Interannual Variations of Stratospheric Water Vapor in MLS Observations and Climate Model Simulations

YOSHIO KAWATANI
Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

JAE N. LEE
Joint Center for Earth Systems Technology, University of Maryland, Baltimore County, Baltimore, Maryland

KEVIN HAMILTON
International Pacific Research Center, University of Hawai'i at Mānoa, Honolulu, Hawaii

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ABSTRACT

By analyzing the almost-decade-long record of water vapor measurements from the Microwave Limb Sounder (MLS) instrument on the NASA Aura satellite and by detailed diagnostic analysis of the results from state-of-the-art climate model simulations, this study confirmed the conceptual picture of the interannual variation in equatorial stratospheric water vapor discussed in earlier papers (e.g., Geller et al.). The interannual anomalies in water vapor are strongly related to the dynamical quasi-biennial oscillation (QBO), and this study presents the first QBO composite of the time–height structure of the equatorial water vapor anomalies. The anomalies display upward propagation below about 10 hPa in a manner analogous to the annual “tape recorder” effect, but at higher levels they show clear downward propagation. This study examined these variations in the Model for Interdisciplinary Research on Climate (MIROC)-AGCM and in four models in phase 5 of the Coupled Model Intercomparison Project (CMIP5) that simulate realistic QBOs. Diagnostic budget analysis of the MIROC-AGCM data and comparisons among the CMIP5 model results demonstrate (i) the importance of temperature anomalies at the tropopause induced by the QBO for lower-stratospheric water vapor variations and (ii) that upper-stratospheric water vapor anomalies are largely driven by advection of the mean vertical gradient of water content by the QBO interannual fluctuations in the vertical wind.

1. Introduction

Although stratospheric water vapor (H$_2$O) mixing ratios are quite small, the water vapor in the stratosphere makes a significant contribution to the global-mean radiative forcing of climate (e.g., Solomon et al. 2010). In addition to its importance as a greenhouse forcing, stratospheric H$_2$O is also of interest as a diagnostic of large-scale stratospheric circulation. In a classic study, Mote et al. (1996) analyzed satellite observations of H$_2$O near the equatorial stratosphere and noted a strong annual cycle of mixing ratio that displayed a slow upward phase propagation. This is consistent with a simple picture of the stratospheric circulation in which the H$_2$O content of the air entering the tropical stratosphere is set by the saturation value for the coldest temperatures experienced as it rises through the very cold tropical tropopause. The annual cycle of temperature near the tropical tropopause results in significant seasonal modulation of the saturation H$_2$O mixing ratio, which is then reflected in the annual cycle in the H$_2$O concentration of air entering the stratosphere. This annual signal is simply advected upward by the slow mean upwelling associated with Brewer–Dobson circulation near the equator, in what is now known as the “tape recorder” effect (Mote et al. 1996).

Mote et al. (1996, 1998) noted that interannual signals in the equatorial stratospheric H$_2$O concentrations could also be seen, and that the deviations from the simple picture of the annual cycle tape recorder could be...
largely explained as effects of the quasi-biennial oscillation (QBO). The QBO is known to modulate the interannual variations in tropical tropopause temperature (Reid and Gage 1985; Randel et al. 1998, 2000, 2004; Zhou et al. 2001). The QBO contribution to temperature variation near the tropopause may be up to 0.5 K (Randel et al. 2000), and these temperature changes could result in significant modulation of the H2O content of the air entering the stratosphere.

Randel et al. (2004) investigated interannual variations of stratospheric H2O during 1992–2003 using Halogen Occultation Experiment (HALOE) data from the National Aeronautics and Space Administration (NASA) Upper Atmospheric Research Satellite (UARS) and found H2O interannual changes of approximately ±0.3 ppmv in magnitude near the equator with a roughly 2-yr periodicity. The anomalies can be traced back to the tropical tropopause and can propagate vertically in a manner similar to the seasonal tape recorder. Fujiwara et al. (2010) investigated H2O variations in the tropical lower stratosphere using balloonborne cryogenic frost-point hygrometer data between 1993 and 2009 during various campaigns. They identified H2O concentration variations that are apparently associated with the QBO in tropopause temperatures. Fujiwara et al. also noted that the vertical gradients of H2O in the westerly shear phase are greater than those in the easterly shear phase and explained this in terms of the advection by the QBO residual meridional circulation.

