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# Stratospheric water vapor feedback and its climate impacts in the coupled atmosphere–ocean Goddard Earth Observing System Chemistry–Climate Model

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## Abstract

The stratospheric water vapor feedbacks on climate for abrupt CO<sub>2</sub> quadrupling are investigated with the coupled atmosphere–ocean Goddard Earth Observing System Chemistry–Climate model. A feedback suppression method is used to quantify the stratospheric water vapor climate feedback parameter and the impacts of stratospheric water vapor increases on temperature and circulation. It is found that increases in stratospheric water vapor change the model’s net climate feedback parameter by 0.11 W m<sup>−2</sup> K<sup>−1</sup>, contributing to 0.5 K, or 10%, of the global-mean surface warming under abrupt CO<sub>2</sub> quadrupling. Stratospheric water vapor increases lead to significant impacts on stratospheric temperature and dynamics. The increases induce stratospheric dynamical changes that strongly modify stratospheric cooling patterns. About 30% of the acceleration of the stratospheric Brewer–Dobson circulation under 4×CO<sub>2</sub> is attributed to the stratospheric water vapor increases. In the troposphere, the stratospheric water vapor feedback plays a role in Arctic amplification and is responsible for 14% of the Arctic warming. It also affects tropospheric circulation, causing greater poleward shift of the northern hemisphere tropospheric midlatitude jet.

**Keywords** Stratospheric water vapor · Climate feedback · Climate impact · Brewer–Dobson circulation · Arctic amplification

## 1 Introduction

As the surface-troposphere climate system warms due to increasing carbon dioxide (CO<sub>2</sub>) and other well-mixed greenhouse gases, stratospheric water vapor (SWV) will increase because of stronger transport (Maycock et al. 2011) and enhanced methane oxidation (Austin et al. 2007). The stronger transport is mainly caused by higher tropical tropopause temperature allowing more tropospheric water vapor to enter the stratosphere (Fueglistaler and Haynes 2005). Since SWV is itself a greenhouse gas, SWV increase will amplify tropospheric and surface warming. Therefore, SWV has a positive climate feedback, although only the part of SWV changes caused by transport is considered to

be feedback. Most of the previous studies found that SWV is an important climate feedback mechanism (Stuber et al. 2001; Dessler et al. 2013; Banerjee et al. 2019). In these studies, the SWV climate feedback parameter is in the range of 0.1–0.3 W m<sup>−2</sup> K<sup>−1</sup>, which is comparable to the albedo feedback (Bony et al. 2006). However, Huang et al. (2016) reported a very small SWV feedback parameter of 0.02 W m<sup>−2</sup> K<sup>−1</sup>.

The different results between Huang et al. (2016) and other studies are likely due to the different methods used to estimate the SWV feedback. Huang et al. (2016) used the radiative kernel method, which calculates the SWV feedback parameter by multiplying increases in SWV with a pre-calculated radiative sensitivity kernel. Other studies assess the SWV feedback parameter by calculating stratosphere adjusted radiative perturbation at the tropopause caused by increases in SWV. Stratosphere temperature adjustment is important in order to correctly assess radiative forcing due to increases in SWV (e.g., Maycock et al. 2011; Banerjee et al. 2019). The common method to adjust stratospheric temperature is the Fixed Dynamical Heating (FDH) approach (Fels

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et al. 1980), which assumes that stratospheric dynamics do not change during the adjustment process. The SWV feedback parameter is then given by the stratosphere adjusted radiative perturbation normalized by the associated surface temperature change. Note that the kernel method of Huang et al. (2016) does not account for stratospheric temperature adjustment, which could be the main reason for the small SWV feedback in their calculation (Banerjee et al. 2019). The stratosphere adjusted radiative perturbation method has its own limitation as the FDH assumption may not hold because stratospheric dynamics can change significantly in response to SWV increase.

The climate feedback parameter is an important diagnostic to quantify the SWV impact on global-mean surface temperature. However, it is also important to understand the impacts of SWV changes on stratospheric and tropospheric temperature and circulation. The two methods described above can only determine the SWV climate feedback parameter, not the details of climate impact from the SWV feedback. Several idealized SWV perturbation studies have investigated the climate impact of SWV. For example, Oinas et al. (2001) studied stratospheric and tropospheric temperature response to a uniform 0.7 ppmv SWV increase above 150 hPa under a 6 ppmv background. Maycock et al. (2013) uniformly doubled SWV concentration from 3 to 6 ppmv and found significant responses in stratospheric temperature, tropical upwelling, and tropospheric circulation. It should be noted that in these studies the SWV increases are arbitrarily set and do not correctly represent the response to global warming.

This study has two purposes. The first purpose is to estimate the SWV feedback parameter in the coupled atmosphere–ocean Goddard Earth Observing System Chemistry–Climate Model (GEOS CCM) using a feedback suppression method. This method does not require the FDH assumption or pre-calculated radiative kernel. The advantage of this method is that it can quantify the climate impact of the SWV feedback. The second purpose of this study is to investigate the climate impact of the SWV feedback on tropospheric and stratospheric temperature and circulation.

## 2 Model and method

Our model is the coupled atmosphere–ocean GEOS CCM. The details of an earlier version of this model were described in Li et al. (2016). The current version has updated the atmosphere model to GEOS Heracles and ocean model to Modular Ocean Model (MOM) 5 (Griffies 2012). The simulations herein use a  $1^\circ$  longitude by  $1^\circ$  latitude resolution for both the atmosphere and ocean. The atmosphere has 72 vertical levels with a 0.01 hPa top. MOM5 has 50 layers with fine resolution in

the upper 200 m. Li et al. (2016) showed that the earlier model version simulated the twentieth century atmosphere and ocean climatology reasonably well. The current version has improved simulations of sea surface temperature and tropospheric jets compared to the earlier version (not shown).

The method we use to estimate the SWV feedback is similar to the online feedback suppression approach (Hall and Manabe 1999). We conducted two experiments, one with SWV feedback and one with suppressed SWV feedback. The SWV feedback and climate impact are then calculated from the differences of these two experiments.

