Responses to the Editor Comments on the revised manuscript by Kuilman et al.

We thank the editor for her extensive comments.

Specific comments:

Title: Since CFRAM play a major role in your study it should also appear in the title.

The title has now been changed to *"*Using the climate feedback response method to quantify climate feedbacks in the middle atmosphere in WACCM*"*

Key points: Here you clearly summarize your study, however, as you write it here it does not appear in the abstracts. Since ACP does not use key points these will be simply lost after publication. Therefore, I would strongly suggest that you include these in your abstract.

These points are included in the abstract.

Abstract: An abstract should be clearly written and summarize the idea and results of a study. Here are too many weird or complicated sentences that distract from the content of the study. Further, not all what you have done is summarized here. Therefore, the abstract should be revised. The abbreviation "CO2" is used throughout the paper but has never been introduced.

The abbreviation CO_2 has now been introduced in the abstract. The abstract has been significantly revised (see below).

P2, L51-52: Already the first sentence is rather weird formulated. I would never "increasingly realized". Please rephrase.

This sentence has been rephrased.

P2, L58-59: "We find the……" This sentence is also not clear. I would suggest to split it into two sentences.

As expected, our results show that the direct forcing of CO_2 cools the middle atmosphere. This cooling becomes stronger with increasing height: the cooling in the upper stratosphere is about three times as strong as the cooling in the lower stratosphere.

P2, L72-73: Same here. It would be better to split this sentence.

However, the changes in SSTs are responsible for dynamical feedbacks that cause large temperature changes. Moreover, the temperature response to the water vapour feedback in the lower stratosphere is almost solely due to changes in the SSTs.

General comment: Why do we consider doubled CO2 atmospheres? One sentence should be included to motivate why such scenarios are of interest.

This has now been added*: "A better understanding of the middle atmosphere and how it reacts to the current increase of the concentration of carbon dioxide (CO_2) is therefore necessary."*

Abstract

Over recent decades it has become clear that the middle atmosphere has a significant impact on surface and tropospheric climate. A better understanding of the middle atmosphere and how it reacts to the current increase of the concentration of carbon dioxide is therefore necessary. In this study, we investigate climate feedbacks in response to a doubling CO_2 in the middle atmosphere using the climate feedback response analysis method (CFRAM). With this method, one can calculate the partial temperature changes due to an external forcing and climate feedbacks in the atmosphere. As this method has the unique feature of additivity, these partial temperature changes are linearly addable. In this study, we discuss the direct forcing of CO_2 and the effects of the ozone, water vapour, cloud, albedo and dynamical feedbacks.
As expected, our results show that the direct forcing of CO_2 cools the middle atmosphere. This cooling becomes stronger with increasing height: the cooling in the upper stratosphere is about three times as strong as the cooling in the lower stratosphere. The ozone feedback yields a radiative feedback that mitigates this cooling in most regions of the middle atmosphere. However, in the tropical lower stratosphere and in some regions of the mesosphere, the ozone feedback has a cooling effect. The increase in CO_2-concentration causes the dynamics to change. The temperature response due to this dynamical feedback is small in the global average, although there are large temperature changes due to this feedback locally. The temperature change in the lower stratosphere is influenced by the water vapour feedback and to a lesser degree by the cloud and albedo feedback. These feedbacks play no role in the upper stratosphere and the mesosphere. We find that the effects of the changed SSTs on the middle atmosphere are relatively small as compared to the effects of changing the CO_2. However, the changes in SSTs are responsible for dynamical feedbacks that cause large temperature changes. Moreover, the temperature response to the water vapour feedback in the lower stratosphere is almost solely due to changes in the SSTs. As CFRAM has not been applied to the middle atmosphere in this way before, this study also serves to investigate the applicability as well as the limitations of this method. This work shows that CFRAM is a very powerful tool to study climate feedbacks in the middle atmosphere. However, it should be noted that there is a relatively large error term associated with the current method in the middle atmosphere, which can be for a large part be explained by the linearization in the method.

Introduction: P2, L89: "ozone is responsible for the existence of the stratosphere…….". This may be indirectly correct, but it sounds really weird and thus should be rephrased.

The introduction has been thoroughly rewritten and this part is no longer there (see below).

P2, L93-95: Also this is a really weird paragraph. This paragraph should be completely rewritten.

The introduction has been thoroughly rewritten and this part is no longer there (see below).