The QBO influences on stratospheric H2O have also been investigated by numerical simulations, although this avenue of research has been complicated by the fact that most comprehensive atmospheric general circulation models (AGCMs) do not simulate a QBO in the tropical stratosphere. Giorgetta and Bengston (1999) conducted AGCM experiments including a simple assimilation of the observed near-equatorial stratospheric zonal-mean winds, effectively forcing a realistic dynamical QBO in their model. They found evidence of both the QBO variation in the dehydration of air rising through the tropical tropopause and QBO modulation of the ascent rate of tropical air. However, their model is rather incomplete, in that it extended only to 10 hPa and did not include the methane oxidation process that is an important source of H2O in the real stratosphere.

Geller et al. (2002) investigated interannual variations of stratospheric H2O associated with both the QBO and El Niño–Southern Oscillation (ENSO) using a twodimensional (2D) chemistry transport model. They showed that QBO variations in cold-point tropopause temperature play a large role in stratospheric H2O variations, which is consistent with the work of Giorgetta and Bengston (1999). Geller et al. (2002) also showed that the ENSO effect produces significant variations from one OBO cycle to another, and that the model results, including both the QBO and ENSO effects, are improved when judged against HALOE observations of equatorial water vapor.

Earlier observational studies have shown that the simple tape-recorder propagation of interannual signals from the equatorial tropopause only explains the observed water vapor anomalies up to at most 25–30 km (~10–15 hPa) and that the upward propagation is not apparent at higher levels (Randel et al. 1998, 2004; Geller et al. 2002). These earlier studies also included diagnostic and simple model investigations of the mechanisms of interannual variability in the tropical upper-stratospheric water vapor. Randel et al. (1998) analyzed HALOE data from 1991 to 1997 and found that equatorial H2O (and CH4) anomalies over the 35–45-km altitude range are correlated with anomalies in the residual-mean vertical velocity, indicating a role for advection of mean vertical gradients in generating these trace-constituent variations.

Geller et al. (2002) analyzed 6 years of HALOE data up to 50 km and showed that the interannual water vapor anomalies slope upward with time below approximately 35 km but variations above that height show no such slope. They also conducted experiments in a 2D model with and without QBO residual meridional circulation. By subtracting the run with no QBO transport from the run that included the QBO transport variations, Geller et al. (2002) demonstrated that the upper-stratospheric H2O interannual anomalies result from the transport of the water vapor by the QBO-induced anomalies of the residual circulation.

The discussion above indicates that considerable progress has been made in characterizing and explaining interannual variations of equatorial water vapor concentrations. However, the extant observational analyses all had significant limitations in the data available. Also, the simulations that have been previously analyzed have come from either simplified 2D models or from AGCMs that lacked adequate treatment of the upper stratosphere. Our study reported in the present paper has been motivated by the availability of new satellite data and more complete comprehensive numerical simulations. By analyzing a long and (arguably) higher-quality observational record and by detailed analysis of the results from a long run of a comprehensive 3D model, including the methane oxidation process, we have improved the characterization of the QBO in equatorial water vapor. We have also confirmed the earlier understanding of the mechanisms driving the QBO variations and placed it on a more secure footing.
The UARS HALOE observations have been used in several previous studies of stratospheric H$_2$O. The HALOE sampling is approximately 15 sunrise and 15 sunset measurements per day, with sunrises and sunsets usually separated in latitude. It takes about 1 month to sample the latitude range from about 60$^\circ$N to 60$^\circ$S (Russell et al. 1993). The UARS HALOE data for trace-gas concentrations are available from 1991 to 2005, but until 1994 they are contaminated by the aerosol signal from the 1991 Mt. Pinatubo eruption. In their original papers Mote et al. (1996, 1998) analyzed the available HALOE data but also showed that the tape-recorder signal was clear in measurements from the UARS Microwave Limb Sounder (MLS) instrument. In fact, the annual cycle data for H$_2$O appear considerably less noisy for the MLS data than for the HALOE data. Unfortunately, the UARS MLS instrument operated for only 18 months and Mote et al. analyzed the MLS H$_2$O retrievals up to only 6.8 hPa (~35 km).

A new MLS instrument, the Earth Observing System (EOS) Microwave Limb Sounder, is now on board NASA’s Aura satellite, which launched in July 2004 (Waters et al. 2006). This instrument detects thermal microwave emission from the edge of Earth’s atmosphere by viewing forward along the spacecraft flight direction. The view is scanned from the ground to about 90 km approximately every 25 s. The satellite makes about 13 orbits per day and retrieves vertical profiles of atmospheric temperature and composition in the vertical range of 8–90 km (Livesey et al. 2006). EOS MLS H$_2$O data for almost 10 years are now available. In this study, we have investigated the interannual variations in H$_2$O content in the equatorial stratosphere using the long record from EOS MLS version 3.3.