Two abrupt  $\text{CO}_2$  quadrupling experiments were performed in order to isolate the SWV feedback: a control experiment (Control) and a fixed SWV experiment (Fixed-SWV). Control consists of a 50-year baseline run under fixed 2000 conditions and a 150-year  $4 \times \text{CO}_2$  run that abruptly increases  $\text{CO}_2$  concentration by 4 times from the end of the baseline run. FixedSWV is the same as the Control, except that SWV in both its baseline and  $4 \times \text{CO}_2$  runs is relaxed to a climatological field. Therefore, the SWV feedback is suppressed in FixedSWV. Note that methane does not contribute to SWV changes in both experiments because they are run without interactive chemistry. In order to avoid an SWV discontinuity at the tropopause when implementing the prescribed SWV climatology, a 50-hPa transition layer is set above the tropopause pressure. The SWV concentration is linearly interpolated from the model calculated value at the tropopause pressure to the prescribed climatological value at 50 hPa above the tropopause pressure. Above the transition layer, SWV is relaxed to the climatological value. Thus, FixedSWV does not completely remove the SWV feedback because SWV in the transition layer will still increase. However, FixedSWV significantly suppresses the increase of SWV. We will show in the next section that SWV in the overworld (above 100 hPa) does not change and SWV in the lowermost stratosphere (between tropopause and 100 hPa) increases about 40% of that in Control.

We quantify the net climate feedback parameter in Control and FixedSWV using the Gregory regression method (Gregory et al. 2004). The SWV feedback parameter is then calculated as the difference between the Control and FixedSWV feedback. Because FixedSWV allows limited increases of SWV in the lowermost stratosphere, the calculated SWV feedback parameter should be regarded as a lower limit.

The climate response to abrupt  $\text{CO}_2$  quadrupling in both Control and FixedSWV is defined as the difference between the last 50 years of the  $4 \times \text{CO}_2$  run and the 50-year baseline run. The climate impact of the SWV feedback is quantified by subtracting climate response in FixedSWV from that in Control.

### 3 Results

#### 3.1 SWV climate feedback parameter

Climate models consistently simulate increases in SWV with warming of the troposphere (e.g., Gettelman et al. 2010; Dessler et al. 2013; Banerjee et al. 2019). Figure 1a shows the annual and zonal mean water vapor response to  $4\times\text{CO}_2$  as a function of latitude and pressure in Control. Water vapor increases throughout the stratosphere. Changes in the lowermost stratosphere have a large vertical gradient, but changes in the overworld are more uniform. SWV response in the lowermost stratosphere is stronger than in the overworld in terms of absolute value: the average SWV increase in the lowermost stratosphere is 6.9 ppmv, nearly twice the average 3.6 ppmv increase in the overworld. In terms of relative changes, the SWV in the lowermost stratosphere and overworld increases 73% and 101%, respectively. The SWV as a whole increase 93%. There are also hemispheric differences with larger increase in the Northern Hemisphere (NH) than in the Southern Hemisphere (SH). The SWV responses in Control are similar to those in CMIP5 abrupt  $4\times\text{CO}_2$  experiments (Banerjee et al. 2019).

FixedSWV has very different SWV responses from those in Control by design. Figure 1b shows that SWV in the overworld doesn't change because it is set to the same climatology in the baseline and the  $4\times\text{CO}_2$  run. SWV increases in the lowermost stratosphere because of the transition layer above the tropopause, although the average increase of 2.9 ppmv is only about 40% of that

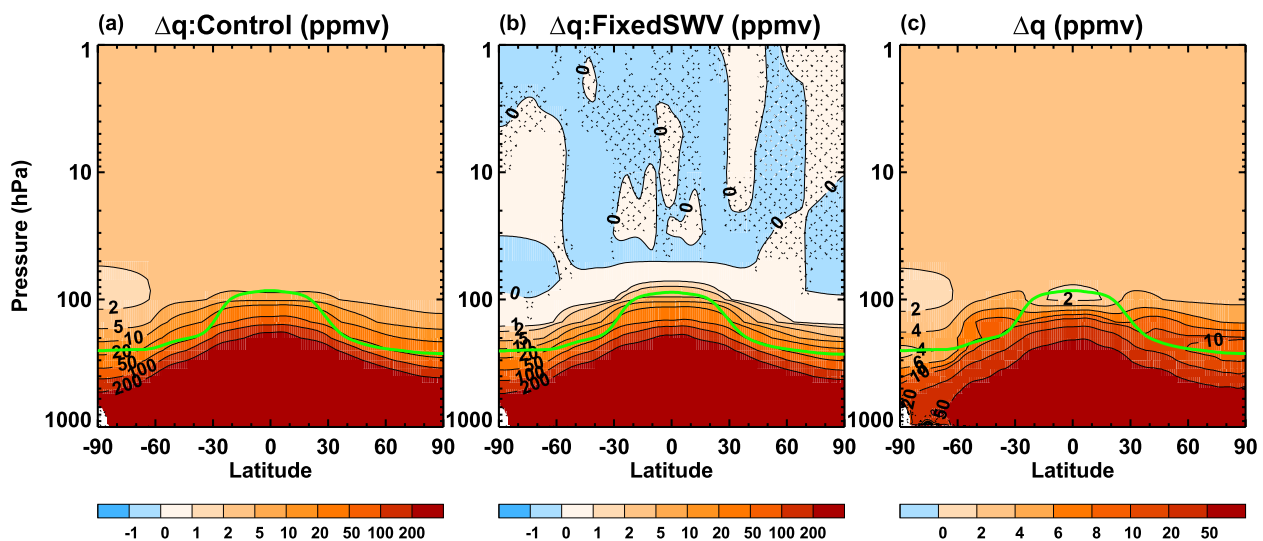
in Control. The different water vapor responses between Control and FixedSWV (Control minus FixedSWV) are shown in Fig. 1c. On average, Control has 3.6 ppmv more SWV in the overworld and 4.0 ppmv more SWV in the lowermost stratosphere than FixedSWV. The differences in the lowermost stratosphere are more pronounced in the NH than in the SH.

Increases in SWV, a greenhouse gas, will cause surface warming (e.g., Solomon et al. 2010). Figure 2a compares the timeseries of annual and global mean surface air temperature anomalies in the  $4\times\text{CO}_2$  runs of Control and FixedSWV. The anomalies are calculated relative to their respective 50-year average baseline results. The two timeseries start to diverge after about 10 years and clearly show enhanced warming in Control. At the last 50 years of the  $4\times\text{CO}_2$  runs, the global-mean surface warming difference is about 0.5 K, or 10%, larger in Control than in FixedSWV. The weaker surface warming in FixedSWV is caused by suppressed SWV response, indicating a significant SWV feedback in the GEOS CCM.

