https://doi.org/10.5194/acp-2019-1169
Preprint. Discussion started: 5 March 2020

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Quantifying climate feedbacks in the middle atmosphere using WACCM

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Key points:

- In a double CO_2 climate, the direct forcing of CO_2 would lead to a cooling of approximately 9 K in the middle atmosphere.

- The ozone feedback mitigates this cooling by warming the middle atmosphere by approximately 1.5 K.

- The dynamical feedback is another important feedback with large effects locally, while the effects of the water vapour feedback and especially the cloud and albedo feedbacks are small.

https://doi.org/10.5194/acp-2019-1169
Preprint. Discussion started: 5 March 2020

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Abstract

The importance of feedback processes in the middle atmosphere for surface and tropospheric climate is increasingly realized. To better understand feedback processes in response to a doubling of CO_2 we use the climate feedback response analysis method (CFRAM). We examine the middle atmosphere response to CO_2 doubling with respect to the pre-industrial state in the Whole Atmosphere Community Climate Model (WACCM). Globally, the simulated temperature decrease between 200 and 0.01 hPa (~12-80 km) is found to be -5.2 K in July and -5.5 K in January in WACCM. The CFRAM calculations show that the direct forcing of CO_2 alone would lead to an even stronger cooling of approximately 9 K in the middle atmosphere in both July and January. This cooling is being mitigated by the combined effect of the different feedback processes.
The contribution from the ozone feedback causes a warming of approximately 1.5 K, mitigating the cooling due to changes in CO_2. Changes in CO_2 also lead to changes in the middle atmosphere dynamics. The changes in dynamics play a large role locally, especially above 0.1 hPa. Other feedback processes, which are known to be important in the tropospheric and surface climate, such as the water vapor, albedo and cloud feedbacks are of minor importance in the middle atmosphere,although some effects are seen in the stratosphere, mainly through the responses to sea surface temperature and sea ice changes. It should be noted that there is a relatively large error term associated with the current method in the middle atmosphere, which can be explained by the linearization in the method.

1. Introduction

The middle atmosphere is the region of the atmosphere that encompassed the stratosphere, where the temperature increases with height, from 10-50 km and the mesosphere, where the temperature decreases with height, from about 50-90 km. A classic study by Manabe and Strickler (1964) shows that in the troposphere, water vapour is the dominant greenhouse gas, followed by CO_2. Ozone is responsible for the existence of the stratosphere and the reversal of the temperature gradient in the stratosphere.

The middle atmosphere is generally controlled by radiative processes although there are important exceptions. The temperature in the winter stratosphere for example, is significantly greater than what is expected from radiative equilibrium considerations alone. This results from adiabatic heating associated with downwelling at higher latitudes as a part of the Brewer-Dobson circulation (Brewer, 1949, Dobson, 1956).

Many chemical, physical and dynamical processes in the middle atmosphere are still often overlooked in climate model simulations. This can be noticed from the description of the experimental design in model intercomparison projects as in e.g. *Kageyama et al.* (2017) and *Taylor et al.* (2012). However, recently, there have been a number of studies that show the importance of the middle atmosphere for the surface and tropospheric climate. It has, for

https://doi.org/10.5194/acp-2019-1169
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example, been shown that cold winters in Siberia are linked to changes in the stratospheric circulation (*Zhang et al.*, 2018).

Other studies show that the representation of ozone in climate models affects climate change projections (*Dietmüller et al.*, 2014; *Noda et al.,* 2017; *Nowack et al.*, 2015, 2018;). It has been found that the ozone representation can result in up to a 20% difference in simulated global mean surface warming (*Nowack et al*., 2015), although the exact importance of changes in ozone seems to be dependent on both the model and the scenario (*Nowack et al*., 2015) and is not found by all studies (*Marsh et al.*, 2016).

Neglecting the effect of ozone may lead to a significant underestimation of the temperature during past climate if we compare with available paleoclimate reconstructions (*Ljungqvist et al*, 2019), and may also alter the future climate projections in different CO_2 scenarios. As the effect is found to be rather large in some studies, and absent in other, there is a need for a better understanding of the behaviour of the middle atmosphere in response to changing CO_2 conditions, as the ozone concentration is influenced by this.

Ozone is an example of a climate feedback, a process that changes in response to a change in CO_2-concentration and in turn dampens or amplifies the climate response to the CO_2 perturbation. These climate feedbacks are a challenging subject of study, as observed climate variations might not be in equilibrium, multiple processes are operating at the same time and moreover the geographical structures and timescales of different forcings differ. However, feedbacks form a crucial part of understanding the response of the atmosphere to changes in the CO_2-concentration.

Various methods have been developed to study these feedbacks, such as the partial radiative perturbation (PRP) method, the online feedback suppression approach and the radiative kernel method (*Bony et al.,* 2006 and the references therein). These methods study the origin of the global cimate sensitivity (Soden and Held, 2006; Caldwell et al., 2016; Rieger et al., 2017). The focus of these methods is on changes in the global mean surface temperature, global mean surface heat and global mean sensible heat fluxes (*Ramaswamy et al., 2019*).

These methods are powerful for this purpose, however, they are not suitable to explain temperature changes on spatially limited domains. They neglect non-radiative interactions between feedback processes and they only account for feedbacks that directly affect the radiation at the top of the atmosphere (TOA).

Recently, a new climate feedback analysis method has been developed, which is an offline diagnostic tool: the climate feedback-response analysis method (CFRAM). With this method, one can calculate the temperature changes on a spatially limited domain: it is possible to isolate partial temperature changes due to an external forcing and internal feedbacks in the atmosphere. The partial temperature changes are linearly addable; their sum is equal to the total temperature change (*Cai and Lu, 2009, Lu and Cai*,

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2009).

CFRAM takes into account that the climate change is not only determined by the energy exchange between the Earth's system and outer space, but is also influenced by the energy flow within the Earth's system itself. The method is based on the energy balance in an atmosphere-surface column.It solves the linearized infrared radiation transfer model for the individual energy flux perturbations.

As a practical diagnostic tool to analyse the role of various forcing and feedback, CFRAM has been used widely in climate change research on studying surface climate change (*Taylor et al.,* 2013; *Song and Zhang,* 2014; *Hu et al.,* 2017; *Zheng et al.,* 2019).

This method has been applied to the middle atmosphere climate sensitivity study as well (*Zhu et al.,* 2016). In that study, the CFRAM method has been adapted and applied to both model output, as well as observations. They study the atmosphere responses during solar maximum and minimum and find that the variation in solar flux forms the largest radiative component. They show that increasing CO_2 cools the middle atmosphere, while the changes in O_3-concentration lead to a cooling in some regions of the middle atmosphere and to a warming in others.

In the present work, we apply CFRAM to climate sensitivity experiments performed by the Whole Atmosphere Community Climate Model (WACCM), which is a high-top global climate system model, including the full middle atmosphere chemistry. We investigate the effects of doubling the CO_2-concentration and the accompanying sea surface temperature changes on the temperature changes in the middle atmosphere as compared to the pre-industrial state. We discuss the total responses and feedbacks, as well as those that are induced by doubling CO_2 and changes in the sea surface temperature and sea ice distribution separately.

2. The model and methods

2.1 Model description

The Whole Atmosphere Community Model (WACCM) is a chemistry-climate model, which spans the range of altitudes from the Earth's surface to about 140 km (*Marsh et al.,* 2013). The model consists of 66 vertical levels with irregular resolution, ~1.1 km in the troposphere above the boundary layer, 1.1–1.4 km in the lower stratosphere, 1.75 km at the stratosphere and 3.5 km above 65 km. The horizontal resolution is 1.9° latitude by 2.5° longitude.

WACCM is a superset of the Community Atmospheric Model version 4 (CAM4) developed at the National Center for Atmospheric Research (NCAR). Therefore, WACCM includes all the physical parameterizations of CAM4 (*Neale et al.,* 2013), and a well-resolved high-top middle atmosphere. The orographic gravity wave (GW) parameterization is based on *McFarlane* (1987). WACCM also includes parameterized non-orographic GWs, which are

https://doi.org/10.5194/acp-2019-1169
Preprint. Discussion started: 5 March 2020

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generated by frontal systems and convection (*Richter et al.,* 2010). The parameterization of non-orographic GW propagation is based on the formulation by *Lindzen* (1981).