P3, L102: I am not aware of that. These are generally roughly parameterized. Which processes are you referring to? Please give some examples.

The introduction has been thoroughly rewritten and this part is no longer there (see below).

P3, L109ff: How is ozone represented in the climate models? This should be added.

The introduction has been rewritten and the study of Nowack has been described more clearly:
Nowack et al. (2015) has found that there is an increase in global mean surface warming of about 1 °C when the ozone is prescribed at pre-industrial levels, as compared with when it is evolving in response to an abrupt 4xCO_2 forcing.

P3, L116ff: How does CO2 affect the middle atmosphere? What role does ozone play in this context?

The introduction has been rewritten and the importance of the changes in CO_2 and O_3 has now been clearer.

General comment on the introduction: The first part of the introduction (L84-L131) is really not well written. I do not see here the relation to your study. There is unfortunately no clear line. Therefore, I would suggest to completely revise this part.

The introduction has been thoroughly rewritten.

P4, L176: Nothing mentioned here that the feedbacks are discussed separately.

This has been added.

Please find below the rewritten introduction:

1. Introduction

The increase of concentration of carbon dioxide in the atmosphere forms a major perturbation to the climate system. It is commonly associated with lower-atmospheric warming. However, in the middle atmosphere, the increase of CO_2 leads to a cooling of the region instead. This cooling has been well documented and is found by both model studies and observations (e.g. Manabe and Wetherald, 1975; Ramaswamy et al., 2001; Beig et al., 2003).

The middle atmosphere is not only affected by the increase in CO_2-concentration, but also by the decrease in ozone-concentration. The depletion of ozone (O_3) also effects the temperature in the stratosphere and leads to a cooling (Shine et al, 2003). A better understanding of the effect of the increased CO_2-concentration on the *middle atmosphere, will help to distinguish the effects of the changes CO_2- and O_3-concentration.*

Another major motivation for this study is the emerging evidence that the middle atmosphere has an important influence on surface and tropospheric climate (Shaw and Shepherd, 2008). It has, for example, been shown that cold winters in Siberia are linked to changes in the stratospheric circulation (Zhang et al., 2018).

Nowack et al. (2015) has found that there is an increase in global mean surface warming of about 1°C when the ozone is prescribed at pre-industrial levels, as compared with when it is evolving in response to an abrupt $4xCO_2$ forcing. It should be noted that 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).

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 climate 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).

The climate feedback-response analysis method (CFRAM) is an alternative method which takes into account that the climate change is not only determined by the energy balance at the top of the atmosphere, but is also influenced by the energy flow within the Earth's system itself (Cai and Lu, 2009, Lu and Cai, 2009). 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. This makes it possible to calculate the partial temperature changes due to an external forcing and these internal feedbacks in the atmosphere. It has the unique feature of additivity, such that these partial temperature changes are linearly *addable.*

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). CFRAM has been applied to study the middle atmosphere climate sensitivity as well (Zhu et al., 2016). In their study, Zhu et al. (2016) have adapted CFRAM and applied it to both model output, as well as observations. The atmospheric responses during solar maximum and minimum were studied and it was found that the variation in solar flux forms the largest radiative component of the middle atmosphere temperature response.

In the present work, we apply CFRAM to climate sensitivity experiments performed with 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 middle atmosphere response to CO_2-doubling. We acknowledge that such idealized equilibrium simulation cannot reproduce the complexity of the atmosphere, in which the CO_2-concentration is changing gradually. However, simulating a double CO_2-scenario still allows us to identify robust feedback processes in the middle atmosphere.

There are two aspects of the middle atmosphere response to CO_2-doubling: there is the effect of the changes in CO_2-concentration directly, as well as the changes in sea surface temperature (SST) which are in itself caused by the changes in CO_2-concentration. It is useful to investigate these aspects separately, as former should be robust, while the effect of the changed SST depends on the changes in tropospheric climate, which can be expected to depend more on the model.

In this study, we investigate the effects of doubling the CO_2-concentration and the accompanying sea surface temperature change on the temperature in the middle atmosphere as compared to the pre-industrial state. We use CFRAM to calculate the radiative contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, and clouds. We refer to the changes in ozone, water vapour, albedo and clouds in response to changes in the CO_2-concentration as the ozone, water vapour, albedo and cloud feedbacks.