The SWV feedback is quantified using the linear regression method of Gregory et al. (2004). This method assumes that, in an abrupt  $\text{CO}_2$  increasing experiment, the annual and global mean radiative imbalance at the top of the atmosphere ( $N$ ) and the surface air temperature response ( $\Delta T$ ) is linearly related by

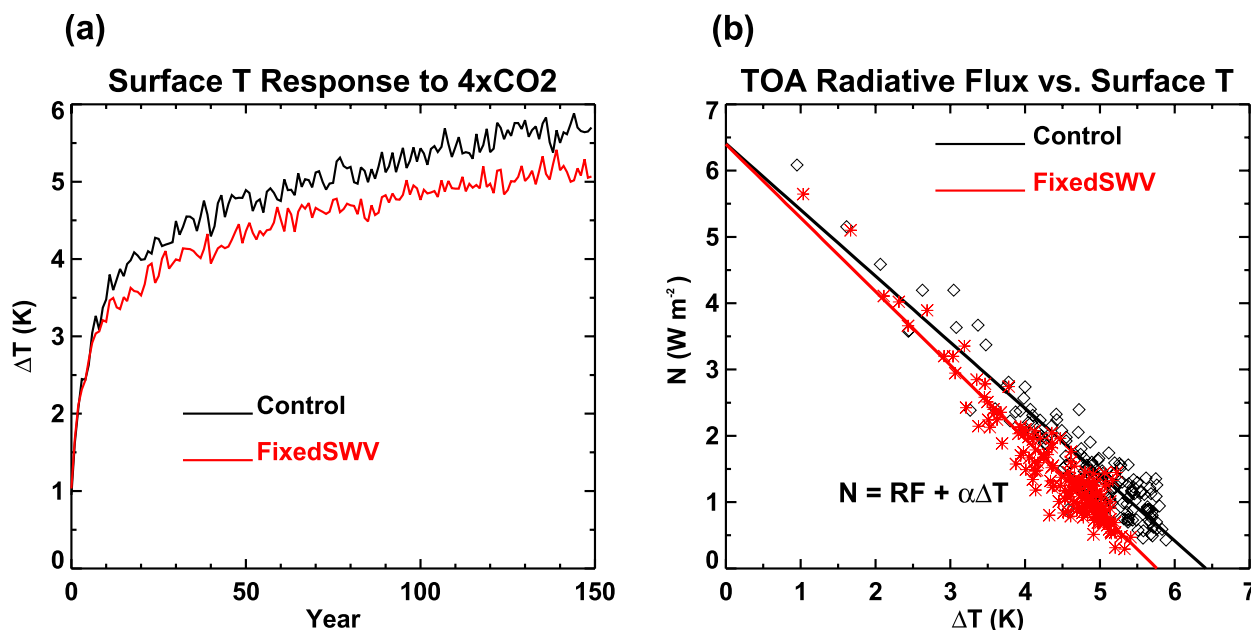
$$N = RF + \alpha \Delta T$$

where  $RF$  is the radiative forcing from increasing  $\text{CO}_2$  and  $\alpha$  is the climate feedback parameter. Simply regressing  $N$  against  $\Delta T$ , the slope gives  $\alpha$  and the y-intercept is  $RF$ . Figure 2b shows the scatter plot of  $N$  against  $\Delta T$  for Control (black diamond) and FixedSWV (red asterisk). In both experiments,  $N$  and  $\Delta T$  are strongly correlated with a



**Fig. 1** **a** The annual and zonal mean water vapor response to  $4\times\text{CO}_2$  in Control. **b** The annual and zonal mean water vapor response to  $4\times\text{CO}_2$  in FixedSWV. **c** The differences between **a** and **b**. The green

line is the tropopause. Stippling indicates that the response is not statistically significant at two-tailed 5% level



**Fig. 2** **a** Timeseries of the annual and global mean surface air temperature anomalies for Control (black) and FixedSWV (red). **b** Scatter plot of the annual and global mean top of atmosphere radiative

flux anomalies ( $N$ ) against surface air temperature anomalies ( $\Delta T$ ) for Control (black diamond) and FixedSWV (red asterisk). The solid lines are linear regression fits

**Table 1** The climate feedback parameter ( $\alpha$ ) and the equilibrium surface air temperature response to  $\text{CO}_2$  quadrupling ( $\Delta T_e$ ) in Control and FixedSWV

Experiment	$\alpha$	( $\text{W m}^{-2} \text{K}^{-1}$ )	$\Delta T_e(\text{K})$
Control	- 1.00		6.42
FixedSWV	- 1.11		5.76

correlation coefficient of 0.94. Thus, the linear assumption in the Gregory method holds (the solid lines in Fig. 2b are regression fits). The regression fit for FixedSWV is steeper than that for Control. As a result, the x-intercept for the regression fit, which represents the equilibrium surface air temperature response to  $\text{CO}_2$  quadrupling, is smaller in FixedSWV than in Control. Table 1 compares the climate feedback parameter  $\alpha$  and the equilibrium surface air temperature response  $\Delta T_e$ . The climate feedback parameter in Control ( $\alpha = -1.00 \text{ W m}^{-2} \text{K}^{-1}$ ) is 10% smaller than that in FixedSWV ( $\alpha = -1.11 \text{ W m}^{-2} \text{K}^{-1}$ ). The SWV feedback amplifies  $\Delta T_e$  from 5.76 K in FixedSWV to 6.42 K in Control. Thus, 0.66 K, or 10% of surface warming in the response to  $4 \times \text{CO}_2$  is attributed to the SWV feedback.

Our method does not directly calculate the SWV climate feedback parameter. However, we can estimate the SWV feedback parameter to be  $0.11 \text{ W m}^{-2} \text{K}^{-1}$  by the difference of climate feedback parameters between Control and FixedSWV. This is because the net climate feedback parameter can be assumed to be a linear combination of feedback

parameters associated with different physical processes, and the only difference between Control and FixedSWV is the SWV feedback (Bony et al. 2006).

Our estimated SWV feedback parameter of  $0.11 \text{ W m}^{-2} \text{K}^{-1}$  should be regarded as a lower limit because our method underestimates the SWV response in the lowermost stratosphere by 40%. This value is within the range of previously reported values derived from the stratosphere adjusted radiative perturbation method. For example, Banerjee et al. (2019) calculated the SWV feedback parameter for 27 CMIP5 models and found that the mean SWV feedback is  $0.15 \pm 0.04 \text{ W m}^{-2} \text{K}^{-1}$ . On the other hand, our results do not support the small value of  $0.02 \text{ W m}^{-2} \text{K}^{-1}$  calculated from the radiative kernel method in Huang et al. (2016).