The chemistry in WACCM is based on version 3 of the Model for Ozone and Related Chemical Tracers (MOZART). This model represents chemical and physical processes from the troposphere until the lower thermosphere. (*Kinnison et al.,* 2007). In addition, WACCM simulates chemical heating, molecular diffusion and ionization and gravity wave drag.

2.2 Experimental set-up

In this study, we first perform a simulation under pre-industrial conditions and take this experiment as a control run, forced with pre-industrial ocean surface conditions such as sea surface temperature and sea ice (referred to SSTs from now on), see table 1. The pre-industrial CO_2-concentration is set as 280 ppm, the SSTs are from the CMIP5 pre-industrial control simulation by the fully coupled earth system model CESM (Hurrel et al., 2013). The atmosphere component of CESM is the same as WACCM, but does not include stratospheric chemistry.

We also run a perturbation experiment by doubling the CO_2 concentration to 560 ppm from pre-industrial level. In WACCM, the CO_2-concentration does not double everywhere in the atmosphere. Only the surface level CO_2 mixing ratio is doubled, and elsewhere in the atmosphere is calculated according to WACCM's chemical model. For the double CO_2 simulation, we run two experiments by using two SSTs forcing, one is keeping the pre-industrial SSTs unchanged, and another one from a double CO_2 equilibrium simulation by CESM. To investigate the effect of SSTs, we further run a simulation for only using the SST forcing from the coupled CESM for double CO_2 condition, but keep the CO_2-concentration as 280 ppm. All the simulations are run for 50 years, of which the last 40 years are used for analysis.

Table 1. Set-up of the model experiments.

Experiment	**CO_2**	**SSTs from CESM equilibrum run**
Pre-industrial	280 ppm	PI control
CO_2 and SSTs high	560 ppm	Double CO2 run
CO_2 high	560 ppm	PI control
SSTs high	280 ppm	Double CO2 run

2.3 Climate feedback-response analysis method (CFRAM)

In this study, we aim to understand and quantify the different climate feedbacks that may play a role in the middle atmosphere in a high CO_2 climate. For this purpose, we apply a climate feedback-response analysis method (CFRAM).

As briefly discussed in the introduction, traditional methods to study climate feedbacks are based on the energy balance at the top of the atmosphere

(TOA). This means that the only climate feedbacks that are taken into account are those that effect the radiative balance at the TOA. However, there are other thermodynamic and dynamical processes that do not directly affect the TOA energy balance, while they do yield a temperature response in the atmosphere.

Contrary to TOA-based methods, CFRAM considers all the radiative and non-radiative feedbacks that result from the climate system due to response to an external forcing. This means that CFRAM starts from a slightly different definition of a feedback process. Note also that as the changes in temperature are calculated simultaneously, the vertical mean temperature or lapse rate feedback per definition do not exist in CFRAM.

Another advantage of CFRAM is that it allows for measuring the magnitude of a certain feedback in units of temperature. We can actually calculate how much of the temperature change is due to which process. The *'climate response'* in the name of this method refers to the changes in temperature in response to the climate forcings and climate feedbacks.

The mathematical formulation of CFRAM is based on the conservation of total energy. At a given location in the atmosphere, the energy balance in an atmosphere-surface column can be written as:

$$\vec{R} = \vec{S} + \vec{Q}^{conv} + \vec{Q}^{turb} - \vec{D}^{v} - \vec{D}^{h} + \vec{W}^{fric} \tag{1}$$

$\vec{R}$ represents the vertical profile of the net long-wave radiation emitted by the different layers in the atmosphere and surface. $\vec{S}$ is the vertical profile of the solar radiation, which is absorbed by these layers. $\vec{Q}^{turb}$ is the convergence of total energy in each layer due to turbulent motions, $\vec{Q}^{conv}$ is convergence of total energy into the layers due to convective motion. $\vec{D}^{v}$ is the large-scale vertical transport of energy from different layers to others. $\vec{D}^{h}$ is the large-scale horizontal transport within the layers and $\vec{W}^{fric}$ is the work done by atmospheric friction.

Due to an external forcing (in this study, the change in CO_2-concentration and/or change in SSTs), the difference in the energy flux terms then becomes:

$$\Delta\vec{R} = \Delta\vec{F}^{ext} + \Delta\vec{S} + \Delta\vec{Q}^{conv} + \Delta\vec{Q}^{turb} - \Delta\vec{D}^{v} - \Delta\vec{D}^{h} + \Delta\vec{W}^{fric} \tag{2}$$

In which the delta (Δ) stands for the difference between the perturbation run and the control run.

CFRAM takes advantage of the fact that the infrared radiation is directly related to the temperatures in the entire column. The temperature changes in the equilibrium response to perturbations in the energy flux terms can be calculated. This is done by requiring that the temperature-induced changes in infrared radiation to balance the non-temperature induced energy flux

https://doi.org/10.5194/acp-2019-1169
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perturbations.

Equation (2) can also be written as:

$$\Delta\left(\vec{S}-\vec{R}\right)_{total} + \Delta dyn = 0 \tag{3}$$

The term $\Delta(\vec{S}-\vec{R})$ we can calculate as the longwave heating rate and the solar heating rate are output variables of the model simulations. We then calculate the difference in these heating rates for the perturbation simulation and the control simulation.

We use the term $\Delta\left(\vec{S}-\vec{R}\right)_{total}$ to calculate the dynamics term Δdyn.

$$\Delta dyn = -\Delta\left(\vec{S}-\vec{R}\right)_{total} \tag{4}$$

WACCM provides us with a heating rate in Ks^{-1}. For the CFRAM calculations, we need the energy flux in Wm^{-2}. We can calculate the energy flux by multiplying with the mass of different layers in the atmosphere and the specific heat capacity.

$$\Delta(\vec{S}-\vec{R}) = \Delta(\vec{S}-\vec{R})_{(WACCM)} * mass_k * c_p \tag{5}$$

with $mass_k = \frac{p_{k+1}-p_k}{g}$ with p in Pa and $c_p = 1004 \ \ J\,kg^{-1}\,K^{-1}$ the specific heat capacity at constant pressure.

WACCM includes a non-local thermal equilibrium (non-LTE) radiation scheme above 50 km. It consists of a long-wave radiation (LW) part and a short-wave radiation (SW) part which includes the extreme ultraviolet (EUV) heating rate, chemical potential heating rate, CO_2 near-infrared (NIR) heating rate, total auroral heating rate and non-EUV photolysis heating rate.

Therefore, we split the term $\Delta\left(\vec{S}-\vec{R}\right)_{total}$ in an LTE and a non-LTE term:

$$\Delta\left(\vec{S}-\vec{R}\right)_{total} = \Delta(\vec{S}-\vec{R})_{LTE} + \Delta(\vec{S}-\vec{R})_{non-LTE} \tag{6}$$

WACCM provides us with the total longwave heating rate as well as the total solar heating rate and the non-LTE longwave and shortwave heating rates for the different runs. This means that we can calculate the term $\Delta(\vec{S}-\vec{R})_{non-LTE}$ as well, where we again need to convert our result from Ks^{-1} to Wm^{-2}:

$$\Delta(\vec{S}-\vec{R})_{non-LTE} = \Delta(\vec{S}-\vec{R})_{non-LTE(WACCM)} \ mass_k * c_p \tag{7}$$

This term can be inserted in equation (3):

$$\Delta(\vec{S}-\vec{R})_{LTE} + \Delta(\vec{S}-\vec{R})_{non-LTE} + \Delta dyn = 0 \tag{8}$$

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The central step in CFRAM is to decompose the radiative flux vector, using a a linear approximation.

We start by decomposing the LTE infrared radiative flux vector $\Delta\vec{R}$

$$\Delta\vec{R}_{LTE} = \frac{\partial\vec{R}}{\partial\vec{T}}\Delta T + \Delta\vec{R}_{CO_2} + \Delta\vec{R}_{O_3} + \Delta\vec{R}_{H_2O} + \Delta\vec{R}_{albedo} + \Delta\vec{R}_{cloud} \qquad (9)$$

where $\Delta\vec{R}_{CO_2}$, $\Delta\vec{R}_{O_3}$, $\Delta\vec{R}_{H_2O}$, $\Delta\vec{R}_{albedo}$, $\Delta\vec{R}_{cloud}$ are the changes in infrared radiative fluxes due to the changes in CO_2, ozone, water vapour, albedo and clouds, respectively.