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (Brewer, 1949; Dobson, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the *middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (Zhu et al., 2016).*

The main goal of this paper is to calculate the contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, clouds and dynamics in the middle atmosphere under a double CO_2-scenario using CFRAM. Our intention is not to give a complete account of the exact mechanisms behind the changes in ozone, water vapour, albedo, clouds and dynamics.

P4, L184: What is the resolution in the boundary layer? As you write it sounds as they use a different resolution in the boundary layer then in the remainder of the atmosphere.

The lowermost levels in WACCM are 1010, 992.556, 970, 929, 867, 788 hPa. The height of this levels of course dependent on the temperature and therefore it wasn't written in the paper. As the boundary layer is not of major importance for this paper, we have now rephrased this as follows.

"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 vertical resolution, which ranges from ~1.1 km in the troposphere, 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.

P5, L219: "……from a double CO2 equilibrium simulation by CESM." Why has this been done? From the description here I count three simulations, but in the table and later there are four simulations.

The use of prescribed SSTs is common for middle atmosphere CCMs (e.g. *Fomichev et al*., 2007).

Four experiments have been performed. Section 2.2 has now been rewritten to make clear what has been done (see below).

In this study, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions. This compset simulates an equilibrium state, which means that it runs a perpetual year 1850. Four experiments have been performed (see Table 1) for this study.

P5, L223: Add the years for which the simulation has been performed.

In this study, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions.

This means that we run the model for a perpetual year 1850, but then with the CO_2 concentration and/or the SSTs changed.

Table 1: What is “PI”? What SSTs have been used? Please add numbers in the table.

PI stands for pre-industrial, this has now been made clear in the text. The experiments are now referred to with a letter number combination.

2.2 Experimental set-up

In this study, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions. This compset simulates an equilibrium state, which means that it runs a perpetual year 1850. Four experiments have been performed for this study (see Table 1).

Experiment C1 is the control run, with the pre-industrial CO_2 concentration (280 ppm) and forced with pre-industrial ocean surface conditions such as sea surface temperature and sea ice (referred to SSTs from now on). These SSTs are generated from the CMIP5 pre-industrial control simulation by the fully coupled Earth system model CESM. The atmospheric component of CESM is the same as WACCM, but does not include stratospheric chemistry (Hurrel et al., 2013). This latter aspect is not considered in this study.

Experiment C2 represents the experiment with the CO_2 concentration doubled as compared to the pre-industrial state (560 ppm) and forced with the same pre-industrial SSTs as in experiment C1. 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.

The compset used in this experiment and all the following ones is still F_1850, which means that other radiatively and chemically active gases, such as ozone, will change only because of the changes in the CO_2-concentration, due to WACCM’s interactive chemistry. Chlorofluorocarbons (CFCs), which have a major impact on the ozone concentration in the real atmosphere, don’t play a role in our experiments.

In experiment S1, we simulate the scenario, in which there is the SSTs forcing from the coupled CESM for double CO_2 condition, but the pre-industrial CO_2-concentration of 280 ppm. S2 represents the experiment with the CO_2-concentration in the atmosphere doubled to 560 ppm and the SSTs prescribed for the double CO_2-climate. Experiment C1, C2, S1 and S2 will be also referred to hereafter by PI, the simulation with high CO2, the simulation with high SSTs and the simulation with high CO_2 and SSTs, respectively.

The experimental setup of this study is similar to the setup performed with the Canadian Middle Atmosphere Model (CMAM) by Fomichev et al. (2007) and with the Hamburg Model of the Neutral and Ionized Atmosphere (HAMMONIA) by Schmidt et al. (2006). The HAMMONIA model is coupled to the same chemical model as WACCM: MOZART3. The setup in their study is similar, however, in their study, they double the CO_2-concentration from 360 ppm to 720 ppm, while in our study, we double from the pre-industrial level of CO_2 (280 ppm).

Note that experiment S1 and C2 are not representing scenarios that could happen in the real atmosphere. These experiments have been used to study the effect of the SSTs separately. Experiment S2 doesn't take into account other (anthropogenic) changes in the atmosphere not caused by changes in the CO_2-concentration and the SSTs.

All the simulations are run for 50 years, of which the last 40 years are used for analysis. In the all results shown, we have used the 40 year mean of our model data.

Table 1. Set-up of the model experiments.