We can evaluate how much the calculated SWV feedback parameter is underestimated. The emissivity of SWV is linearly proportional to its column concentration under the water vapor weak line approximation. Therefore, we can use the SWV column concentration as a proxy for SWV radiative forcing. The SWV column concentration (integrated from tropopause to 1 hPa) increases  $5.9 \text{ gm}^{-2}$  in Control and  $1.7 \text{ gm}^{-2}$  in FixedSWV. Thus, our method underestimates the SWV column concentration response to  $4 \times \text{CO}_2$  by 29%, suggesting the calculated  $0.11 \text{ W m}^{-2} \text{K}^{-1}$  is underestimated by 29%.

The impact of SWV feedback on global-mean surface temperature depends not only on the climate feedback parameter, but also on the equilibrium climate sensitivity of the model (Dessler et al. 2013). It may also depend on

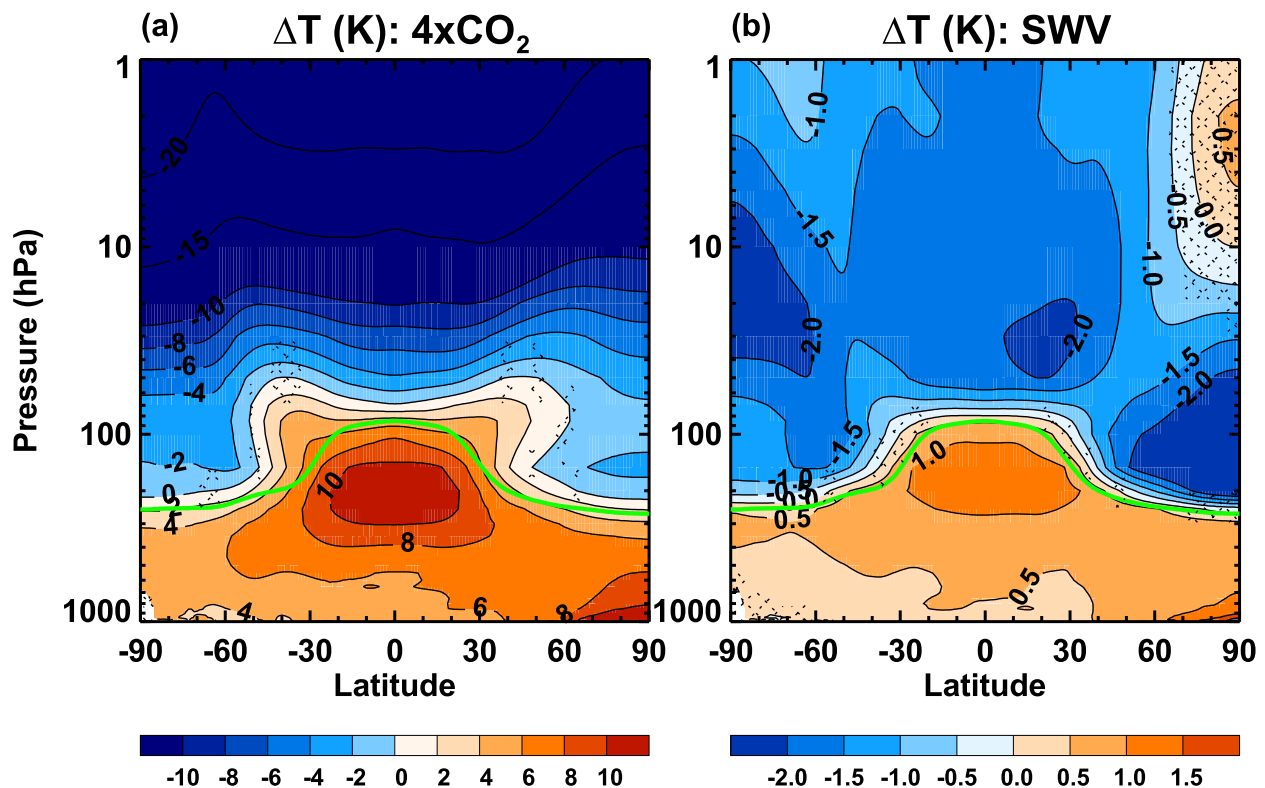
the experimental design. For example, Stuber et al. (2001) performed a CO<sub>2</sub> perturbation experiment by increasing 76 ppmv CO<sub>2</sub> and found a SWV climate feedback parameter of 0.24 W m<sup>-2</sup> K<sup>-1</sup>. They also conducted an experiment to eliminate the SWV feedback by fixing the SWV in the perturbation run. Surprisingly, even with the large SWV feedback parameter, Stuber et al. (2001) only found a 1% decrease of the global-mean surface air temperature when the SWV feedback was eliminated. Our results contrast to Stuber et al. (2001): our SWV feedback parameter is about half of that in Stuber et al. (2001) (0.11 W m<sup>-2</sup> K<sup>-1</sup> vs. 0.24 W m<sup>-2</sup> K<sup>-1</sup>), but the impact on global-mean surface air temperature is 10 times larger (10% vs. 1%). Note that in our experiment the CO<sub>2</sub> perturbation is much stronger (1106 ppmv) and the associated SWV increase is much larger. The different results between this study and Stuber et al. (2001) are possibly due to nonlinear response of surface air temperature to SWV increase.

### 3.2 Stratospheric impacts of SWV feedback

The advantage of our method is that we can quantitatively determine the contribution of the SWV feedback to climate change under CO<sub>2</sub> quadrupling. Figure 3a shows the

annual and zonal mean temperature response in Control, while Fig. 3b shows temperature change due to the SWV feedback (i.e., Control minus FixedSWV). As expected, increases in SWV cause warming of the troposphere and cooling of the stratosphere. In the troposphere, the patterns of temperature change induced by the SWV feedback, such as enhanced warming in the tropical upper troposphere and Arctic lower troposphere, are very similar to those induced by CO<sub>2</sub>. However, stratospheric temperature responds very differently to increases in CO<sub>2</sub> and SWV. The response to increasing CO<sub>2</sub> is a stratospheric cooling that increases with height and reaches a maximum at the stratopause (Fig. 3a). But the response to increasing SWV is a stronger cooling in the lower stratosphere with localized patterns of enhanced cooling (Fig. 3b). Enhanced cooling of more than 1.5 K is found in the Arctic lower stratosphere, Antarctic lower to middle stratosphere, and tropical stratosphere. Reduced cooling of less than 1.5 K occurs in the subtropical lower stratosphere and extends upward and poleward. The Arctic upper stratosphere actually shows a warming, although it is not statistically significant.

