\cos \phi \bar{\bar{v}} \cos \phi \bar{\phi}) \phi. \]

The latitude–height cross section of the climatological annual mean EP-flux divergence plus zonal forcing due to the mountain gravity wave parameterization (i.e., total zonal wave forcing) and that of residual streamfunction in the present run are shown in Figs. 5a and 5b. Climatological annual and zonal-mean zonal wind is contoured. Positive and negative values of wave forcing correspond to eastward and westward wave forcing, and positive and negative values of the streamfunction indicate clockwise and anticlockwise circulations, respectively. In a steady
FIG. 5. (a),(b) Climatological annual mean (a) EP-flux divergence plus zonal forcing due to mountain gravity wave parameterization and (b) residual streamfunction in the present climate run. Zonal wind is contoured with intervals of 10 m s\(^{-1}\). (c)–(h) Anomalies of (c),(e),(g) zonal wave forcing and (d),(f),(h) residual streamfunction in the (c),(d) future, (e),(f) FS, and (g),(h) FC runs. Differences with statistical significances of ≥95% are colored. The color contours are at (a) 0, ±0.2, ±0.4, ±0.8, ±1.2, ±1.6, ±2.0, and ±2.4 m s\(^{-1}\) day\(^{-1}\) and (b) 0, ±5, ±10, ±20, ±30, ±50, ±100, ±200, and ±400 kg m\(^{-1}\) s\(^{-1}\). The color contours of (c),(e),(g) and (d),(f),(h) are 0.25 times and 0.1 times as much as those in (a) and (b), respectively.
FIG. 6. (a),(b) Climatological annual mean (a) EP flux divided by the atmospheric density (vector) and EP flux divergence (color), and (b) zonal forcing due to mountain gravity wave parameterization in the present climate run. Zonal wind is contoured with an interval of 10 m s$^{-1}$. (c)–(h) Anomalies of (c),(e),(g) the EP flux, its divergence, and (d),(f),(h) parameterized wave forcing in the (c),(d) future, (e),(f) FS, and (g),(h) FC runs. Differences with statistical significances of ≥95% are colored. The vertical components of the EP flux are multiplied by a factor of 660. Color contours are at 0, ±0.2, ±0.4, ±0.8, ±1.2, ±1.6, ±2.0, and ±2.4 m s$^{-1}$ day$^{-1}$ for (a),(b). The arrow units are $2.4 \times 10^8$ m$^3$ s$^{-1}$ for (a). The color contours and arrow units in (c)–(h) are 0.25 times larger than those in (a) and (b).
state in the zonal-mean field, by the principle of down
ward control, westward wave forcing results in poleward
flow, whereas eastward forcing induces equatorward
flow (Haynes et al. 1991). Large westward wave forcing
distributes within the midlatitude westerly jets in both
hemispheres (Fig. 5a), and poleward residual mean flows
are formed from the low to high latitudes (Fig. 5b). As
a result, residual vertical velocity in the equatorial region
is upward.

As described here, mean tropical upwelling is closely
related to the distribution of wave forcing off the equator.
Next, the residual circulation changes associated with
changes in wave forcing are investigated. Anomalies of
total zonal wave forcing and mean residual circulation
in each run are shown in Figs. 5c–h. In the FS run, the
large westward forcing anomalies exist at 10°–80°N and
5°–50°S around 30–80 hPa (Fig. 5e). Corresponding to
these westward wave-forcing anomalies, the residual
mean circulation anomalies are upward and generally
poleward in the equatorial stratosphere (Fig. 5f).

In the FC run, distributions of wave-forcing anom-
aliies in the lower stratosphere and off the equator are
quite different (Fig. 5g). Eastward wave-forcing anom-
ailies spread around 60–100 hPa in both hemispheres,
resulting in the equatorward and downward circulation
anomalies below 70 hPa near the equator (Fig. 5h). On
the other hand, westward forcing anomalies are distrib-
uted off the equator from 10 to 60 hPa in the Northern
Hemisphere. In the Southern Hemisphere, statistically
significant westward forcing anomalies are distributed
around 20 hPa with relatively small values. Correspond-
ing to these westward forcing anomalies, the residual
mean circulation anomalies in the equatorial region are
upward above 70 hPa.