Experiment	***CO_2***	***SSTs from CESM equilibrium run***
C1	*280 ppm*	*PI control*
C2	*560 ppm*	*PI control*
S1	*280 ppm*	*Double CO_2 run*
S2	*560 ppm*	*Double CO_2 run*

P7, L287: we can calculate as -> can be calculated since ……

This has been rephrased.

General on this section: The description is much too long for the main part of the manuscript. You should shorten this Section and put the other parts to an Appendix.

We have shortened the method section substantially by putting a large part of the method in the appendix section, where we give the complete formulation of CFRAM diagnostics using outputs of WACCM.

General on Section 3: Also here I would suggest some restructuring. I would add subsections, on for the temperature responses and one for the feedbacks or directly use different Sections.

Section 3 has now been restructured. Section 3.1 has become section 3. Figs. 5-12 have now moved to section 4 (meridional-vertical profiles of partial temperature changes), which has 5 sub-sections, which discuss the temperature response to the different feedbacks. Fig. 2-4 are moved to section 5 (regional and global means of partial temperature changes due to feedbacks).

P10, L448: In section 2.2, it was….. -> As described in Section 2.2 four experiments…..

This has been changed as suggested.

P10, L453: Are you using a 40 year mean? This should be clearly stated in the manuscript.

We have added this now both the method section as well as in section 3.

For this figure, as well as for all the results shown in this paper, we have used the 40 year mean of our data.

P10, L455: Add references to the earlier studies.
References have now been added.

In line with what was shown in earlier studies (e.g. Akmaev, 2006; Fomichev et al., 2007), 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.

P11, L478-479: "all as compared the pre-industrial control simulation". That means the difference between these? Clearly state this. As it is written now it is rather confusing.

Yes, that is right. This has been rewritten for this figure as well as for all the other ones which had a similar formulation.

Changes in the zonal and monthly mean temperature (K) for July (top) and January (bottom) due to (a,d) combined effect of the CO_2 increase and SSTs changes (experiments C2 -B1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments B2 - B1) and the (c,f) SSTs (experiments C1 - B1).

P11, L488: Why can only the radiative feedback been calculated?

This has now been explained in the text.

This can be understood as we use the Fu-Liou radiative transfer model (Fu and Liou, 1992, 1993) to do offline calculations of the total local thermal equilibrium (LTE) radiative heating rate perturbation fields between the control experiment C1 and one of the other three experiments (i.e, C2, or S1, or S2). We use the standard outputs of atmospheric compositions (e.g., CO_2 and O_3) and thermodynamic fields (e.g., pressure, temperature, water vapour, clouds, surface albedo) as well as partial LTE radiative heating rate perturbation fields due to perturbations in individual atmospheric composition or thermodynamic fields (e.g., the terms on the right hand side of (A.9) except the first term).

We use the difference between the offline calculation of the total LTE radiative heating rate perturbations and the original total LTE radiative heating rate perturbations derived directly from the standard WACCM outputs as the error term of our offline LTE radiative heating perturbations. We note that the standard WACCM output fields also include non-LTE radiative heating fields, but do not include non-radiative heating rates. Therefore, we use the sum of the total LTE radiative heating rate perturbations and non-LTE radiative heating fields derived from the standard WACCM output fields to infer non-radiative heating rate perturbations under the equilibrium condition, namely Eq. (A.8).

P12, L498: It would be quite helpful if you would give your experiments names as it usually done in the modelling community and then use these names throughout the manuscript. It would make it much easier to follow which experiment you are actually discussing.

Experiment numbers (C1, C2, S1 and S2) have been added throughout the paper.

P12, L500-501: “the pressure level of which is an output of WACCM”. What do you mean with that?The tropopause is an output level? But it is always at a different height. P12, L505: Why do you not use temperature to derive the location of the tropopause?

WACCM has an output field which is the tropopause height in hPa (in the same way it outputs the temperature, water vapour etc.). This is a field that indeed varies with latitude, longitude and time. This has been taken into account, as can be seen in Fig. 2: the tropopause height varies with latitude (we have averaged over the time and longitude). WACCM also has the tropopause temperature and the temperature as separate output fields. It is indeed also possible to use the temperature profile to find the position of the tropopause, which would give you the same pressure level. We have taken the output of WACCM directly, as it is more convenient.

This has now been written in the text more clearly:

The pressure level of the tropopause is simulated in WACCM for each latitude and longitude. We use this pressure level to demarcate between the troposphere and stratosphere.