We also conducted climate model simulations using a fine-horizontal-and-vertical-resolution (T106L72) version of the Model for Interdisciplinary Research on Climate (MIROC) AGCM to clarify the mechanism of the observed H$_2$O interannual variability. This model spontaneously simulates a rather realistic dynamical QBO in the tropical stratosphere (Kawatani et al. 2011, 2012). The standard version of this model includes a simple parameterization of the effects of the methane oxidation source of H$_2$O, and we have conducted experiments with and without the methane oxidation parameterization to elucidate the mechanisms of H$_2$O variability in the upper stratosphere. In addition, the interannual variability of H$_2$O is investigated in simulations from several other global models that were included in phase 5 of the Coupled Model Intercomparison Project (CMIP5).

This paper is arranged as follows. Section 2 describes observational data and provides a description of the model. Section 3 analyzes the interannual variation of MLS H$_2$O. Section 4 investigates the interannual variation of equatorial water vapor concentration as simulated in the MIROC model. Section 5 compares the interannual variations of water vapor concentration in the equatorial stratosphere in four CMIP5 global models. Section 6 summarizes the study and provides concluding remarks.

2. Observational data and model description

a. MLS observation

Monthly-mean EOS Aura MLS H$_2$O concentration data from August 2004 to January 2014, derived from latest version 3.3 (v3.3) of daily-mean observations, are analyzed in this study. Extensive assessment has been conducted for MLS v2.2 H$_2$O product through validation studies (Lambert et al. 2007; Read et al. 2007). For MLS v2.2 H$_2$O data, the single-profile precision is about 0.2–0.3 ppmv (4%–9%) in the stratosphere and the accuracy is estimated to be 0.2–0.5 ppmv (4%–11%) for the pressure range 68–0.01 hPa (Lambert et al. 2007). This precision is not achieved in the lower-stratosphere-and-upper-troposphere region with values of 10%–20% from 121 to 82.5 hPa (Read et al. 2007). The MLS v3.3 H$_2$O product is expected to be about 0.2–0.3 ppmv wetter than the v2.2 product in the pressure range 82.5–0.1 hPa (Livesey et al. 2011). For pressures greater than 21 hPa, the precisions of the two versions are nearly identical. The H$_2$O data mapped into a 4$^\circ$ (latitude) × 8$^\circ$ (longitude) grid from 146.8 to 0.46 hPa (29 levels) are analyzed here. The vertical resolution is about 2.5 km at 316–215 hPa, 3.0 km at 100–1.0 hPa, and 3.4 km above 1 hPa.

b. General circulation model

We use a version of the MIROC-AGCM almost identical to that described in Kawatani et al. (2011). This version of the model has a horizontal resolution of T106 spectral truncation that corresponds to a grid interval of approximately 120 km (about 1.125$^\circ$ in latitude and longitude). The model uses 72 vertical numerical levels (L72) with the top boundary at 1.2 hPa (~47 km). The vertical resolution is close to 550 m from about 300 up to 5 hPa, which should provide adequate representation of mean-flow interaction with vertically propagating waves. Starting at 4.5 hPa, the model includes an artificial damping in a “sponge layer.” The topographic gravity wave parameterization of McFarlane (1987) is employed, but no parameterization of nonstationary gravity wave effects is included. Hence, the simulated QBO is driven by explicitly resolved waves in the model. The parameterization of the methane oxidation process used in the European Centre for Medium-Range
Weather Forecasts (ECMWF) is included in the model. Methane oxidation is a primary process of H\textsubscript{2}O production in the middle atmosphere. The chemical source in the water vapor mass mixing ratio tendency equation is expressed as \( k(Q - q) \), where \( k \) is a rate (specified as a function of pressure), \( Q \) is a parameter set at \( 4.25 \times 10^{-6} \) (corresponding to 6.8 ppmv), and \( q \) is the model H\textsubscript{2}O mass mixing ratio [see ECMWF (2013) for more details and references therein].

Our T106 MIROC AGCM integrations reported in this paper were conducted with annually repeating sea surface temperatures (SSTs) based on present-day observed climatology (Kawatani et al. 2011).

We conducted two model integrations: (i) a control run with the methane oxidation parameterization and (ii) another run without the methane oxidation source for water vapor. The model in each case was integrated for 30 years after a spinup to equilibrate the mean stratospheric H\textsubscript{2}O concentrations.

### 3. Interannual variation of MLS H\textsubscript{2}O

Figure 1 shows the time–height cross section of the Aura MLS H\textsubscript{2}O volume mixing ratio averaged over 12\textdegree{}S–12\textdegree{}N. We show data only to 0.46 hPa (a little higher than stratopause level) because the main purpose of this study is to investigate stratospheric interannual H\textsubscript{2}O variations and particularly those associated with the QBO (which becomes quite weak above 1 hPa; Hamilton 1981; Baldwin et al. 2001; Baldwin and Gray 2005). The minimum annual-mean H\textsubscript{2}O mixing ratio is observed near the tropical tropopause, and it is clear that the vertical gradient of annual-mean H\textsubscript{2}O concentration is positive and becomes larger above about 10 hPa. This vertical stratification of the mean H\textsubscript{2}O concentration agrees with earlier observations (e.g., Randel et al. 2004).