Next, we investigate the relative roles of resolved and
parameterized waves in the total wave forcing to indi-
vidually clarify their contribution to the residual circu-
lation changes. First, we confirmed the characteristics
of the EP flux and its divergence and parameterized wave
forcing in the present run, as shown in Figs. 6a and 6b.
Resolved waves with westward momentum around
30°–50°N and 30°–50°S propagate upward (i.e., EP-flux
vectors directed upward) and equatorward (Fig. 6a).
Equatorward propagations become very weak around
0 m s\(^{-1}\) lines of the zonal wind. Around 60°S and 60°N,
resolved waves propagate directly upward toward the
upper stratosphere. Westward forcing by resolved waves
distributes in association with the westerly jet. On the
other hand, large westward wave forcing due to mountain
gravity wave parameterization exists in the midlatitude
UTLS region in both hemispheres where the westerlies
are weak. This forcing plays a substantial role in de-
celerating the upper part of the subtropical westerly jet.

Figures 6c–h show anomalies of the EP flux and its
divergence and parameterized wave forcing in each run.
In the FS run, resolved waves show large westward
forcing anomalies at 10°–80°N and 5°–50°S around 30–
100 hPa (Fig. 6c). The EP-flux anomalies indicate that
more waves with westward momentum preferentially
propagate into the lower stratosphere. Equatorward-
propagating waves from the troposphere into the lower
stratosphere are apparent at 25°–50°N and 30°–60°S. These
anomalies are mainly due to waves with 1 ≤ s ≤ 11 (not
shown; s is zonal wavenumber); more Rossby waves
propagate equatorward and upward due to the equator-
ward and upward shifts of zero-wind lines. As a result,
westward wave forcing shifts upward and equatorward,
and westward forcing anomalies lie above the eastward
wave-forcing anomalies in the UTLS region.

Westward forcing anomalies due to the mountain
gravity wave parameterization occur in the midlatitudes
around 30–50 hPa in both the Northern and Southern
Hemispheres, whereas eastward forcing anomalies are
seen below the westward forcing anomalies (Fig. 6f).
Absolute values of the upper part of this eastward forc-
ing anomalies around 70 hPa are smaller than those of
westward forcing anomalies due to the resolved waves.
As a result, the total wave-forcing anomalies around
70 hPa is westward forcing (Fig. 5e). As discussed by
Kawatani et al. (2011), shifting the zero line of the zonal
wind anomaly causes more upward propagation of the
parameterized mountain gravity wave; westward wave-
forcing anomalies are then formed at a higher altitude,
whereas eastward wave-forcing anomalies are formed at
a lower altitude.

In the FC run, both resolved and parameterized waves
show eastward forcing anomalies off the equator in the
UTLS region (Figs. 6g,h). Resolved wave anomalies
propagating equatorward from the midlatitudes are not
seen, but upward directed EP-flux anomalies are domi-
nant around 30°–70°N (Fig. 6g). Some Rossby waves
are prevented from propagating into the equatorial re-
gion by easterly anomalies around 40°S–40°N (Fig. 4h),
and these waves preferentially propagate upward. Less
equatorward-propagating Rossby waves cause eastward
wave-forcing anomalies off the equator.

Eastward forcing anomalies due to parameterized waves
in the UTLS region could be explained by weakened
wave sources rather than by background zonal wind
changes. Large eastward forcing anomalies are formed
around 80 hPa at 35°N (Fig. 6h), where large parameter-
ized westward forcing exists in the present run (Fig. 6b).
Zero-wind lines of the zonal-mean zonal wind are similar
between the present and FC runs (Fig. 4h), and distri-
butions of parameterized wave forcing are also similar
(not shown). However, the absolute values of the forcing
FIG. 7. (a) Latitude–height cross section of the climatological annual mean residual vertical velocity $\mathbf{w}^*$ in the present run, and (b)–(d) $\mathbf{w}^*$ anomalies in the (b) future, (c) FS, and (d) FC runs. (e) Profiles of climatological $\mathbf{w}^*$ in the present (black), future (red), FS (blue), and FC (green) runs. (f) Ratio of the future (red), FS (blue), and FC (green) runs to the present $\mathbf{w}^*$ at 20°S–20°N.
in the FC run are smaller than those in the present run. These results indicate that eastward wave-forcing anomalies around 80 hPa are due not to vertical shifting of the forcing profile but to smaller wave generations. Surface zonal winds are smaller in the midlatitudes associated with the warming 2-m temperature around 50°–60°N (Fig. 3e). Weakening of surface westerly winds reduces the generation of mountain gravity waves.

To quantitatively investigate the mean ascent in the equatorial lower stratosphere, $w^*$ is examined. Figure 7a shows the latitude–height cross section of the climatological annual mean $w^*$ in the present run, and Figs. 7b–d show the $w^*$ anomalies in each run. In the present run, the annual mean upward residual velocity has a narrow meridional width from about 10°S to 10°–15°N at 200 hPa. The meridional widths expand broadly in the latitudinal direction as the height increases from the upper troposphere to the stratosphere. In the FS run, the meridional widths of positive $w^*$ anomalies in the equatorial region are narrow at 200 hPa and expand widely as the height increases (Fig. 7c; see Fig. 5f for residual streamfunction anomalies). On the other hand, in the FC run, negative $w^*$ anomalies are distributed in the equatorial region around 70–150 hPa, and positive anomalies exist over 70 hPa (Fig. 7d), as shown in Fig. 5h.