For equation (9), we assumed that radiative perturbations can be linearized by neglecting the higher order terms of each thermodynamic feedback and the interactions between these feedbacks. This is also commonly done in the PRP method (*Bony et al.,* 2006).

The term $\frac{\partial\vec{R}}{\partial\vec{T}}\Delta T$ represents the changes in the IR radiative fluxes related to the temperature changes in the entire atmosphere-surface column. The matrix $\frac{\partial\vec{R}}{\partial\vec{T}}$ is the Planck feedback matrix, in which the vertical profiles of the changes in the divergence of radiative energy fluxes due to a temperature change are represented.

We calculate this feedback matrix using the output variables of the perturbation and the control run of WACCM and inserting these in the CFRAM radiation code: atmospheric temperature, surface temperature, reference height temperature, ozone, surface pressure, solar insolation, downwelling solar flux at the surface, net solar flux at the surface, dew point temperature, cloud fraction, cloud ice amount, cloud liquid amount, ozone and specific humidity.

Similarly, the changes in the LTE shortwave radiation flux can be written as the sum of the change in shortwave radiation flux due to the direct forcing of CO_2 and the different feedbacks:

$$\Delta\vec{S}_{LTE} = \Delta\vec{S}_{CO_2} + \Delta\vec{S}_{O_3} + \Delta\vec{S}_{H_2O} + \Delta\vec{S}_{albedo} + \Delta\vec{S}_{cloud} \qquad (10)$$

Similarly, to equation (9), we perform a linearization.

Substituting (9) and (10) in equation (8) yields:

$$\Delta(\vec{S}-\vec{R})_{CO_2} + \Delta(\vec{S}-\vec{R})_{O_3} + \Delta(\vec{S}-\vec{R})_{H_2O} + \Delta(\vec{S}-\vec{R})_{albedo} + \Delta(\vec{S}-\vec{R})_{cloud} - \frac{\partial\vec{R}}{\partial\vec{T}}\Delta T$$
$$+\Delta(\vec{S}-\vec{R})_{non-LTE} + \Delta dyn = 0 \qquad (11)$$

This can be written as:

$$\frac{\partial\vec{R}}{\partial\vec{T}}\Delta T = +\Delta(\vec{S}-\vec{R})_{non-LTE} + \Delta dyn \qquad (12)$$

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Equation (12) can be solved for the individual temperature perturbations. We calculate the temperature changes due to the direct effect of CO_2 as well as the different feedback processes.

$$\Delta T_{CO_2} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{CO_2} \tag{13}$$

This can be done in a similar way for the different feedback processes:

$$\Delta T_{O_3} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{O_3} \tag{14}$$

$$\Delta T_{H_2O} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{H_2O} \tag{15}$$

$$\Delta T_{albedo} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{albedo} \tag{16}$$

$$\Delta T_{cloud} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{cloud} \tag{17}$$

The factors $\Delta\left(\vec{S} - \vec{R}\right)_{CO_2}$, $\Delta\left(\vec{S} - \vec{R}\right)_{O_3}$, $\Delta\left(\vec{S} - \vec{R}\right)_{H_2O}$, $\Delta\left(\vec{S} - \vec{R}\right)_{albedo}$ and $\Delta\left(\vec{S} - \vec{R}\right)_{cloud}$ are calculated by inserting the output variables from WACCM in the radiation code of CFRAM. Here, one takes the output variables from the control run, apart from the variable that is related to the direct forcing or the feedback.

This means that for the direct forcing of CO_2, one takes the CO_2 from the perturbation run, while one takes the other variables from the control run. For the ozone feedback, one takes the ozone from the perturbation run. For the water vapour feedback, one takes the specific humidity, surface pressure, surface temperature and dew point temperature. While for the albedo feedback, one takes the downwelling solar flux at surface and net solar flux at surface from the perturbation run and the other variables from the control run. For the cloud feedback, one takes the cloud fraction, cloud ice and cloud liquid amount from the perturbation run. For all these feedbacks, one takes the other variables from the control run.

Similarly to equations (13)-(17), we also calculate the temperature change due to non-LTE processes and the dynamical feedback. We already calculated the terms $\Delta\left(\vec{S} - \vec{R}\right)_{non-LTE}$ and Δdyn in (4) and (7).

$$\Delta T_{non-LTE} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{non-LTE} \tag{18}$$

$$\Delta T_{dyn} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta dyn \tag{19}$$

The calculated partial temperature changes can be added, their sum being

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equal to the total temperature change. It is important to note that this does not mean that the individual processes are physically independent of each other.

$$\Delta T_{CFRAM} = +\Delta T_{O_3} + \Delta T_{H_2O} + \Delta T_{albedo} + \Delta T_{cloud} + \Delta T_{non-LTE} + \Delta T_{dyn} \tag{20}$$

The linearization done for equations (9) and (10) introduces an error between the temperature difference as calculated by CFRAM and as seen in the model output. Another source of error is that the radiation code of the CFRAM calculations is not exactly equal to the radiation code of WACCM.

$$\Delta T_{CFRAM} = \Delta T_{WACCM} - \Delta T_{error} \tag{21}$$

For more details on the CFRAM method, please refer to *Lu and Cai* (2009).

Note that the method used in this study differs slightly from the Middle Atmosphere Climate Feedback Response Analysis Method (MCFRAM) used by *Zhu et al.* (2016). The major difference is that in this study, we perform the calculations using the units of energy fluxes (Wm^{-2}) instead of converting to heating rates (Ks^{-1}). Another difference is that our method is not applicable above 0.01 hPa (~80 km), while *Zhu et al.* (2016) added molecular thermal conduction to the energy equation, to perform the calculations beyond the mesopause.

3. Results

3.1 Temperature responses in a double CO_2 scenario

In section 2.2, it was discussed that four experiments were performed with WACCM: a simulation with pre-industrial conditions, a simulation with changed SSTs only, a simulation with only a changed CO_2-concentration and a final simulation with both changed SSTs and CO_2-concentration.

Figure 1 shows the temperature changes for the different experiments with respect to the pre-industrial state, as modelled by WACCM. As shown in earlier studies, we observe that an increase in CO_2 causes a cooling in the middle atmosphere with the exception of the cold summer upper mesosphere region (*Akmaev*, 2006). We also observe that changing the SSTs alone, while leaving the CO_2-concentration at the pre-industrial levels (Fig 1c and 1f) also yields significant temperature changes over a large part of the middle atmosphere, and contributes to the observed warming in the cold summer mesopause region.

In line with *Fomichev et al.* (2007) and *Schmidt et al.* (2006), we find that the sum of the two separate temperature changes in the experiment with changed CO_2 only and changed SSTs only is approximately equal to the changes observed in the combined simulation. *Shepherd* (2008) has explained this phenomenon as follows: climate change affects the middle atmosphere in two ways: either radiatively through in situ changes associated with changes in CO_2 or dynamically through changes in stratospheric wave forcing, which are primarily a result of changing the SSTs (*Shepherd,* 2008). Even though the

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radiative and dynamic processes are not independent, at first approximation, these processes are seen to be additive (*Sigmond et al.*, 2004, *Schmidt et al.,* 2006, *Fomichev et al.,* 2007).

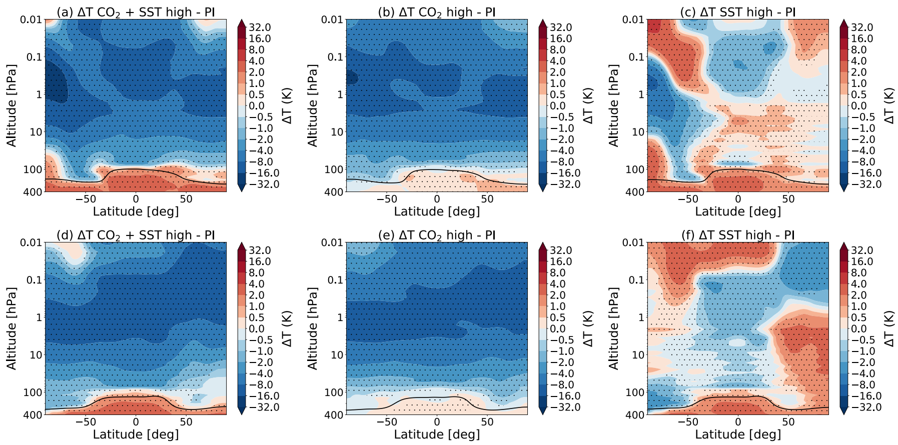

Figure 1: The total change in temperature in July (top) and January (bottom) for (a,d) the simulation with high CO_2 and SSTs, (b,e) the simulation with high CO_2, (c,f) the simulation with high SSTs, all as compared the pre-industrial control simulation. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

The CFRAM makes it possible to separate and estimate the temperature responses due to an external forcing and various climate feedbacks, such as ozone, water vapour, cloud, albedo and dynamical feedbacks. Note that for the ozone, water vapour, cloud and albedo feedback, we can only calculate the radiative part of the feedback.