P12, L501: Why 24 hPa?

We consider 24 hPa as a crude estimate for the boundary between the lower and upper stratosphere. This has now also been added to text.

P12, L506: Why is a mass weighting important? What is the error/uncertainty of not doing this?

I added this sentence because a reviewer asked for this. But there is no error or uncertainty involved in such average calculations regardless of whether we consider mass weighting or not. The mass weighting would emphasize the values in the lower levels much more heavily as the mass decreases with height exponentially. By not considering mass weight, the vertical average is just simple arithmetical mean and we can directly compare the vertical average values with their counterparts showing in the vertical-latitude cross-section diagrams.

General comment: How do your results agree with previous studies? Are your feedbacks higher or lower?

The CFRAM method allows to calculate for each location in the middle atmosphere, how much of the temperature change is due to which process. No other method before could do that. We have made this more explicit in the text.

The ozone feedback generally yields a radiative feedback that mitigates the cooling, which is due to the direct forcing of CO_2. This has been suggested in earlier studies, such as Jonsson et al., 2004, Dietmüller et al., 2014. With CFRAM, it is possible to quantify this effect and to compare it with the effects of other feedbacks in the middle atmosphere. Note that no other method before has been able to quantify how much *of the temperature change in the middle atmosphere is due to the different feedback processes before.*

P13, L559: Something missing here after "radiative"?

That is correct. What we meant to write was as follows:

Another part of the error is due to the fact that the radiative transfer model used in the offline CFRAM calculations is different than the radiative transfer model used in the climate simulations with WACCM.

General comment: Discuss a bit more in which altitude/latitude region the impact is highest/lowest and give numbers.

The current Section 5 has been rewritten and now includes more quantification, as shown below (the same is done for the Conclusion and Discussion section).

3. Regional and global means of partial temperature changes due to feedbacks

To study the relative importance of the different feedback processes globally we show the average change in global mean temperature for the lower stratosphere, the upper stratosphere and the mesosphere for the S2 experiment with the changed CO_2-concentration and changed SSTs in Figure 11. We also show the average change in temperature in the polar regions (90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) for the lower and upper stratosphere and the mesosphere.

In order to calculate the lower stratosphere temperature changes, we take the average value of the temperature change from the tropopause until 24 hPa. The pressure level of the tropopause is simulated in WACCM for each latitude and longitude, we use this pressure level to demarcate between the troposphere and stratosphere. We consider 24 hPa as a crude estimate for the boundary between the lower and upper stratosphere.

The tropopause is not exactly at the same pressure level in the perturbation experiments as compared to the pre-industrial control run (C1). We always take the tropopause of the perturbation experiment which is a bit higher at some latitudes, to make sure that we do not use values from the troposphere. We add the values for each latitude up and take the average. This average is not mass weighted. By taking the in this way, we can directly compare the vertical values in different regions of the atmosphere. The temperature changes in the upper stratosphere and in the mesosphere are calculated in the same way, but then for the altitudes 24 hPa-1 hPa and 1 hPa-0.01 hPa respectively.

Figure 11 shows the radiative feedbacks due to ozone, water vapour, clouds, albedo and the dynamical feedback, as well as the small contribution due to the Non-LTE processes, as calculated by CFRAM. The 'total'-column shows the temperature changes in WACCM, while the column 'error' shows the difference between *temperature change in WACCM and the sum of the calculated temperature responses in CFRAM. Note that the range of values on the y-axis is not the same for the different subplots.*

Figure 11 shows that the temperature change in the lower stratosphere due to the direct forcing of CO_2 is around 3 K in the global mean. There is a stronger cooling in the tropical region of about 4 K in July and 3.5 K in January. We also observe that there is a cooling of about 1 K due to ozone feedback in the tropical region while there is a slight warming taking place in the summer hemispheres in both January and July. We also see that the temperature change in the lower stratosphere is influenced by the water vapour feedback. There is a cooling of about 0.5 K in the lower stratosphere, apart from in the southern polar area. There is some small influence from the cloud and albedo feedback, which can be negative or positive (see also Fig. 9).

In the upper stratosphere, the cooling due to the direct forcing of CO_2 is with about 9 K in the global mean considerably stronger than in the lower stratosphere. The cooling is stronger in the summer polar regions, where the cooling due to the direct forcing of CO_2 reaches 11K. In the winter polar region, this cooling is only about 8K.