In the FS run, the negative $w^*$ anomalies in the mid- to high-latitude stratosphere (Fig. 7c) coincide with a horn-like warming temperature structure (Fig. 4d). Downward vertical motion induces adiabatic heating and thus positive temperature anomalies. Zonal wind changes associated with temperature anomalies alter properties of wave propagation and the residual circulation. Changes in these physical quantities are closely linked with each other.

Figures 7e and 7f show the vertical profiles of $w^*$ at 20°S–20°N in each run and those for the ratios of the future, FS, and FC $w^*$ to the present $w^*$. Our model’s present climate equatorial residual upwelling velocity profile compares well with various observations as mentioned by Kawatani et al. (2011). In the FS run, $w^*$ is strengthened by about 40% around 70 hPa, by 5% around 10 hPa, and by 23% in the 10–70-hPa average. Much more strengthening is found in the lower stratosphere than in the middle to upper stratosphere. In the
FC run, \( \overline{\mathbf{w}}^* \) is weaker below 70 hPa and stronger above 70 hPa; \( \overline{\mathbf{w}}^* \) is strengthened by about 6.5% around 40 hPa and 4.3% in the 10–70-hPa average.

If wave forcing relevant to the QBO does not change, and \( \overline{\mathbf{w}}^* \) primarily determines the QBO period, the 4.3% increase in \( \overline{\mathbf{w}}^* \) in the FC run lengthens the QBO period (24 months in the present run) by about 1 month, which is consistent with the results shown in Fig. 2d. Concerning the FS run, 23% changes in the average \( \overline{\mathbf{w}}^* \) lengthen the QBO period by about 3 months. The period of the QBO in the FS run becomes 1–3 months longer (Fig. 2c). Some aspects were not explained by the \( \overline{\mathbf{w}}^* \) changes only because \( \overline{\mathbf{w}}^* \) changes in the FS run are highly dependent on the height, which would not be valid for such a rough estimation. Additionally, wave sources must be changes associated with significant precipitation changes around the equator (Fig. 3), which is discussed in the next section.

The increased \( \overline{\mathbf{w}}^* \) above 70 hPa in the FC run should also be related to decreased buoyancy frequency in the stratosphere (Fig. 4i) in addition to the westward forcing anomaly off the equator (Fig. 5g; Zhou et al. 2006). The quantitative separation of the contributions from these two factors requires further sensitivity experiments and is beyond the scope of this study.

b. Changes in wave momentum flux

In this subsection, we analyze the changes in wave momentum flux using hourly sampled data. Here, we discuss the results from year 81 to year 90. The statistical significance of differences is calculated by the two-sided Student’s \( t \) test using 10 individual yearly mean samples. First, the differences in the equatorial moist heating (i.e., cumulus convective heating plus large-scale condensation heating), which is the strongest wave source, are investigated. Mean moisture heating and its variation in the FS run are larger than those in the present run, whereas they are slightly smaller in the FC run (not shown), consistent with precipitation changes (Fig. 3). These results imply that more waves could be generated in the FS run but fewer in the FC run [see Fig. 10 of Kawatani et al. (2011) for heating profiles].

Following Kawatani et al. (2011), we discuss the absolute value of the vertical flux of zonal momentum \(|\overline{u'w'}|\) [\(|\overline{u'w'}| = |\overline{u'}| |\overline{w'}| + |\overline{u'} w'|\); sum of absolute values of positive and negative momentum fluxes] associated with waves due to \( 12 \leq s \leq 106 \). Waves with \( 12 \leq s \leq 106 \) are regarded as internal inertia–gravity waves, which play a significant role in driving the QBO (Kawatani et al. 2010a,b).

Figure 8 shows the latitude–height cross sections for the ratios of the future, FS, and FC runs to the present run for \(|\overline{u'w'}|\). The results in the FS run are similar to those in the future run. In the FS run, \(|\overline{u'w'}|\) is larger above about 170 hPa in the equatorial region, whereas \(|\overline{u'w'}|\) decreases below about 170 hPa (Fig. 8b). These differences result from the upward shift of the vertical profile of \(|\overline{u'w'}|\) in the FS run compared with the present run [see Fig. 11b of Kawatani et al. (2011) for the future run]. At 60–150 hPa in the equatorial region, \(|\overline{u'w'}|\) increases by up to about 16%, which would be related to the strengthened variances in moisture heating. The \(|\overline{u'w'}|\) associated with waves due to \( 1 \leq s \leq 11 \) also increases significantly by 10%–15% in the equatorial region at 60–100 hPa (not shown). On the other hand, in the FC run, \(|\overline{u'w'}|\) slightly decreases in the upper troposphere (Fig. 8c), which would be associated with decreasing precipitation in the equatorial region (Fig. 3). Increasing SST contributes to more wave generation propagating into the stratosphere.