Because we are using an atmosphere-only model, in our experiment, the external forcing is either the change in CO_2-concentration or the change in SSTs or both. In an atmosphere-ocean model (such as CESM) and, of course, in reality the changes in sea surface temperature and sea ice distributions are responses to the changed CO_2-concentration.

Figure 2 shows the average change in global mean temperature for the middle atmosphere between 200 and 0.01 hPa (12-80 km) for the experiment with the changed CO_2-concentration and changed SSTs. The 'total'-column shows the temperature changes in WACCM. It is found that the average change in global mean temperature for the middle atmosphere between 200 and 0.01 hPa for this experiment is approximately -5.5 K in July and -5.2 K in January in WACCM.

Figure 2 also shows the radiative feedbacks due to ozone, water vapour, clouds, albedo and the dynamical feedback, as well as the small contribution

https://doi.org/10.5194/acp-2019-1169
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due to the Non-LTE processes, as calculated by CFRAM. The column 'error' shows the difference between temperature change in WACCM and the sum of the calculated temperature responses in CFRAM. In sections 3.3-3.6, we will discuss the different feedbacks separately in more detail, at this point we give an overview of the general effects and relative importance of the different feedback processes.

It can be seen that the direct forcing of the changed CO_2-concentration alone, as calculated by CFRAM, would lead to an even stronger cooling in the middle atmosphere. Globally, the cooling would be approximately 9 K in both July and January. The cooling is being mitigated by the combined effect of the different feedbacks.

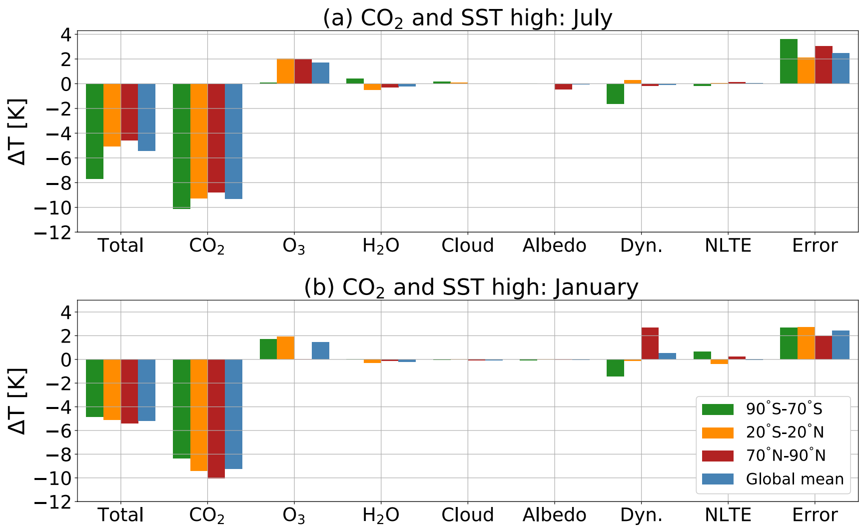

Figure 2: The mean temperature responses to the changes in CO_2 and various feedback processes in July (top) and January (bottom) for the experiment with double CO_2 and changed SSTs in the atmosphere between 200 and 0.01 hPa, for polar regions (90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) and the global mean, for experiment with double CO_2 and changed SSTs.

Figure 2 shows that ozone feedback causes significant warming, by heating the middle atmosphere globally by approximately 1.5 K in both July and January, and thus counteracting the cooling effect due to the direct forcing of CO_2. It has been suggested in earlier studies that the ozone response to changes in the CO_2-concentration will yield a radiative feedback that will mitigate the cooling, which is due to the direct forcing of CO_2 (*Jonsson et al.,* 2004). With CFRAM, it is possible to quantify this effect and to compare it with the effects of other feedbacks in the middle atmosphere.

The temperature response due to dynamical feedbacks is small in global average. This can be understood as waves generally do not generate momentum and heat, but redistribute these instead (*Zhu et al.,* 2016).

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However, the local responses to dynamical changes in the high latitudes are large, as we will see in section 3.3.

While clouds, albedo and water vapour feedbacks are of crucial importance for the tropospheric and surface climate, in the middle atmosphere their role is very small globally. The feedback due to water vapour shows some contribution locally, as will be discussed in section 3.6. There are some small temperature responses due to non-LTE effects as well.

The error term is relatively large. In CFRAM, we assumed that the radiative perturbations can be linearized by neglecting the higher order terms of each thermodynamic feedback and the interactions between these feedbacks, this yields an error. Cai and Lu (2009) show that this error is larger in the middle atmosphere than for similar calculations in the troposphere. In the middle atmosphere, the density of the atmosphere is smaller, which leads to smaller numerical values of the diagonal elements of the Planck feedback matrix. As a result, the linear solution is very sensitive to forcing in the middle atmosphere. Another part of the error is due to the fact that the radiative scheme. The scheme we use in CFRAM is relatively simple, as compared to the one that is used in WACCM.

Figure 2 does not only show the global mean temperature responses, but also those in the polar regions and the tropics to give a more complete picture. It can be seen that the total temperature response in WACCM in the winter polar region (-7.7 K) is much stronger than in the summer polar region (-4.6 K) during July, resulting in 3.1 K difference. In January, there is also a larger temperature response in the winter polar region as compared to the summer polar region, although the difference in the temperature response is only 0.6K.

It is observed that the total temperature change in the polar regions during the local winter is stronger than that in local summer. At the same time, the cooling due to the direct forcing of CO_2 alone is stronger in the summer polar region than in the winter polar region. The strong CO_2 cooling effect in the summer polar region is mitigated by the warming caused by the ozone feedback.

In addition, the vertical profile of the temperature response to various feedbacks is shown in Figure 3. Here, one can see that the increase in CO_2 leads to a cooling over almost the whole middle atmosphere: an effect that increases with height. We also observe that in the summer upper mesosphere regions, the increased CO_2-concentration leads to a warming. Ozone leads to a warming almost everywhere in the atmosphere. In some places, this warming exceeds 5 K. In the polar winter the effect of ozone is small due to lack of sunlight.

There is also a relatively large temperature response to the changes in dynamics. In Fig. 3, it can be seen that there is a cooling in the summer mesosphere, while there is warming in the winter mesosphere. The water vapour, cloud and albedo feedback play only a very small role in the middle atmosphere, as we observed in Figure 2. The Non-LTE effects are also small,

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but start to play a small role above 0.1 hPa, the exact mechanism of which is outside the scope of this paper.

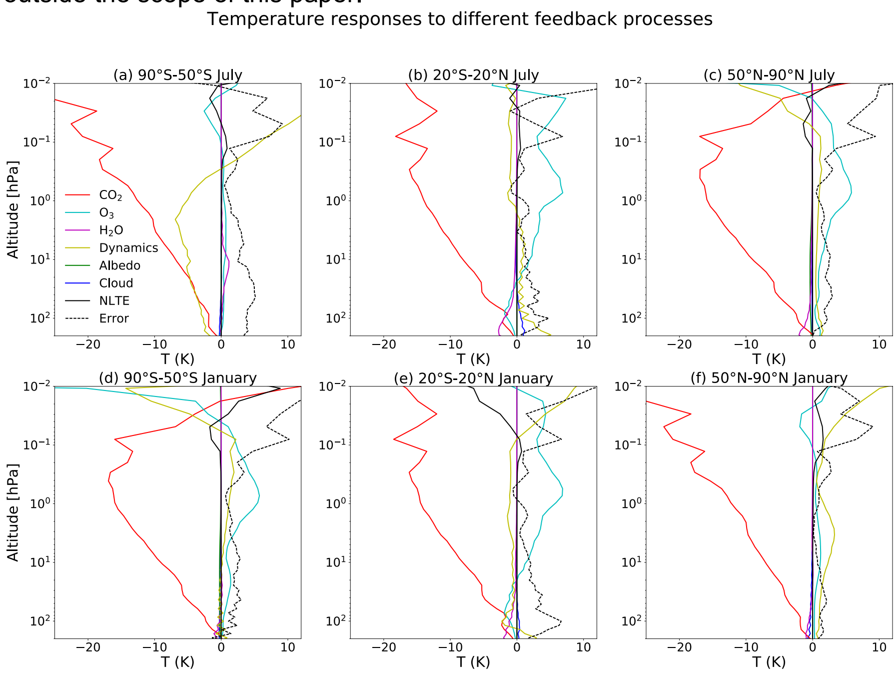

Figure 3: Vertical profiles of the temperature responses to the changes in CO_2 and various feedback processes in July (top) and January (bottom) for the experiment with double CO_2 and changed SSTs in the atmosphere between 200 and 0.01 hPa, for regions from 50° N/S polewards and the tropics (20°S-20°N), as calculated by CFRAM.