The SWV feedback plays a particularly important role in the extratropical lowermost stratosphere. Increases in SWV cause more than 2 K cooling in the NH lowermost



**Fig. 3** **a** The annual and zonal mean temperature response to 4×CO<sub>2</sub> in Control. **b** The impact of the stratospheric water vapor feedback on the annual and zonal mean temperature. The green line is the tropo-

pause. Stippling indicates that the response is not statistically significant at two-tailed 5% level

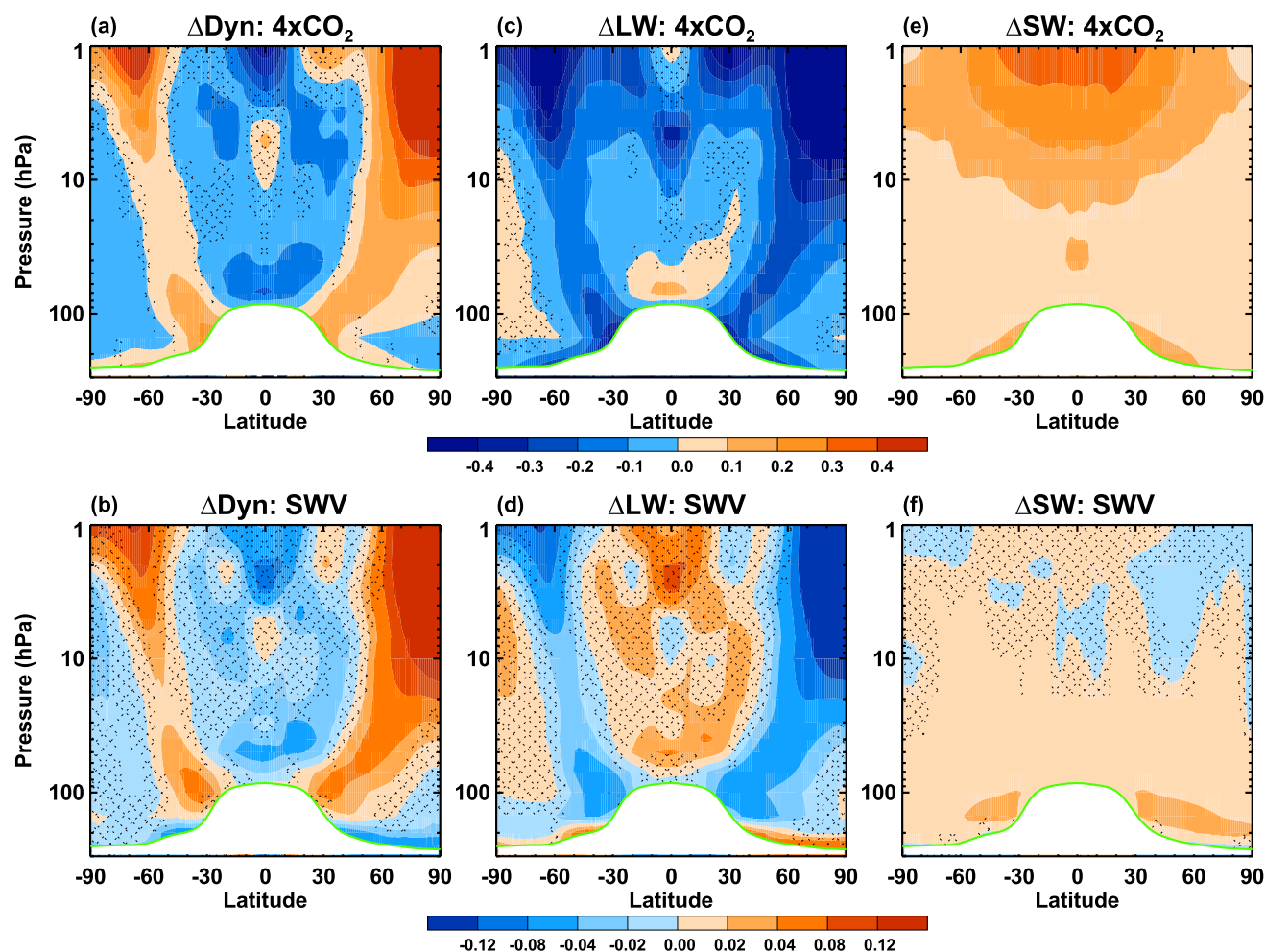


stratosphere (Fig. 3b), which accounts for all the cooling in this region in Control (Fig. 3a). In the SH lowermost stratosphere, the SWV-induced cooling of 1–1.5 K contributes to half of the temperature response to  $\text{CO}_2$  quadrupling in Control. The strong cooling in the lowermost stratosphere changes the latitudinal temperature gradient, which affects stratospheric dynamics. Changes in stratospheric dynamics could further modify the stratospheric temperature response.

The patterns of stratospheric temperature response to SWV increase, with regions of enhanced and reduced cooling, strongly suggest a modification of the stratosphere's dynamics. Figure 4a and b show the stratospheric dynamical heating rate response to  $4\times\text{CO}_2$  and to SWV increase (i.e., Control minus FixedSWV), respectively. Both figures show similar structure, with dynamical cooling in the tropics and dynamical heating that extends from the subtropics at the tropopause to the polar region in the upper stratosphere. This

structure indicates a stronger tropical upwelling and extra-tropical downwelling, i.e., an acceleration of the Brewer–Dobson circulation. About 20–25% of the dynamical heating rate response in Control is attributed to the SWV increase. Comparing Fig. 4b with Fig. 3b reveals that dynamical heating corresponds to reduced cooling (or warming in the case of Arctic upper stratosphere) and vice versa, confirming strong dynamical modification of stratospheric temperature response to SWV increase.