The annual cycle in Fig. 1 is interpreted as resulting from more upward H\textsubscript{2}O transport from June to October and less upward H\textsubscript{2}O transport from December to April, because of the seasonal temperature cycle at the tropical tropopause and consequent variations in saturation mixing ratios at the tropical tropopause (Mote et al. 1996, 1998).

Figure 2a shows the frequency power spectra of MLS H\textsubscript{2}O in 12\textdegree{}S–12\textdegree{}N as a function of height. Before calculating the power spectra, the linear trend was removed. There are three major spectral peaks in the stratosphere—at 6, 12, and 20–40 months—that correspond to semiannual, annual, and QBO variability, respectively. These three peaks are evident at all heights, but they are weaker near 10 and 0.5 hPa. In the troposphere and near the tropopause around 70–100 hPa, rather than a QBO peak, one finds variability spread over a broader spectral range, possibly indicating the significance of ENSO variations on this region of the tropical atmosphere. The procedure to extract interannual variability is as follows: the mean seasonal cycle (i.e., annual cycle) is calculated using data from August 2004 to January 2014, and then the values are subtracted.
from the raw data. The resulting series were then smoothed by taking 5-month running means (e.g., Kawatani and Hamilton 2013). Figure 2b shows the spectra for the deseasonalized and smoothed time series. Most components extracted by this method are concentrated in periods of 20–40 months in the stratosphere—that is, the QBO period’s ranges—while components with periods longer than 40 months and those around 8–11 months remain in the troposphere and tropopause regions.

Figure 3 shows vertical profiles of the time-mean 12°S–12°N MLS H2O mixing ratio, along with the total standard deviation and standard deviations due to the annual cycle and due to interannual components. The MLS H2O mixing ratio in the tropics reaches its minimum at 82.5 hPa, at or just above the tropopause, and it increases with height in the stratosphere. The cold-point tropopause typically lies between 100 and 82 hPa, and the mean water vapor seen in earlier HALOE observations also shows a minimum at 82 hPa (Randel et al. 2004). The annual cycle is dominant in the upper troposphere and lower stratosphere, while interannual variability becomes comparable to annual variability in the upper stratosphere.

Figure 4 illustrates the time–height cross section of the interannual anomaly of MLS H2O in 12°S–12°N compared with the observed deseasonalized and smoothed (5-month running mean) zonal wind over Singapore from August 2004 to January 2014 [data from Kunze (2014)]. Note that blue colors correspond to positive H2O anomalies. Inspection shows this period included roughly 4.5 cycles of both the wind QBO and the dominant interannual variation of H2O in the stratosphere. Upward-propagating anomalies are clearly seen from the lower stratosphere to the middle stratosphere, and in their rather uniform upward propagation they resemble the annual tape-recorder signal apparent in the water vapor in this altitude range (e.g., Figure 1). The interannual H2O anomalies display more variability than the wind signals, which could reflect the presence of other sources of interannual variability for H2O, including ENSO (Geller et al. 2002).

At higher levels—say, above 10–15 hPa—the anomalies in H2O seem to propagate downward. The interannual variability in the upper and lower stratosphere appear either unrelated or perhaps have a phase cancellation around 10 hPa (note the power spectrum of interannual anomalies has minimum values around 10 hPa; Fig. 2).

To isolate the effects of the QBO on interannual variations of H2O, a composite based on the phase of the zonal wind QBO was computed. Month 0 of the composite is taken to be when the zonal winds at 30 hPa in the deseasonalized and smoothed Singapore wind series changes from westerly to easterly. Composite values were then computed for ±18 months around these zero months (i.e., February 2007, May 2009, and September 2011; see Fig. 4b).
The time–height cross section of the QBO composite for H$_2$O mixing ratio is shown in Fig. 5. While a composite based on just three cycles cannot remove all the extraneous signals, the QBO effect on H$_2$O is quite apparent in Fig. 5. Below about 20 hPa, the upward-propagating tape-recorder signal is clearly seen as discussed in previous studies (Randel et al. 1998, 2004; Giorgetta and Bengtsson 1999; Geller et al. 2002). The difference between the maximum and minimum anomalies around the tropopause is approximately 0.25 ppmv. The anomalies in Fig. 5 propagate vertically between 100 and 20 hPa with an estimated mean speed of about 8.5 km yr$^{-1}$ ($\sim$0.27 mm s$^{-1}$), similar to the propagation speed of the annual cycle in H$_2$O (Mote et al. 1998; Niwano et al. 2003) and also to other estimates of the mean upwelling in the tropical lower stratosphere (Rosenlof 1995).