Figure 4 shows temperature responses for the experiment with double CO_2 (a,b) and changed SSTs (c,d) separately. Again, the 'total'-column shows the temperature changes as found by WACCM, the columns CO_2, O_3, H_2O, cloud, albedo, dynamics, Non-LTE shows the temperature responses as calculated by CFRAM. Error shows the difference between temperature change in WACCM and the sum of the calculated temperature responses in CFRAM.

The effects of the changed SSTs on the middle atmosphere are relatively small as compared to the effects of changing the CO_2. The temperature response to the water vapour feedback is, however, almost solely due to changes in the SSTs.

Earlier, we discussed that the sum of the two separate temperature changes in the experiment with double CO_2 and changed SSTs is approximately equal to the changes observed in the combined simulation. We find that the same is true for the temperature responses to the different feedback processes.

In the rest of the paper, we discuss the temperature responses to the direct forcing and the various feedbacks during July and January in further detail.

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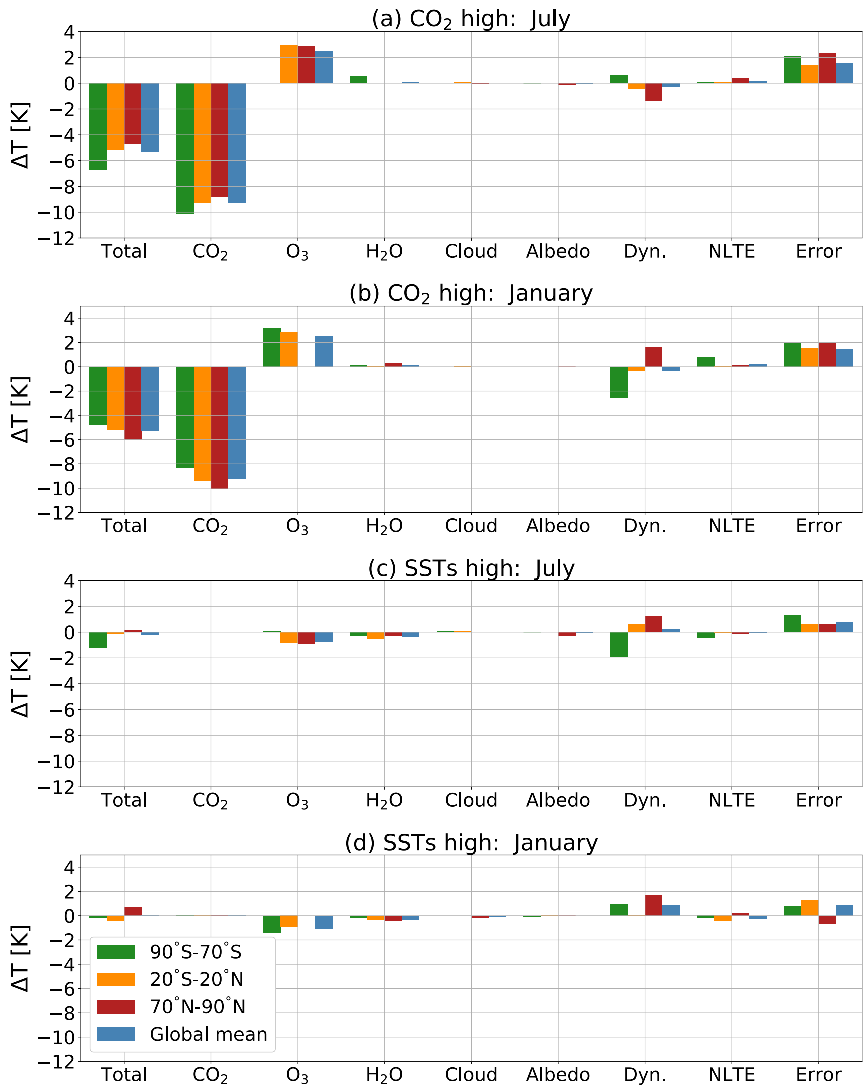

Figure 4: The mean temperature responses to the changes in CO_2 and various feedback processes in July (a,c) and January (b,d) in the atmosphere between 200 and 0.01 hPa, for polar regions (90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) and the global mean for the experiment with double CO_2 (a,b) and changed SSTs (c,d) separately.

3.2 Temperature direct response to CO_2

In Figure 5, we see that increasing CO_2 leads to a cooling almost everywhere in the middle atmosphere, except in the cold summer upper mesosphere. The higher the temperature, the more cooling there is taking place due to the increasing CO_2-concentration (*Shepherd,* 2008). The reason for this is that

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the outgoing longwave radiation strongly depends on the Planck blackbody emission (*Zhu et al.,* 2016).

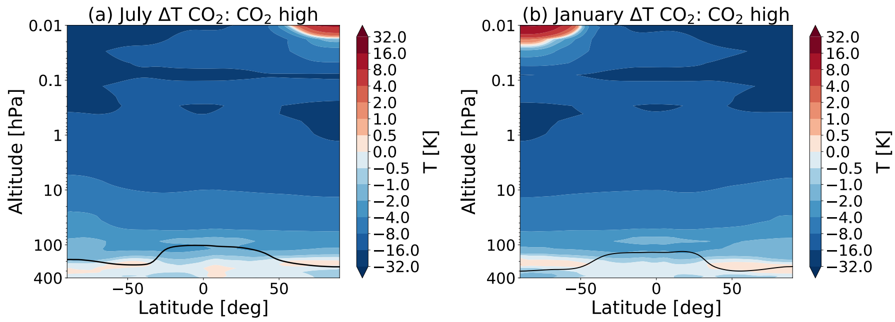

Figure 5: Temperature changes to direct CO_2 forcing in July (left) and January (right) for the high CO_2 simulation as compared to pre-industrial control simulation, as calculated by CFRAM. The black line indicates the tropopause height for the runs with changed CO_2-concentration.

Changing the SSTs does not lead to a change in CO_2-concentration, therefore the temperature response to changes in CO_2 is not present for the run with only changed SST (Figures not shown). We observed that the temperature responses to CO_2 and changed SST are approximately linearly additive, which means that the temperature response to CO_2 in the combined experiment is equal to the temperature response to CO_2 alone.

3.3 Dynamical feedback

Ozone is the largest feedback, which mitigates the cooling due to the direct CO_2 forcing in most parts of the middle atmosphere (Fig. 2). Climate change affects stratospheric ozone through changes in middle atmospheric chemistry, through changes in dynamics or a combination of these two factors. To understand how the ozone concentration is changing, it is necessary to understand how the dynamics in the middle atmosphere is altered in a double CO_2 climate. Therefore, we will first go into the dynamics of the middle atmosphere and the temperature responses to dynamical changes.

The transport of air in the stratosphere and mesosphere is controlled by the the Brewer-Dobson circulation (BDC). The Brewer-Dobson circulation consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. The BDC in the mesosphere spans from the summer to the winter pole (*Butchart et al*. 2010).

The zonal mean residual circulation forms an important component of the mass transport by the BDC. It consists of a merional ($\bar{v}^*$) and a vertical ($\bar{v}^*$) component as defined in the Transformed Eulerian Mean (TEM) framework. The residual circulation consists of a shallow branch which controls the transport of air in the tropical lower stratosphere, as well as a deep branch in the mid-latitude upper stratosphere and mesosphere.