Changes in the stratospheric dynamical heating rate are balanced by changes in the longwave and shortwave radiation. In response to  $4\times\text{CO}_2$ , the longwave radiation change dominates the shortwave radiation change and shows nearly opposite patterns to those in dynamical heating rate (Fig. 4c). The response of shortwave radiation is warming throughout the stratosphere and becomes stronger in the upper stratosphere (Fig. 4e). The longwave radiation response to SWV



**Fig. 4** Upper panel: the annual and zonal mean response to  $4\times\text{CO}_2$  in Control for **a** dynamical heating rate, **c** longwave radiative heating rate, and **e** shortwave radiative heating rate. Lower panel: the impact of stratospheric water vapor feedback on the annual and zonal mean **b**

dynamical heating rate, **d** longwave radiative heating rate, and **f** shortwave radiative heating rate. The unit is  $\text{K day}^{-1}$ . Stippling indicates that the response is not statistically significant at two-tailed 5% level. Only the response in the stratosphere is plotted

increase also dominates the shortwave radiation response (Fig. 4d and f) and nearly balances the dynamical heating rate response (Fig. 4b). The longwave radiation response to SWV is caused by changes in stratospheric temperature and SWV. Comparing Fig. 4d to Fig. 3b reveals that the spatial pattern of the longwave radiation response is determined by temperature changes. Regions of enhanced cooling in Fig. 3b, such as tropical stratosphere and Antarctic lower to middle stratosphere, have longwave heating, whereas regions of warming or reduced cooling, such as the subtropics and Arctic upper stratosphere, show longwave cooling. Increases in SWV cause shortwave heating in the lower stratosphere, although the magnitude of shortwave radiation response is smaller than that in the longwave radiation.

A robust stratospheric dynamical response to increasing  $\text{CO}_2$  concentration is the acceleration of the Brewer-Dobson circulation (e.g., Butchart et al. 2006; Li et al. 2008). We have shown that the dynamical heating rate response in Fig. 4a and b indicates an acceleration of the Brewer-Dobson circulation. In order to quantify the impact of the SWV feedback on the Brewer-Dobson circulation, we compare the timeseries of the anomalous upward mass flux at 70 hPa, a commonly used measure of the Brewer-Dobson circulation, in Control and FixedSWV. Figure 5 shows larger increase of upward mass flux in Control than in FixedSWV throughout the  $4\times\text{CO}_2$  integration. During the last 50 years of the run, the increase in FixedSWV is 72% of that in Control. Thus, 28% of the Brewer-Dobson circulation increase in Control is attributed to the SWV increase.

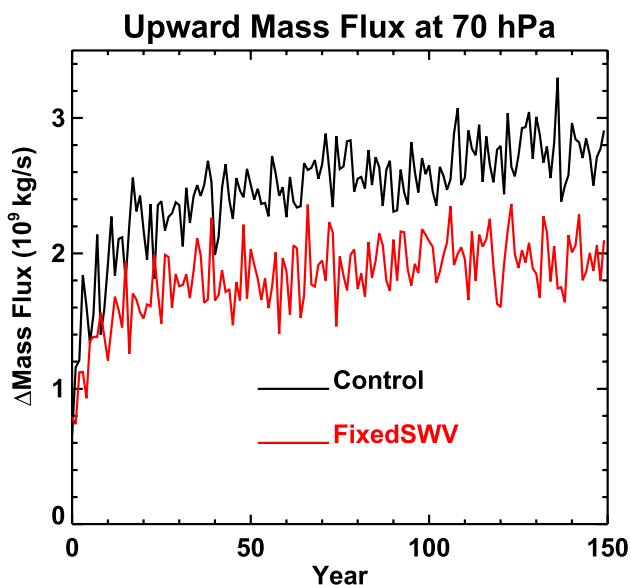
Increases in SWV significantly affect the stratospheric zonal-mean zonal wind (Fig. 6). The largest impact is in

the subtropical lower stratosphere in both hemispheres, where SWV causes a larger than  $2\text{ m s}^{-1}$  westerly increase (Fig. 6b), or about 25% of the response to  $4\times\text{CO}_2$  (Fig. 6a). The westerly increase is coherent with the increased latitudinal temperature gradient (via the thermal wind relationship) due to SWV-induced cooling in the extratropical lower stratosphere.

Our results are consistent with previous studies on the response of stratospheric temperature and circulation to SWV increase. The patterns of stratospheric temperature changes shown in Fig. 3b are broadly consistent with previous studies (Oinas et al. 2001; Forster and Shine 2002; Maycock et al. 2011, 2013), who also showed stronger cooling in the lower stratosphere, especially in the polar regions. One notable difference is the enhanced cooling in the tropical stratosphere in Fig. 3b, which is not found in the previous studies. This difference could be due to the FDH approximation used in previous studies that does not capture the tropical dynamical cooling associated with the strengthened tropical upwelling. It might also be caused by the differences in experiment setup as previous studies prescribed idealized uniform SWV increase. The stratospheric circulation response also qualitatively agrees between this study and previous studies. For example, Maycock et al. (2013) found more than  $5\text{ m s}^{-1}$  increase of the lower stratosphere subtropical zonal wind and about 10% increase of the Brewer-Dobson circulation when the SWV is uniformly increased from 3 to 6 ppmv.

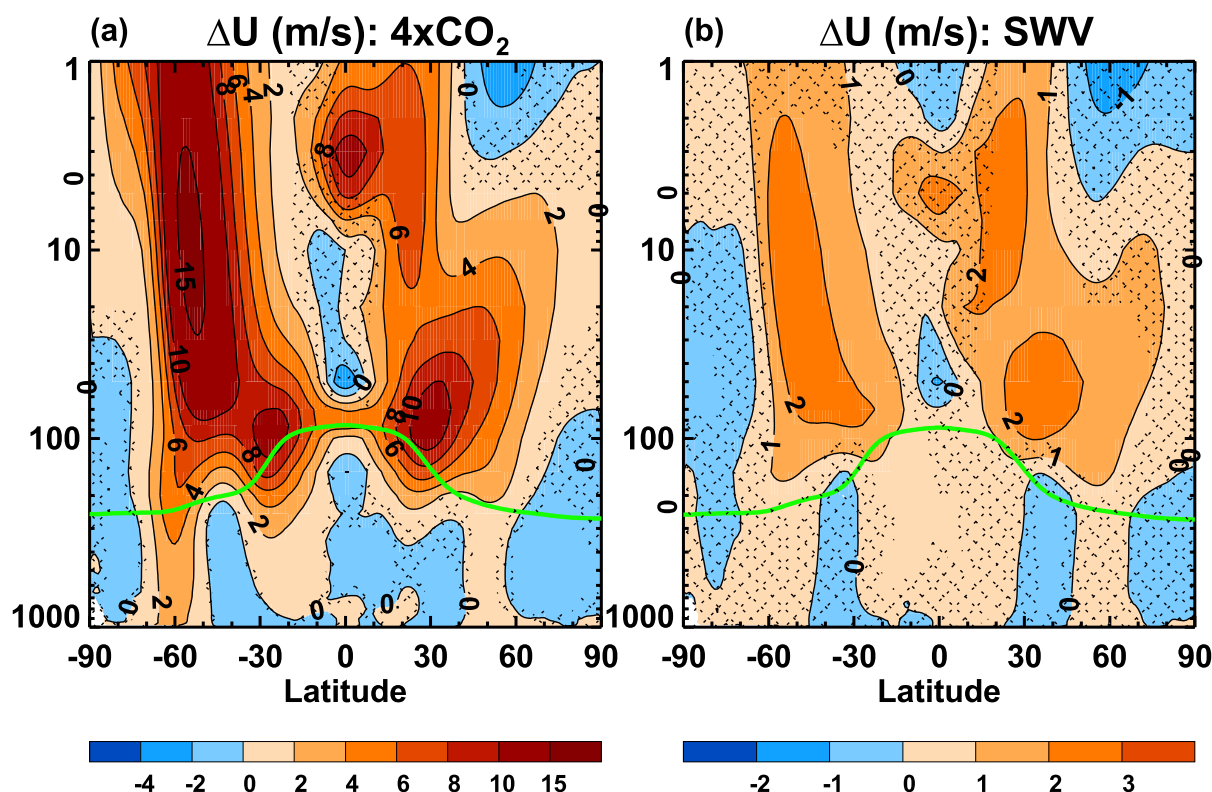
### 3.3 Tropospheric impacts of SWV feedback