In the upper stratosphere, downward-propagating signals are apparent. Seen together, the upward-propagating and downward-propagating signals form a “boomerang” pattern in the time–height plot. In earlier studies of satellite data, Randel et al. (1998) and Geller et al. (2002) noted that the regular upward propagation of the water vapor anomalies in the lower stratosphere was not seen above about 30 km, but they did not characterize further the downward propagation of the anomalies. With our analysis of about 10 years of MLS observations, the upper-stratospheric variations are better characterized and the dominant downward propagation of QBO-related water vapor anomalies there is quite clear.

![Fig. 4](Image)

**Fig. 4.** (a) Time–height cross section of the 12°S–12°N interannual anomaly of MLS water vapor mixing ratio and (b) observed deseasonalized and smoothed zonal wind over Singapore from August 2004 to January 2014, provided by the Free University of Berlin (FUB). The color intervals are (a) 0.05 ppmv and (b) 5 m s$^{-1}$. For H$_2$O, blue colors correspond to positive values (more water vapor).

![Fig. 5](Image)

**Fig. 5.** Composite of the QBO in interannual variation of 12°S–12°N average H$_2$O where month 0 corresponds to the westerly-to-easterly transition of the zonal wind at 30 hPa. The color interval is 0.05 ppmv.
4. Interannual variation of H₂O in the MIROC-AGCM

We now turn to results from the long MIROC-AGCM simulations described in section 2. As noted earlier, the model has fine vertical resolution up to 5 hPa, but coarser resolution and an artificial damping are imposed higher up. So, we show results only up to 5 hPa. The red curve in Fig. 6a shows the vertical profile of the mean H₂O mixing ratio averaged over 12°S–12°N in the control simulation, and it is compared with the MLS observed result (black curve). The basic pattern of minimum water vapor concentration near 80–90 hPa with rising values above that point is seen in both the model and MLS data. However, the minimum is deeper in the observations (3.7 vs 4.1 ppmv in the model) and the vertical gradient is larger in the observations at least above approximately 70 hPa. The simulated long-term annual-mean and zonal-mean temperature in the equatorial region is nearly identical to that in the Interim ECMWF Re-Analysis (ERA-Interim). However, the MIROC-AGCM generally has warm biases in the coldest tropical tropopause temperatures, which occur over the equatorial western Pacific (Holton and Gettelman 2001; Zhou et al. 2004), compared with ERA-Interim. The MIROC-AGCM has approximately 1-K warm biases around there, corresponding to a saturation mixing ratio bias of about 0.5 ppmv. The blue curve in Fig. 6a shows results from the MIROC simulation without the methane oxidation source [note that Giorgetta and Bengtsson (1999) also had no methane oxidation source]. In this case the vertical gradient of H₂O concentration is actually negative above 70 hPa.

Figure 6b shows the standard deviations of the annual cycle of the H₂O mixing ratios as a function of height for the two model experiments compared with the MLS observations. The model results significantly underestimate the variance above about 40 hPa. Figure 6c shows the standard deviation of the interannual anomalies, again comparing the two model runs with the MLS observations. The interannual standard deviation is smaller in the model than in the MLS observations at all altitude ranges. Possible contributors to the smaller interannual variation in the modeled water vapor mixing ratio include the use of climatological SSTs and an amplitude of the simulated dynamical QBO that is somewhat smaller than observed (Kawatani et al. 2011).

Figures 7a–d show the QBO composite of zonal-mean zonal wind, temperature, residual vertical velocity, and H₂O mixing ratio in the MIROC T106 control simulations. For each variable, the results presented are averaged over 12°S–12°N. The procedure for making the composite is the same as for that described in section 3 for observations (Fig. 5) but a total of 13 cycles from the 30-yr simulations are averaged for the model. The model simulates a QBO-like oscillation in the zonal wind with a period close to 24 months (Fig. 7a). The simulated QBO amplitude is smaller than that in the real world, especially in the lower stratosphere. Given the presumed role of the cold-point tropopause in dehydration of air entering the stratosphere, it is of interest to characterize the model QBO at that level (83 hPa in the