The water vapour, cloud and albedo feedback play no role in the upper stratosphere nor in the mesosphere. The ozone feedback results in the positive partial temperature changes in the upper stratosphere, of about 2 K in the global mean. The changes in ozone don't result in temperature changes in the winter hemisphere, as discussed in section 4.2.

The picture in the mesosphere is similar as in the upper stratosphere. The main difference is that the temperature changes are larger. The global temperature change due to direct forcing of CO_2 is about 15 K. The O_3-feedback results in a partial temperature changes of about 3 K in the mesosphere in the global mean. The temperature change due to ozone in the equatorial mesosphere is about 4 K, while the warming due to ozone in the summer polar region a bit smaller: around 3K. Just like in the upper stratosphere water vapour, cloud and albedo feedback play no role.

We see, that the ozone feedback generally yields a radiative feedback that mitigates the cooling, which is due to the direct forcing of CO_2. This has been suggested in earlier studies, such as Jonsson et al., 2004, Dietmüller et al., 2014. With CFRAM, it is possible to quantify this effect and to compare it with the effects of other feedbacks in the middle atmosphere. Note that no other method before has been able to quantify how much of the temperature change in the middle atmosphere is due to the different feedback processes before.

The temperature response due to dynamical feedbacks is small in the global average: less than 1 K. This can be understood as waves generally do not generate momentum and heat, but redistribute these instead (Zhu et al., 2016). However, the local responses to dynamical changes in the high latitudes are large, as we have seen in section 4.2. There are some very small temperature responses due to non-LTE effects as well, which contribute to the temperature change in the mesosphere mostly.

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 transfer model used in the offline CFRAM calculations is different than the radiative transfer model used in the climate simulations with WACCM.

Figure 2: I see here the highest feedbacks for the mesosphere. This is not discussed like that in the main text.

This was stated in the text as follows: *The picture in the mesosphere is similar. The main difference is that the temperature changes are larger, note the difference of the range for the temperature change between Fig. 10c, d and Fig. 10e,f.*

As this was not clear enough, it has been rewritten as follows:

The picture in the mesosphere is similar as in the upper stratosphere. The main difference is that the temperature changes are larger. The global temperature change due to direct forcing of CO_2 is about 15 K, which is stronger than in the upper stratosphere. The O_3-feedback results in a partial temperature changes of about 3 K in the mesosphere. Just like in the upper stratosphere water vapour, cloud and albedo feedback play no role.

Further, you use different y-axis scale which masks a bit the differences between the atmospheric regions. At least you should mention the different y-scales in the figure caption.

This was done to be able to make out the relative importance of the feedbacks in the different regions. We have now made note of this point in both the text and the caption.

Note that the range of values on the y-axis is not the same for the different subplots.

Figure 3: Here, a strong cooling due to CO2 is visible while all others rather show a warming. This is not discussed like this in the text. At least it is not clearly stated.

This has been added to the text and is also mentioned in section 4.1 *We see that increasing CO_2 leads to a cooling almost everywhere in the middle atmosphere, except at the high latitudes in the cold summer upper mesosphere, where we see a warming instead.*

P17, L606, Figure 4: Why has the upper stratosphere picked? Changes seem to be highest in the mesosphere. Why has the separation not been done there?

The only reason that the separation has not been done here is that one would end up with a figure consisting of 3 pages. Or three separate big figures. This can be added, if needed, but we think it will not add much information.

The aim of this figure is to show the temperature response of the CO_2 and the SST separately. The upper stratosphere is a bit more interesting then showing the mesosphere as there is still some albeit small effect from the albedo and water vapour, while in the mesosphere this is not the case. The text about this has been edited:

Figure 13 shows temperature responses in the upper stratosphere for the experiment with double CO_2 (a,b) and changed SSTs (c,d) separately. This has been done to give insight in the temperature response of the CO_2 and the SST separately. These temperature changes were calculated in the same way as for Fig. 11. Again also, the 'total'-column shows the temperature changes as simulated 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.

We learn from this figure that the effects of the changed SSTs on the upper stratosphere are relatively small as compared to the effects of changing the CO_2. We show the temperature changes for the upper stratosphere as an example. In the lower stratosphere and the mesosphere, we see the same pattern: the effect of the CO_2 on the temperature is generally much larger than the effect of the SSTs on the temperature. This finding is consistent with the study of Fomichev et al. (2007), where it is concluded that the impact of changes in SSTs on the middle atmosphere is relatively small and localized as compared to the combined response of changing the CO_2-concentration and the SSTs.