We have shown in Sect. 3.1 that the SWV feedback enhances global-mean surface warming by 10%. It is of interest to find out the latitudinal structure of the surface air temperature response to the SWV feedback. Figure 7 shows the annual and zonal mean surface air temperature response to  $4\times\text{CO}_2$  (black line) and to the SWV feedback (blue line). A major feature of surface warming under  $\text{CO}_2$  quadrupling in Control is the enhanced warming in the Arctic region, i.e., Arctic amplification. The area-averaged warming in the Arctic ( $60^\circ\text{N}$ – $90^\circ\text{N}$ ) is 12.1 K, more than twice the global-mean surface warming of 5.6 K. The SWV increase amplifies this enhanced Arctic warming, with Arctic surface air temperature increase (1.7 K) three time larger than the global-mean surface air temperature increase (0.5 K). Thus, SWV feedback is an important Arctic amplification mechanism, accounting for 14% of Arctic warming under  $\text{CO}_2$  quadrupling. Note that the SWV feedback does not cause significant surface air temperature change in the Antarctic region. It is not clear why the Antarctic surface air temperature is not affected by the SWV feedback, although the much smaller SWV increase over the Antarctic (Fig. 1c) might partly explain the weak surface temperature response.



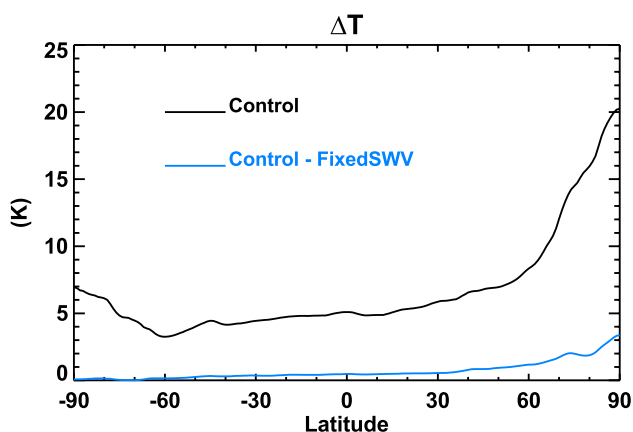
**Fig. 5** Timeseries of the anomalous upward mass flux at 70 hPa for control (black) and FixedSWV (red)





**Fig. 6** **a** The annual and zonal mean zonal wind response to  $4\times\text{CO}_2$  in Control. **b** The impact of the stratospheric water vapor feedback on the annual and zonal mean zonal wind. The green line is the tropo-

pause in Control. Stippling indicates that the response is not statistically significant at two-tailed 5% level

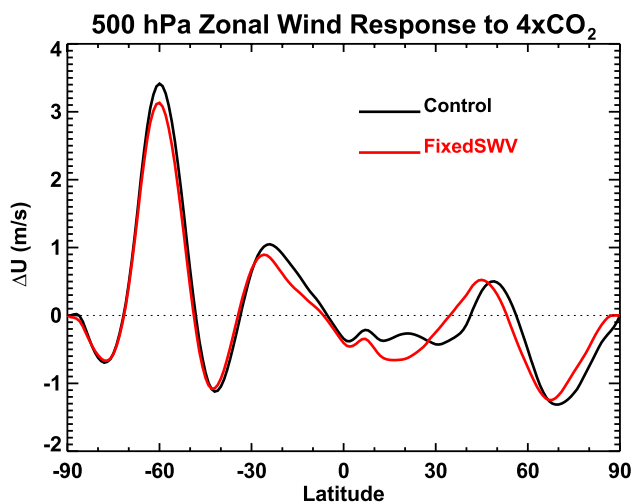


**Fig. 7** The black line is the zonal mean surface air temperature response to  $4\times\text{CO}_2$  in Control. The blue line is the response caused by the stratospheric water vapor feedback

Previous idealized SWV perturbation studies have found significant impact on the tropospheric extratropical jet (Tandon et al. 2011; Maycock et al. 2013). Figure 6 shows that the tropospheric zonal-mean zonal wind changes have similar patterns in response to SWV increase and  $4\times\text{CO}_2$ . The changes are characterized by a dipole structure in the

midlatitude in both hemispheres, indicating a poleward shift of the midlatitude westerly jet. However, the magnitudes of the response for the two cases are very different in the SH midlatitude. In order to investigate the impact of the SWV increase on the tropospheric jet in more detail, we plot in Fig. 8 the latitudinal distribution of the 500 hPa zonal-mean zonal wind response in Control (black) and FixedSWV (red). In the SH, Control and FixedSWV have similar 500 hPa zonal wind response. The magnitudes of the maximum westerly and easterly response in the SH midlatitude are slightly larger in Control than in FixedSWV, but the difference is not statistically significant. Thus, the SWV feedback does not significantly impact the SH midlatitude jet. In the NH midlatitude, the 500 hPa zonal wind response has different latitudinal structure between Control and FixedSWV: the westerly and easterly dipole response in Control is shifted poleward to that in FixedSWV. The latitude of maximum westerly increase shifts from  $45^\circ\text{N}$  in FixedSWV to  $49^\circ\text{N}$  in Control and this latitudinal shift is statistically significant. Therefore, the SWV feedback causes a poleward shift of the NH midlatitude jet.