![Fig. 6. Vertical profiles of (a) the 12°S–12°N mean H₂O and its standard deviation due to (b) annual and (c) interannual components. Profiles of MLS and model simulation with and without methane parameterization are drawn by black, red, and blue lines, respectively. Intervals of the abscissa are (a) 0.5, (b) 0.2, and (c) 0.05 ppmv.](http://journals.ametsoc.org/jas/article-pdf/71/11/4072/3818577/jas-d-14-0164_1.pdf)
The QBO amplitude at 80 hPa calculated using the FUB monthly-mean winds for Singapore (1.4°N) during the 2000s is about 4.1 m s\(^{-1}\), whereas the simulated QBO amplitude at 83 hPa for 1.7°S–1.7°N is about 1.2 m s\(^{-1}\). The MIROC model-simulated QBO in temperature near the tropopause is \(\pm 0.2\) K at 83 hPa and \(\pm 0.1\) K around 100 hPa (Fig. 7b), whereas it is known to range up to about \(\pm 0.5\) K in the real world (Randel et al. 2000). The variations of MLS \(\text{H}_2\text{O}\) mixing ratio at 83 hPa in the QBO composite are about \(\pm 0.25\) ppmv (Fig. 5), while those in the model are about \(\pm 0.06\) ppmv (Fig. 7d). Although the simulated QBO variations in zonal wind, temperature, and \(\text{H}_2\text{O}\) are smaller than those from observations, the overall qualitative characteristics are similar in observations and the model simulation.

The QBO component of residual vertical velocity is downward (upward) when the vertical shear of the QBO zonal wind is positive (negative) (Fig. 7c), consistent with expectations (e.g., Plumb and Bell 1982). The simulated \(\text{H}_2\text{O}\) mixing ratio variation shows a similar boomerang pattern as that seen in the MLS data (Fig. 5): there are upward-propagating signals from the upper troposphere to the middle stratosphere and downward-propagating signals in the upper stratosphere. The transition between upward and downward propagation in the model is around 15–20 hPa, which is a little lower down than in the MLS observations.

From the model results, a detailed budget for the near-equatorial \(\text{H}_2\text{O}\) can be calculated. We regard the time series of any quantity as being composed of a long-term mean, a component related to the dynamical QBO, and everything else (including the annual cycle and non-QBO-related interannual variations). So, for example, the \(\text{H}_2\text{O}\) mixing ratio consists of long-term mean \(\bar{q}\), QBO component \(q'\), and others, \(Rq\), so

\[
q = \bar{q} + q' + Rq. \tag{1}
\]

In practice the overbar is a time mean over the whole record and the prime indicates the component that is isolated by the QBO compositing procedure outlined above. We also divide the residual-mean vertical velocity \([\vec{w}]\) in the transformed Eulerian-mean (TEM) formalism (cf. Andrews et al. 1987); hereinafter, just denoted as \(w\) into a time mean, QBO, and other components. The zonal-mean \(\text{H}_2\text{O}\) budget for the QBO component of water vapor mixing ratio is then expressed as

\[
\frac{\partial q'}{\partial t} = \vec{w} \frac{\partial q'}{\partial z} - w \frac{\partial \bar{q}}{\partial z} + \text{residual}, \tag{2}
\]

where “residual” includes the effects of quadratic terms involving the annual cycle and other components as well as effects of meridional advection and chemical sources or sinks as they might project onto the QBO. The first term on the right-hand side expresses the mean advection of the QBO variation of \(\text{H}_2\text{O}\) mixing ratio by mean
upwelling. The second term on the right-hand side indicates the advection of the mean vertical gradient of H$_2$O by the QBO component of vertical velocity.

Figures 8a and 8c show the time variation of the H$_2$O budget terms through the composite QBO cycle in the control run. Results for 5–15 and 30–50 hPa are shown. At 30–50 hPa, the actual tendency in the H$_2$O mixing ratio is mainly accounted for by the advection of QBO H$_2$O anomalies by the mean upwelling (–w$^\theta$q/∂z). Conversely, at 5–15 hPa, the total tendency is driven by QBO variation in vertical velocity advecting the mean gradient (–w$^\theta$q/∂z) and is opposed by the advection of QBO mixing ratio anomalies by the mean upwelling. Below 15 hPa, the QBO-related water vapor anomalies propagate upward via the familiar tape-recorder effect, while above 15 hPa the water vapor anomalies display a downward propagation characteristic of the propagation of the dynamical QBO itself.

Figure 7e is as in Fig. 7d, but for the simulation without the parameterization of methane oxidation, as discussed in section 2 above. The zonal wind, temperature, and residual vertical velocity composites (not shown) in this case are nearly identical to those in the control simulation, which includes the methane oxidation parameterization. The H$_2$O mixing ratio composite in Fig. 7e is similar to that in the control run up to about 20 hPa, but quite different at higher levels. In fact, the positive and negative phases around 5–15 hPa in Fig. 7e are opposite to those in the composite from the simulation with the methane oxidation source (Fig. 7d).