P17, L621: Thus, SSTs are important for the water vapour feedback on temperature, but lower for the CO2 feedback on temperature. This could also be more clearly stated.

Here, we discuss the calculated temperature responses for the experiment with double CO_2 (C2) and changed SSTs separately (S1). We make this now clearer in the text.

We learn from this figure that the effects of the changed SSTs on the upper stratosphere are relatively small as compared to the effects of changing the CO_2. In the lower stratosphere and the mesosphere, we see the same pattern: the effect of the CO_2 on the temperature is generally much larger than the effect of the SSTs on the temperature.

The changes in SSTs are, however, responsible for large temperature changes as a result of the dynamical feedbacks, especially in the winter hemispheres. A similar figure for the lower stratosphere (not shown) shows that the temperature response to the water vapour feedback is almost solely due to changes in the SSTs.

P17, L631: Temperature direct response -> Direct temperature response

This has been changed.

P18, L669: Add also the paper by Brewer, 1949, QJRMS
This reference has been added.

P19, L681ff: Most of this paragraph rather belongs to the introduction.

Part of what was written here has now been moved to the introduction.

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (Brewer, 1949; Dobson, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (Zhu et al., 2016).

This is to explain why we consider the dynamical feedback and what we mean by that in this study. As for the other feedbacks, I go into the background of this feedback in each of the different sections of section 4.

P21, L757ff: Please quantify your results and give the percentages.

The percentages are given now:

We find, as expected, that an increase in CO_2, leads to an increase of ozone in most of the middle atmosphere. The increase of O_3 is about 20% around 2 hPa in the tropical region for experiment S2 with respect to C1. This corresponds with what is seen by Fomichev et al., (2007), however they find that the increase in ozone in January is a bit lower in this region (around 15%, see their Figure 7).

There are some regions where the O_3-concentration is decreasing. In the tropical lower stratosphere, a decrease of about 20% is seen, in the summer polar mesosphere (around 0.01 hPa) ozone decreases by 3%, while in the mesosphere (around 0.02 hPa), ozone decreases by over 30%. Fig. 3c and f show that changing the SSTs also has a significant impact on the ozone concentration. A complete account of the ozone changes is out of the scope of this paper.

P21, L771: What symbol has been used for the statistical significance? Please add that to the figure caption.

This has been added: *The dotted regions indicate the regions where the data reaches a confidence level of 95%.*

P21, L780: Before you always wrote “ozone” but now you write “O_3” without introducing the abbreviation.

The chemical notation O_3 has now been introduced in the introduction of the paper, so that is clear that it signifies ozone.

P21, L787-789: Ozone is destroyed in polar winter in the lower stratosphere, not at 0.1 hPa.

We would like to refer to Schmidt et al., 2006. They don’t use the WACCM model, but HAMMONIA, however this is coupled to the same chemical model MOZART3. In their study, they perform a similar experiment as we do in our study, however they start from 360 ppm of CO2 and double to 720, while we start from the pre-industrial level of CO2 (280 ppm). Our results are very similar.

They write *“The large decrease in the atomic oxygen mixing ratio at high summer latitudes above 0.01 hPa results from increased upwelling and leads also to an ozone decrease at this level. The ozone decrease in the polar winter around 0.1 hPa (approx. 65 km) is mainly caused by the increase of NO and (to a lesser extent) Cl mixing ratios due to stronger subsidence of NO and Cl-rich air”*

P22, L804 and several other occasions: the 2 in CO2 should be written as subscript.
All instances of CO_2 are now written with a subscript in the paper.

General comment: Discuss also the statistical significance. Which results do you derive with which significance?

Comments about the statistical significance have been added in the figures that show the difference in a field between a perturbation run and the control run (such as Fig. 1). For the temperature responses we cannot calculate the statistical significance, but we can calculate the error, which we show in Fig. 10 for example.

Figure 10 caption: Add information what symbol has been used for the statistical significance.

This has been added: *The dotted regions indicate the regions where the data reaches a confidence level of 95%.*

P23, L853: Sentence with “This has been explained……..” is too complicated and should be rephrased,e.g. This can be explained by a ……leading to an …...