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Both of these branches are driven by atmospheric waves. In the winter hemisphere, planetary Rossby waves propagate upwards into the stratosphere, where they break and deposit their momentum on the zonal mean flow, which in turns induces a meridional circulation. The two-cell structure in the lower stratosphere, which is present all-year round, is driven by synoptic waves. The circulation is also affected by orographic gravity wave drag in the stratosphere and by non-orographic gravity wave drag in the upper mesosphere (*Oberländer et al*., 2013).

Most climate models show that the BDC and the upwelling in the equatorial region will speed up due to an increase in CO_2-concentration (*Butchart el al.,* 2010). It has been shown that the strengthening of the Brewer-Dobson circulation in the lower stratosphere is caused by changes in transient planetary and synoptic waves, while the upper stratospheric changes are due to changes in the propagation properties for gravity waves (*Oberländer et al*., 2013).

It has been explained that the increased stratospheric resolved wave drag is caused by an increase of the meridional temperature gradient in the stratosphere, which leads to a strengtening of the upper flank of the subtropical jets. This in turn shifts the critical layers for Rossby wave breaking upward, which allows for more Rossby waves to reach the lower stratosphere, where they break and deposit their momentum, enhancing the BDC (*Shepherd and McLandress*, 2011)

The changes in the meridional component of the residual circulation ($\bar{v}^*$) for the different simulations are shown in Fig. 6. Figure 6b and 6e show that only doubling the CO_2 leads to a stronger pole-to-pole flow in the mesosphere. Changing the SSTs also leads to changes in the residual circulation as can be seen in Fig. 6c and 6f. Oberländer et al. (*2013*) have shown that the rising CO_2-concentration affects the upper stratospheric layers, while the signals in the lower stratosphere are almost completely due to changes in sea surface temperature. The warmer sea surface temperatures enhance the activity of transient planetary waves and orographic gravity waves in the lower and middle stratosphere. The changed SSTs also leads to enhanced dissipation of plantery waves, as well als orographic and non-orographic waves in the upper stratosphere.

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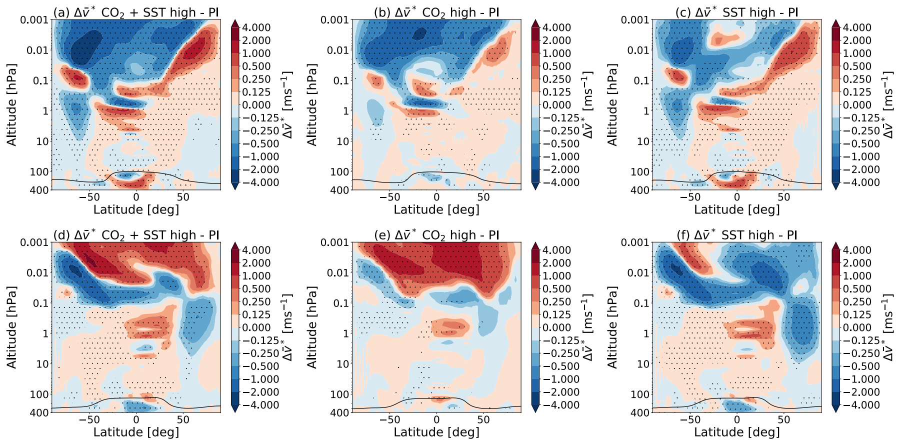

Figure 6: The changes in transformed Eulerian-mean residual circulation horizontal velocity $\bar{v}^*$ in July (top) and January (bottom) for (a,d) the simulation with high CO_2 and SSTs, (b,e) the simulation with high CO_2, (c,f) the simulation with high SSTs, all as compared to pre-industrial control simulation, as found by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

We are interested in the temperature responses due to the dynamical feedbacks in the different experiments. These temperature responses are shown in Figure 7. Figure 7b and 7e show that there is cooling in the summer mesosphere, while there is warming in the winter mesosphere, which is consistent with a stronger summer-to-winter pole flow.

Figure 7c and 7f show the temperature responses due to changes in the SSTs. It is seen that there is mostly a warming in the summer mesosphere and mostly a cooling in the winter hemisphere, which would weaken the effect of the changed CO_2-concentration. Most of the temperature responses in the lower stratosphere are caused by the changes in SSTs, as expected.

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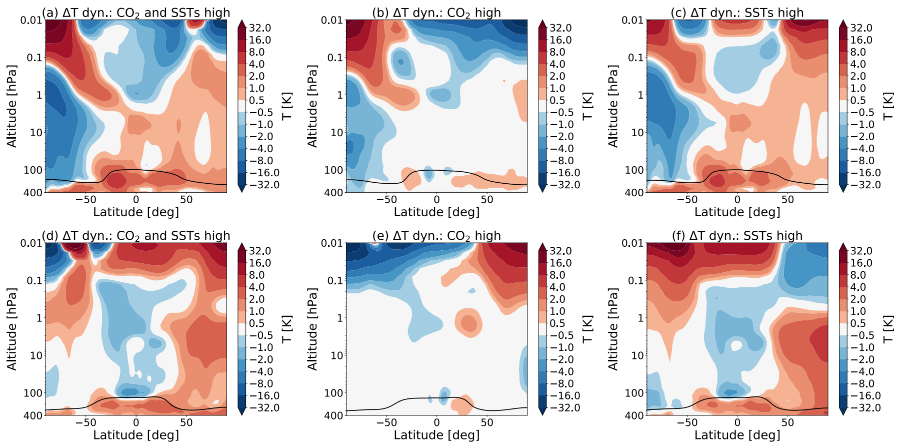

Figure 7: Temperature responses to changes in dynamics, as calculated by CFRAM, in July (top) and January (bottom) for (a,d) the simulation with high CO2 and SSTs, (b,e) the simulation with high CO2, (c,f) the simulation with high SSTs, all as compared to pre-industrial control simulation. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

In summary, doubling the CO_2 leads to a stronger pole-to-pole flow in the mesosphere, which leads to cooling of the summer mesosphere and a warming of the winter mesosphere. Changing the SSTs weakens this effect, but leads to temperature changes in the stratosphere and lower mesosphere.

3.4 Ozone feedback

The ozone concentration in the middle atmosphere is for a large part determined by so-called Chapman reactions, in which molecular oxygen (O_2) is photodissociated to two atomic oxygen atoms, which may recombine in a three-body process or react with molecular oxygen to produce ozone (*Brasseur and Solomon,* 2005).

Changes in the circulation and dynamical transport have a significant impact on the ozone distribution (*Salby et al.,* 2011, *Weber et al.,* 2011). The O_3-concentration is further determined by interactions between other chemical constituents, which are also impacted when the CO_2-concentration is increased (*Brasseur and Solomon,* 2005).

Fig. 8 shows the percentage changes in O_3-concentration when the CO_2-concentration and/or the SSTs change. An increase in CO_2, leads to an increase of ozone in most of the middle atmosphere. However, in the tropical lower stratosphere, the summer polar mesosphere, the winter and equatorial mesosphere, a decrease in ozone is seen. Fig. 8c and f show that changing the SSTs also has a significant impact on the ozone concentration. A

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complete account of the ozone changes is out of the scope of this paper, but the main processes responsible for ozone changes will be discussed.

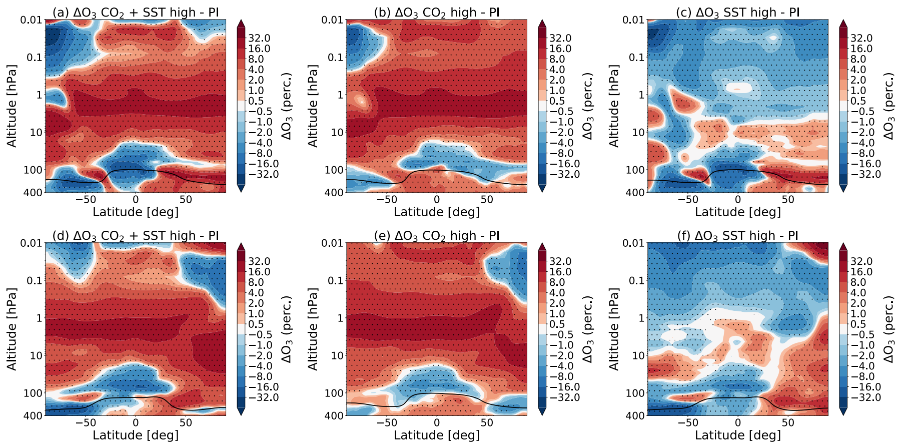

Figure 8: The percentage changes in ozone concentration in July (top) and January (bottom) for (a,d) the simulation with high CO2 and SSTs, (b,e) the simulation with high CO2, (c,f) the simulation with high SSTs, all as compared to the pre-industrial control simulation, as found by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

The increase in ozone due to an increase in CO_2, has been shown to be primarily a result of the temperature dependence of the reaction $O + O_2 + M \rightarrow O_3 + M$ (*Jonsson et al.,* 2004). In the upper stratosphere, ozone destruction is slowed down, and leads to an increase in O_3 (*Bekki et al.,* 2011). The decrease of ozone at the high latitudes in the summer mesosphere, is due to a decrease in atomic oxygen which results from increased upwelling. The decrease in O_3 concentration in the polar winter around 0.1 hPa is due to a stronger subsidence of NO and Cl, which are both ozone-destroying constituents (*Schmidt et al.,* 2006).