Our results are qualitatively consistent with Maycock et al. (2013), but not quantitatively, particularly in the SH. We show that SWV increase does not significantly affect the



**Fig. 8** The response of the annual and zonal mean zonal wind at 500 hPa to  $4\times\text{CO}_2$  in Control (black) and FixedSWV (red)

SH tropospheric midlatitude jet, but Maycock et al. (2013) found that doubling SWV from 3 to 6 ppmv causes more than  $2\text{ m s}^{-1}$  westerly increase in the SH midlatitude. There are a few reasons that may explain the differences. First, different types of models are used. This study uses a coupled atmosphere–ocean model and Maycock et al. (2013) used an atmosphere-only model that does not account for changes in surface temperature caused by the SWV feedback. Changes in surface temperature gradient could have large impact on the extratropical jet (Chen et al., 2010). Second, the experiment setups are different. Maycock et al. (2013) uniformly doubled SWV from 3 to 6 ppmv. Our experimental design reflects SWV changes associated with abrupt  $\text{CO}_2$  quadrupling. The mean SWV increase is 3.8 ppmv and has large vertical gradient and hemispheric asymmetry in the lower stratosphere in this study. Third, the stratospheric temperature response is different. The stratospheric cooling in this study is weaker in the polar region and stronger in the tropics than reported in Maycock et al. (2013). Tandon et al. (2011) found that while cooling in the extratropical lower stratosphere causes a poleward shift and strengthening of the tropospheric jet, cooling in the tropical stratosphere produces opposite changes of the tropospheric jet.

## 4 Discussion and conclusions

The stratospheric water vapor feedback and its climate impacts are investigated with the coupled atmosphere–ocean GEOS CCM. A feedback suppression method is used to quantify the SWV feedback. Two abrupt  $\text{CO}_2$  quadrupling experiments, Control and FixedSWV, were conducted. Control has the SWV feedback, but FixedSWV strongly suppresses the SWV feedback by prescribing SWV to a

climatological field. The SWV feedback and its impact on stratospheric and tropospheric temperature and circulation are assessed by the differences between Control and FixedSWV.

The results show that SWV is an important feedback mechanism in GEOS CCM. The SWV feedback changes the net climate feedback parameter in GEOS CCM from  $-1.11$  to  $-1.00\text{ W m}^{-2}\text{ K}^{-1}$ , i.e., a 10% decrease in magnitude. It is responsible for 0.66 K, or 10%, of global-mean surface warming under abrupt  $\text{CO}_2$  quadrupling. Note that our estimated SWV feedback parameter of  $0.11\text{ W m}^{-2}\text{ K}^{-1}$  should be considered as a lower limit, because our method underestimates SWV increase in the lowermost stratosphere. A rough estimation suggests our calculated SWV feedback parameter is underestimated by 29%. Using a totally different approach, our results support previous studies using the stratosphere adjusted radiative perturbation method that the SWV feedback is important. It should also be noted that a  $0.11\text{ W m}^{-2}\text{ K}^{-1}$  change in net climate feedback parameter does not necessarily lead to a 10% change in surface warming as shown in this study, because the surface temperature response also depends on the climate sensitivity (or the net climate feedback) that varies considerably among climate models. Thus, it is important to quantify the relative contribution of the SWV feedback to the net climate feedback.

The advantage of our feedback suppression method over the widely used stratosphere adjusted radiative perturbation method is that it can quantitatively assess the climate impacts of the SWV feedback on the stratosphere and troposphere. It is found that the SWV feedback is particularly important in the lowermost stratosphere. This is because although increasing  $\text{CO}_2$  and SWV both cause stratospheric cooling, their vertical structures are different. First,  $\text{CO}_2$ -induced cooling increases with height, whereas SWV-induced cooling decreases with height. Second, even though  $\text{CO}_2$ -induced cooling is much stronger than SWV-induced cooling in most of the stratosphere, they are comparable in the lowermost stratosphere. Indeed, all the cooling in the NH extratropical lowermost stratosphere and half of the cooling in the SH extratropical lowermost stratosphere under  $\text{CO}_2$  quadrupling is attributed to the SWV feedback. The SWV feedback's cooling in the extratropical lowermost stratosphere further increases the latitudinal temperature gradient in the upper troposphere and lower stratosphere. As a result, the SWV feedback causes a more than  $2\text{ m s}^{-1}$  westerly increase in the subtropical lower stratosphere, accounting for 20% of stratospheric zonal wind change in this region. The SWV feedback also contributes significantly to the acceleration of the Brewer–Dobson circulation, explaining nearly 30% of tropical upward mass flux increase.

We find that SWV increases induce a dynamical feedback that strongly modifies the stratospheric temperature response and generate localized cooling/warming patterns. Our results

clearly demonstrate the interactions between increases in SWV, stratospheric temperature, and dynamics. Thus, while the FDH method is able to reasonably capture the main patterns of stratospheric temperature response to changes in SWV and other trace gases (e.g., Maycock et al. 2011), cautions should be taken when applying the FDH approximation.

An interesting result is that the SWV feedback plays a role in Arctic amplification. The SWV feedback causes a 1.7 K Arctic warming, which is more than three times larger than the SWV induced global mean surface warming. The 1.7 K warming contributes to about 14% of the Arctic surface warming under abrupt CO<sub>2</sub> quadrupling. Although water vapor feedback has been known to be an important mechanism of Arctic amplification (Graversen and Wang 2009), our study is the first to quantify the role of the SWV feedback in Arctic amplification. It is also found that the SWV feedback does not cause statistically significant surface temperature changes in the Antarctic region. The different impacts in the Arctic and Antarctic might be related to the hemispherical differences in the SWV response to CO<sub>2</sub> quadrupling, but at this stage the mechanism is not clear.

The SWV feedback causes a significantly larger poleward shift of the NH tropospheric midlatitude jet, but it does not have significant impact on the SH tropospheric jet. Our results are not consistent with Maycock et al. (2013), who showed that doubling SWV concentrations leads to SH tropospheric jet changes an order of magnitude stronger than found in this study. The different results could be caused by differences in models, experiment setups, and stratospheric temperature responses. Further studies are needed to understand the impact of SWV feedback on the tropospheric circulation.

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**Availability of data and materials** The simulations used in this study are stored in the data storage facility of NASA Center for Climate Simulation and are fully available upon request to F. L. Code availability: The code used to analyze the model data is available upon request to F. L.

## Compliance with ethical standards

**Conflicts of interest** The authors declare no conflicts of interest.

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