The changes in the zonal wavenumber and frequency spectra of the vertical component of the EP flux and precipitation in the FS run are similar to those in the future run; the wave momentum fluxes with \( 2 \leq C_x \leq 20 \text{ m s}^{-1} \) and \(-30 \leq C_x < -15 \text{ m s}^{-1} \), whose spectral ranges are relevant to the QBO forcing, do not increase much [see details in Figs. 12 and 13 of Kawatani et al. (2011)]. Consequently, the effect of the enhanced mean tropical upwelling in the equatorial stratosphere overwhelms counteracting influences from the strengthened wave fluxes in the FS run, as mentioned by Kawatani et al. (2011), for the future run. On the other hand, in the FC run, the spectra of the EP flux and precipitation are similar to those in the present run, with slightly weaker values. For this reason, the residual circulation changes are conclusive for the QBO changes in the FC run.

6. Summary and concluding remarks

The effects of SST and CO2 on the QBO changes are investigated using a climate model that simulates a fairly realistic QBO with only explicitly resolved nonstationary waves. We analyzed data from four different runs: 1) the control run (present climate run) with the present SST and sea ice and present CO2, 2) future SST and sea ice and future CO2 (future climate run), 3) future SST and sea ice and present CO2 (FS run), and 4) present SST and sea ice and future CO2 (FC run). We conducted detailed analysis of the changes in wave forcing in the lower stratosphere and their relation to changes in mean equatorial upwelling. In this paper, “differences” are defined by subtracting the present run from each of the future, FS, and FC runs. We confirmed that the nonlinearity—found by subtracting anomalies of FS plus FC runs from those of the future run—is nearly zero for fundamental physical values (e.g., zonal-mean temperature) and is thus negligible.
The period of the QBO becomes longer by about 1–3 months in the FS run and 1 month in the FC run. In the FS run, the amplitude of the QBO becomes smaller in the lower stratosphere. On the other hand, in the FC run, the QBO extends farther downward into the lowermost level of the stratosphere. The mechanisms of the QBO changes are as follows:

1) In the FS run, zonal-mean precipitation increases significantly in the equatorial regions. In the FC run, precipitation decreases over the ocean, and zonal-mean equatorial precipitation slightly decreases.

2) Warming in the troposphere is mainly due to the SST change, whereas cooling in the stratosphere is due to the increase in CO$_2$. In the FS run, the westerlies in the midlatitudes become stronger, and there is a clear upward/equatorward shift in the 0 m s$^{-1}$ zonal wind line. In the FC run, the tropopause regions become warmer, and warming anomalies extend down to the middle to lower troposphere in the midlatitudes, which cause easterly anomalies in the 40$^\circ$S–40$^\circ$N region. These zonal wind changes alter the propagation property of resolved waves and parameterized waves.

3) In the FS run, westward wave-forcing anomalies are formed off the equator in the lower stratosphere due to both resolved waves and parameterized waves. On the other hand, in the FC run, eastward wave-forcing anomalies caused by resolved waves and parameterized waves are evident. In the middle to upper stratosphere, westward forcing anomalies are formed off the equator in both the FS and FC runs.

4) In the FS run, the residual vertical velocity increases by 30%–40% in the equatorial lower stratosphere, preventing the QBO from propagating into the lower stratosphere. In the FC run, the residual vertical velocity decreases below 70 hPa, and the QBO extends farther downward into the lower stratosphere. The residual upward velocity increases above 70 hPa.

5) Wave momentum fluxes in the equatorial lower stratosphere increase by 10%–16% in the FS run. However, momentum fluxes relevant to the QBO forcing do not increase much. In the FC run, momentum fluxes slightly decrease.

6) As a consequence, increases in the SST and in CO$_2$ concentration both lengthen the QBO period, but they have different effects on the vertical structure of the QBO.

This study reveals that the QBO will be significantly affected by expected late-century changes in SST, and the direct stratospheric effect of increased CO$_2$ will make a more modest contribution. The fact that the ocean surface warming and the direct effect of the CO$_2$ change produce different responses in the mean equatorial upwelling allowed us to demonstrate the strong effects of the mean upwelling in controlling the QBO behavior in a comprehensive model.

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