Figures 8b and 8d present the same budget terms as Figs. 8a and 8c, but for the simulation without methane oxidation. At 30–50 hPa, the dominant term that drives the H$_2$O tendency is the advection of QBO mixing ratio anomalies by the mean upwelling, just as that in the simulation with the methane oxidation source (Fig. 7d).
opposed by both the advection by the mean vertical wind and the residual term, leaving a small net tendency. In the simulation without the methane oxidation source, the sign of the vertical gradient of mean H_2O (\partial[H_2O]/\partial z) in the stratosphere is opposite to that in the control simulation (Fig. 6a) and this leads to very different results for the QBO in water vapor concentration above 15 hPa.

5. Interannual variations of H_2O in CMIP5 models

We have investigated the variability of stratospheric equatorial H_2O in long-term climate simulations for four of the coupled ocean–atmosphere global models whose results are available through CMIP5. CMIP5 included four models that simulate a reasonable QBO in the equatorial zonal wind (Kawatani and Hamilton 2013). Specifically, these are the Max Planck Institute Earth System Model, medium resolution (MPI-ESM-MR; Roeckner et al. 2006; Schmidt et al. 2013), the Hadley Centre Global Environment Model, version 2 - Carbon Cycle (HadGEM2-CC; Collins et al. 2011; Jones et al. 2011; Martin et al. 2011), the Model for Interdisciplinary Research on Climate, Earth System Model, Chemistry Coupled (MIROC-ESM-CHEM), and the Model for Interdisciplinary Research on Climate, Earth System Model (MIROC-ESM; Watanabe et al. 2011). For each of the four models, we analyzed a single realization of the CMIP5 “historical” run (forced with observed greenhouse gas and aerosol concentrations; Taylor et al. 2012) from 1950 to 1999. Since the response to changing climate forcing is not an issue for this paper, the first step in our analysis of each of the CMIP5 model time series was to remove any linear trend over the 50 years considered.

The left panels in Fig. 9 show the QBO composite of H_2O for each of the four CMIP5 models. The top boundary of all four CMIP5 models considered is roughly 0.01 hPa, but we show results only up to 0.5 hPa. Note that all these models include some representation of the methane oxidation process and simulate positive vertical gradients of mean H_2O (right panels). All four models clearly show the boomerang structures of H_2O (\partial[H_2O]/\partial z) in the stratosphere is opposite to that in the control simulation (Fig. 6a) and this leads to very different results for the QBO in water vapor concentration above 15 hPa.

One feature in which the models differ markedly is the simulated mean equatorial water vapor profile in the region above about 20 hPa. Two models, MPI-ESM-MR (Fig. 9a) and MIROC-ESM (Fig. 9d), have vertical gradients in this region that are reasonably similar to those observed by MLS (although the models have an overall dry bias throughout the equatorial stratosphere). The other two models, HadGEM2-CC (Fig. 9b) and MIROC-ESM-CHEM (Fig. 9c), have unrealistically small vertical gradients above 20 hPa. In the upper stratosphere, models with stronger H_2O vertical gradients simulate stronger downward-propagating QBO anomalies compared with models with weaker gradients. Although TEM vertical velocity fields for these CMIP5 models are not known, these results are at least consistent with our view that the mean H_2O vertical gradient plays a key role in generating downward-propagating water vapor anomaly signals.

Figure 10 shows the time variation of the QBO composite temperature at 100 hPa, and water vapor at 100 and 70 hPa in each of the CMIP5 model simulations and in our T106 MIROC-AGCM control simulation (note that choice of levels to analyze was constrained by requiring data availability for all four CMIP5 models). It is clear that models with larger QBO temperature variability at 100 hPa have larger QBO water vapor concentration variability at 100 hPa, and there seems to be little phase lag between the QBO temperature variations and the water vapor variations. This is consistent with the notion that simple cold trapping determines the water vapor mixing ratios at 100 hPa. Figure 10c shows the same composites for the QBO, but for water vapor at 70 hPa. The QBO water vapor signals are similar to those seen at 100 hPa but are somewhat smaller and delayed by about 2 months. This would be consistent with the usual tape-recorder effect, assuming some dilution of the upward-propagating air near the equator.

These intermodel comparisons clarify the important role of QBO tropopause temperature variations in upward-propagating H_2O anomalies in the lower stratosphere and that of the mean H_2O vertical gradient in downward-propagating anomalies in the upper stratosphere, which support our conclusions obtained from our MIROC-AGCM experiments.