In addition to changes in the chemistry in the middle atmosphere, we have seen that the dynamics are changing, which leads to changes in the ozone concentration. The BDC affects the O_3-concentration directly, but also determines the amount of ozone depleting species. The decrease in the ozone concentration in the tropical lower stratosphere is due to an acceleration of the BDC. In a similar fashion, the ozone concentration in middle and high latitudes is increased (*Braesicke et al*., 2018). As can be seen in Fig. 8, the decrease of the O_3-concentration in the tropical upper troposphere region is mainly associated with changes in the SSTs.

The temperature responses due to the ozone feedback are shown in Figure 9. It can be seen that there is a warming in the regions, where there is an increase of the O_3-concentration, while there is a cooling for the regions with a

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decrease of the O_3-concentration. However, this is not the case for the winter polar region, where there is no sunlight. Note that the temperature responses to the changes in CO_2- and O_3- concentration behave differently in this respect: the temperature responses due to the direct forcing of CO_2 follow the temperature distribution quite closely, while the temperature responses due to O_3 follow the ozone concentration, as also seen by *Zhu et al.,* (2016).

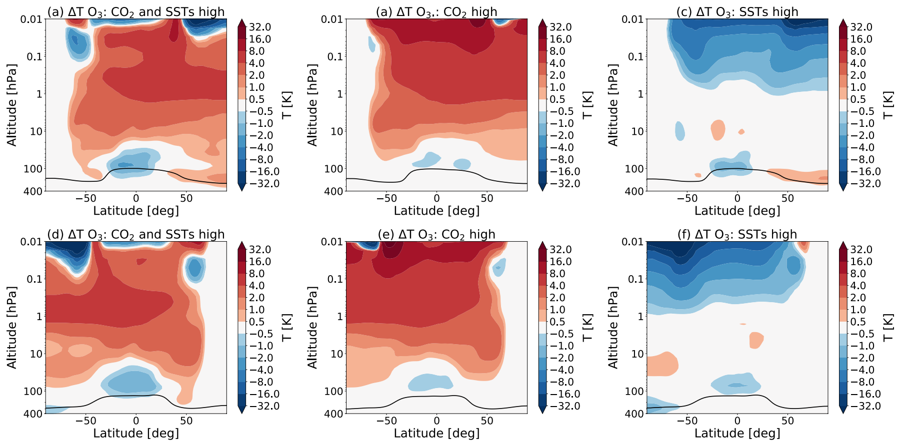

Figure 9: Temperature responses to changes in O_3-concentration, as calculated by CFRAM, in July (top) and January (bottom) for (a,d) the run with high CO2 and SSTs, (b,e) the run with high CO2, (c,f) the run with high SSTs, all as compared to pre-industrial conditions. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

3.5 Water vapour feedback

Water vapour plays a secondary but not negligible role in determining the middle atmosphere climate sensitivity. In Figure 3, we saw the temperature responses to the different feedbacks between 20°S and 20°N in July and January for changes in both the SSTs and the CO_2-concentration. It can be seen that in part of the lower stratosphere, H_2O is the main cooling agent instead of CO_2. Above 30 hPa**,** the water vapour contribution to the energy budget is negligible, as also seen by *Fomichev et al.* (2007).

Figure 10 shows how the water vapour is changing in the middle atmosphere if the CO_2-concentration and/or the SSTs are enhanced. Increasing the CO_2-concentration alone leads to a decrease of water vapour in most of the middle atmosphere. This is due to the decrease in temperature in the tropical tropopause for the double CO_2 experiment in WACCM of about -0.25 K. The cold temperatures in the tropical tropopause lead to a reduction of water vapour of between 2 and 8% due to freeze-drying in this region.

It can be seen that changing the SSTs leads to an increase in water vapour almost everywhere in the middle atmosphere (Fig. 10c and f). The increase in SSTs leads a higher and warmer tropopause, which can explain this increase

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of water vapour. There is a considerable decrease in water vapour in the winter polar stratosphere in the experiment with only doubling CO_2. The same phenomenon is seen for the experiment with double CO_2 and changed SSTs.

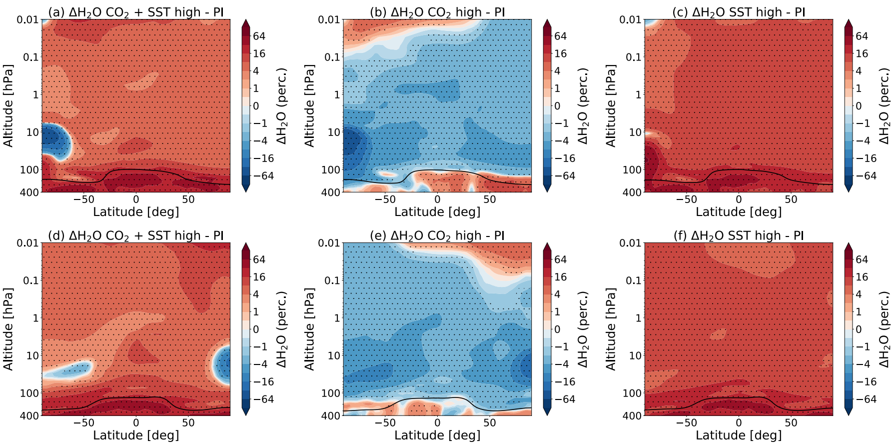

Figure 10: The percentage changes in water vapour mixing ratio in July (top) and January (bottom) for (a,d) the run with high CO2 and SSTs, (b,e) the run with high CO2, (c,f) the run with high SSTs, all as compared to pre-industrial conditions, as found by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

Figure 11 shows the temperature responses due to the changes in water vapour. It can be seen that the regions where there is an increase (decrease) in the water vapour, there is a cooling (warming). An increase (decrease) in water vapour in the middle atmosphere leads to an increase (decrease) in longwave emissions of water vapour in the mid and far-infrared, which leads to a cooling (warming) (*Brasseur and Solomon,* 2005). Higher up in the atmosphere, there are large percentage changes in water vapour, but the absolute concentration of water is small there, which explains why there is no temperature response to these changes.

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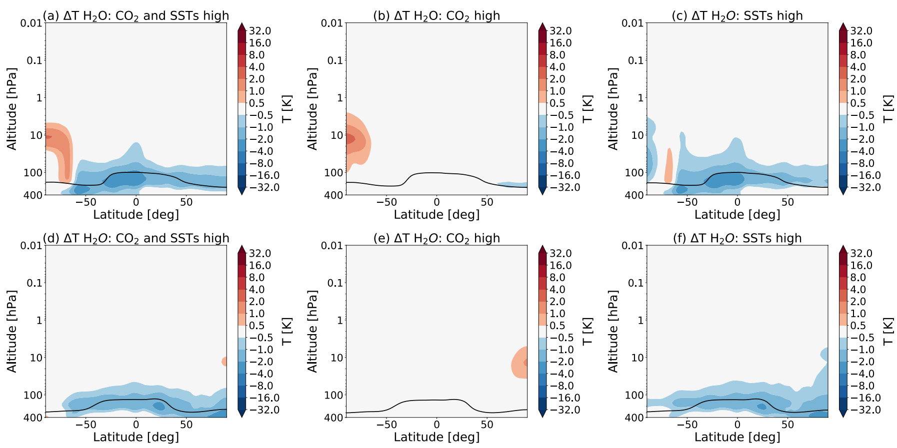

Figure 11: Temperature responses to changes in H_2O-concentration in July (top) and January (bottom) for (a,d) the run with high CO_2 and SSTs, (b,e) the run with high CO2, (c,f) the run with high SSTs, all as compared to pre-industrial conditions, as calculated by CFRAM. The black line indicates the tropopause height for the runs with changed CO_2-concentration and SSTs (a,d), with changed CO_2-concentration (b,e) and with changed SSTs (c,f).

3.6 Cloud and albedo feedback

It is known that feedbacks due to changes in clouds and surface albedo play a crucial role in determining the tropospheric and surface climate (*Boucher et al.,* 2013, *Royer et al.,* 1990). We have seen in Figure 2 that these feedbacks play only a very small role in the middle atmosphere temperature response to the doubling of CO_2 and SSTs. However, there are some small radiative effects from the cloud and albedo feedback, that are the result of the changes in SSTs, as shown in Figure 12.

Changes in SSTs yield an overall increase in the cloud cover in the upper troposphere, while this is not the case if one only increases the CO_2 concentration. We see that in the tropical region there is a warming due to changes in clouds, while there is a cooling at higher latitudes in July (see Figure 12a). In January, the pattern looks slightly different (see Figure 12c). These temperature changes are due to changes in the balance between the increased reflected shortwave radiation and the decrease of outgoing longwave radiation.

We also see an effect of the changes in surface albedo in the stratosphere (see Figure 12 b and d). The temperature responses shown in Figure 12 b and d are due to radiative changes. The decrease in surface albedo would cause less shortwave radiation being reflected. We suggest that this leads to the cooling seen in the summer polar stratosphere, but more research is needed.

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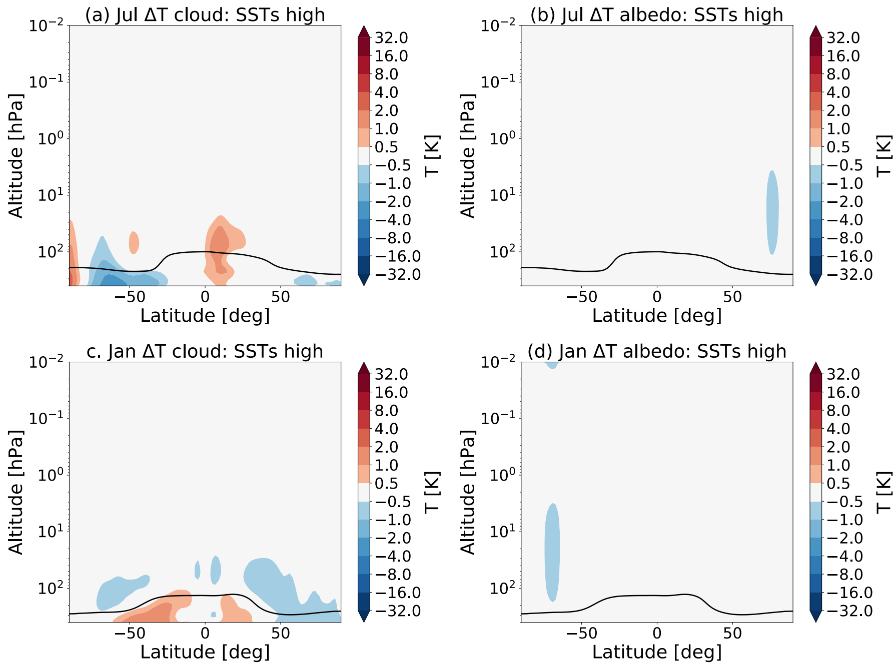

Figure 12: Temperature responses to changes in cloud (left) and albedo (right) in July (top) and January (bottom) for the run with high SSTs as compared to pre-industrial conditions, as calculated by CFRAM. The black line indicates the tropopause height for the runs with changed SSTs.

4. Discussion and conclusions

In this study, we have applied the climate feedback response analysis method to climate sensitivity experiments performed by WACCM. We have examined the middle atmosphere response to CO_2 doubling with respect to the pre-industrial state. We also investigated the combined effect of doubling CO_2 and subsequent warming SSTs, as well as the effects of separately changing the CO_2 and the SSTs.

It was seen before that the sum of the two separate temperature changes in the experiment with only changed CO_2 and only changed SSTs is, at first approximation, equal to the changes observed in the combined simulation (see e.g. *Fomichev et al. (2007) and Schmidt et al. (2006)*). This is also the case for WACCM.

We have found that, even though changing the SSTs yields significant temperature changes over a large part of the middle atmosphere, the effects of the changed SSTs on the middle atmosphere are relatively small as compared to the effects of changing the CO_2 without changes in the SSTs.

We also have seen that the cooling due to the direct forcing of CO_2 alone would lead to a cooling of about 9 K between 200 and 0.01 hPa (~10-80 km), for both January and July. However, the cooling in this region for the

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experiment with a high CO_2 and SSTs is only 5.2 K in January and 5.5 K in July. This means that the cooling due to the direct forcing of CO_2 is being mitigated by the combined effect of the different feedback processes.

Different chemical components, in particular ozone, respond to changes in CO_2 and/or SSTs due to changes in chemical reaction rate constants and atmospheric density (due to changes in temperature) and due to the strength of the up- and downwelling (*Schmidt et al.,* 2006). The ozone feedback counteracts the cooling effect, by mitigating the cooling due to changes in CO_2 by approximately 1.5 K in both July and January.

The ozone concentration increases in most parts of the middle atmosphere, with the biggest change seen around 1 hPa. The ozone decreases in the tropical lower stratosphere and certain parts of the mesosphere. The temperature response of the ozone feedback follows the concentration of O_3. The ozone feedback leads to a stronger warming for the case where only the CO_2-concentration is changed, while the changes in SSTs weaken this effect in the mesosphere.

We also have seen that the global mean temperature response due to dynamical feedbacks is small, while the local responses to the changes in dynamics are large. Doubling the CO_2 leads to a stronger summer-to-winter-pole flow, which leads to cooling of the summer mesosphere and a warming of the winter mesosphere. Changing the SSTs weakens this effect in the mesosphere, but leads to temperature changes in the stratosphere and lower mesosphere.

The role of the water vapour feedback in the middle atmosphere is generally quite small. However, the water vapour feedback forms the largest contribution to the temperature change in some regions of the lower stratosphere, mainly as a result of changing SSTs. Changing the SSTs leads to an increase of water vapour, which leads to a cooling in the lower stratosphere. Changing the CO_2-concentration leads to a decrease in water vapour, due to enhanced freeze-drying.

While the cloud and surface albedo feedbacks are of crucial importance for the tropospheric and surface climate (*Boucher et al.,* 2013, *Royer et al.,* 1990), in the middle atmosphere their role is very small. There are some temperature responses to these feedbacks, that are the result of the changed SSTs.

It would also be interesting to investigate the exact mechanisms behind the feedback processes in more detail. Some processes can influence the different feedback processes, such as ozone depleting chemicals influencing the ozone concentration and thereby the temperature response of this feedback. Other studies have shown that a surface albedo change, which is associated with sea ice loss, can influence the middle atmosphere dynamics, which in turn influences the temperature response (*Jaiser et al.,* 2013). The CFRAM cannot unravel the effects of these different processes.

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There is also a need for a better understanding of how different feedbacks in the middle atmosphere affect the surface climate. As discussed in the introduction, the exact importance of ozone feedback is currently not clear While this paper focused on the temperature changes in the middle atmosphere, similar analysis can be done to quantify the effects of feedbacks on the surface climate.

In conclusion, we have seen that CFRAM is an efficient method to quantify climate feedbacks in the middle atmosphere, although some refinement of the method is necessary to reduce the error in this region. The CFRAM allows for separating and estimating the temperature responses due an external forcing and various climate feedbacks, such as ozone, water vapour, cloud, albedo and dynamical feedbacks. More research into the exact mechanisms of these feedbacks could help us to understand the temperature response of the middle atmosphere and their effects on the surface and tropospheric climate better.

Acknowledgement

The computations and simulations were performed on resources provided by the Swedish National Infrastructure for Computing (SNIC) at National Supercomputer Center (NSC) in Linköping.

Hamish Struthers NSC is acknowledged for assistance concerning technical aspects in making the WACCM code run on NSC supercomputer Tetralith. We thank Qiang Zhang for helping to make the radiation model code applicable to WACCM model data.

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