6. Summary and concluding remarks

The classic studies of Mote et al. (1996, 1998) show the seasonal cycle of water vapor mixing ratio in the equatorial lower stratosphere can be explained by introduction
FIG. 9. (left) Composite of the QBO in the interannual variation of H$_2$O mixing ratio for (a) MPI-ESM-MR, (b) HadGEM2-CC, (c) MIROC-ESM-CHEM, and (d) MIROC-ESM from 1950 and 1999 in the historical run. (right) Profile of mean H$_2$O from each model (red) with that from MLS observations (black). The color contours are $\pm 0.01$, $\pm 0.02$, $\pm 0.04$, $\pm 0.08$, and $\pm 0.12$ ppmv.
of variations in the saturation mixing ratio of air passing upward through the cold-point tropopause; these variations are then advected upward by the mean vertical upwelling. As noted in these early papers, this basic tape-recorder mechanism also can account for the upward propagation of QBO-related anomalies that are observed to appear near the tropopause, a conclusion supported by the modeling study of Giorgetta and Bengtsson (1999). The early observational and modeling work was extended to the equatorial upper stratosphere by Geller et al. (2002), who pointed out that the interannual fluctuations in the upper stratosphere could not be explained by the tape recorder and must depend on the interannual variations of the transport circulation itself.

These previous studies characterized interannual variations of water vapor concentrations, but the supporting evidence from models and observations in each case has some significant limitations. In the present study, we revisited this issue of interannual variations in equatorial water vapor through application of recent satellite data and results from several state-of-the-art comprehensive global simulation models. We use nearly 10 years of observations from the MLS instrument on the NASA Aura satellite, which allowed us to make a three-cycle QBO composite keyed off the dynamical QBO, and it provides a much nicer view of the systematic QBO-related variations than can be seen in any of the earlier published observational records of equatorial stratospheric H₂O anomalies. We investigated the time–height structure of interannual variations in equatorial H₂O concentration using these longer and higher-quality observational records. From the upper troposphere to the middle stratosphere, H₂O concentration anomalies were found to propagate upward in a manner analogous to the seasonal “tape recorder” (Mote et al. 1996), which is consistent with previous observational and modeling studies (Randel et al. 1998, 2004; Giorgetta and Bengtsson 1999; Geller et al. 2002). On the other hand, clear downward-propagating anomalies are found above about 10–15 hPa.

We examined the interannual equatorial stratospheric water vapor variations in the control integrations conducted with a fine-horizontal-and-vertical-resolution (T106L72) version of the MIROC-AGCM and in four models in the CMIP5 that are known to simulate fairly realistic dynamical QBOs (Kawatani et al. 2011, 2012; Kawatani and Hamilton 2013). We showed that the global models all simulate somewhat realistic interannual water vapor variations in the equatorial stratosphere. In particular, the model-simulated H₂O concentration displays the same basic “boomerang” pattern as the MLS data with rather uniform upward propagation from the tropopause to some midstratospheric level and downward propagation of anomalies at higher levels. It is apparent that the interannual water vapor anomalies in both models and observations are dominated by the familiar stratospheric QBO.

The detailed data available from the high-resolution MIROC-AGCM simulation allowed a budget analysis of the zonal mean H₂O mixing ratio based on a QBO compositing procedure. This showed that the upward propagation in the equatorial lower stratosphere is indeed caused by the mean advection of interannual water content anomalies induced by the QBO at the tropopause, while the downward propagation is primarily due to the advection of the mean vertical gradient of water content by the QBO fluctuations in vertical wind. We are also able to demonstrate the central role of the mean H₂O vertical gradient in the downward propagation with our experiment with the methane oxidation source turned
off. The importance of these two mechanisms had been proposed earlier by Geller et al. (2002) and Fujiwara et al. (2010), but the pictures that we present are more complete and our conclusion is more secure.

We also analyze QBO-related water vapor variations using the four CMIP5 models that simulate a reasonable QBO. The models with larger tropopause temperature anomalies induced by the QBO have larger lower-stratospheric water vapor anomalies, while the models with stronger mean H2O vertical gradients display stronger upper-stratospheric water vapor variations. The intermodel comparisons support our conclusions from our MIROC-AGCM simulations.

The high-resolution MIROC model results for the QBO in H2O concentration shown here, while qualitatively similar to observations, do display significant differences. These are likely related in large part to deficiencies in the mean water vapor simulation and in the detailed structure of the simulated dynamical QBO. Efforts to improve the model in these respects should be continued. Also, while we have produced a reasonably straightforward picture for the nature and causes of the QBO-related water vapor concentration anomalies, more observational and modeling research could help also explain those interannual variations that are not directly related to the QBO, including variations that may be linked to ENSO variability in the troposphere.

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