This sentence has been rewritten and split into several sentences:

This can be understood as increasing the water vapour in the middle atmosphere leads to an increase in longwave emissions in the mid and far-infrared by water vapour. This in turns leads ot a cooling of the region. Similarly, a decrease in water vapour leads to a warming of the region (Brasseur and Solomon, 2005).

P24, L888: Discuss Figure 12 a bit more. The figures should be first described. What is shown, what do you see…

This part has been rewritten:

Forcing the model with SSTs from the double CO_2-climate (as in experiment S1 and S2) yields 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 (as in experiment C2). Figure 10 shows the temperature responses to changes in cloud (left) and albedo (right) in July (top) and January (bottom) for experiment S1, as calculated by CFRAM.

Fig. 10 shows 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 10a). In January, the pattern looks slightly different (see Figure 10c). These temperature changes are due to changes in the balance between the increased reflected shortwave radiation and the decrease of outgoing longwave radiation.

P24, L893: Which cooling are you talking about? Up to date cooling for future cooling?

We were talking about the cooling as calculated by CFRAM. This paragraph has now been rewritten.

We also see an effect of the changes in surface albedo in the stratosphere (see Figure 10 b and d). The cooling in the summer polar stratosphere shown in Figure 10 b and d is due to radiative changes. We suggest that this cooling is due to a decrease in surface albedo, which would lead to less shortwave radiation being reflected. However, more research is needed.

P26, L924: Add a number. How much stronger?

This paragraph has been rewritten: *We find that the temperature change due to the direct forcing of CO_2 increases with increasing height in the middle atmosphere. The temperature change in the lower stratosphere due to the direct forcing of CO_2 is around 3 K. In the upper stratosphere, the cooling due to the direct forcing of CO_2 is about 9 K, which is considerably stronger than in the lower stratosphere. In the mesosphere, the cooling due to the direct forcing of CO_2 is about 15 K.*

General comment: How do the results agree with previous studies? What is the importance of you results for future predictions or climate change etc.? This should also be discussed.

The discussion and conclusion section has been rewritten to make these aspects clearer.

5.0 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 is important to note that no other method before has been able to quantify how much of the temperature change in the middle atmosphere is due to the different feedback processes before.

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 have given an overview of the mean temperature responses to the changes in CO_2 and various feedback processes in the lower stratosphere, upper stratosphere and in the mesosphere in January and July. We find that the temperature change due to the direct forcing of CO_2 increases with increasing height in the middle atmosphere. The temperature change in the lower stratosphere due to the direct forcing of CO_2 is around 3 K. There is a stronger cooling in the tropical lower stratosphere of about 4 K in July and 3.5 K in January.

In the upper stratosphere, the cooling due to the direct forcing of CO_2 is about 9 K, which is considerably stronger than in the lower stratosphere. The cooling is stronger in the summer polar regions, where the cooling reaches a value of 11K, than in the winter polar region, where the cooling is only about 8K. In the mesosphere, the cooling due to the direct forcing of CO_2 is even stronger: 15 K.

The ozone concentration changes due to changes in the CO_2-concentration as well as by changes in the SSTs. The temperature changes caused by this change in ozone concentration generally mitigate the cooling caused by the direct forcing of CO_2. However, in the tropical lower stratosphere and in some regions of the mesosphere, the ozone feedback cools these regions further. In the tropical lower stratosphere, for example, there is a cooling of 1K due to the ozone feedback.

We also have seen that the global mean temperature response due to dynamical feedbacks is small in the global average in all regions: less than 1 K. However, local responses to the changes in dynamics can be large. Doubling the CO_2 leads to a stronger summer-to-winter-pole flow, which leads to a cooling of the summer mesosphere and a warming of the winter mesosphere. Changing the SSTs weakens this effect in the mesosphere, but affects the temperature response in the stratosphere and lower mesosphere.

Using CFRAM on WACCM data shows that the change in water vapour leads to a cooling of up to 2 K in the lower stratosphere. It should be noted that climate models currently have a limited representation of the processes determining the distribution and variability of lower stratospheric water vapour. This means that the temperature response to the water vapour feedback might be different using a different model. We have also seen a small effect of the cloud and albedo feedback on the temperature response in the lower stratosphere, while these feedbacks play no role in the upper stratosphere and the mesosphere.

The results seen in this study are consistent with earlier findings. As in Shepherd et al., (2008), we find that the higher the temperature at a region in the atmosphere, the more cooling there is seen due to the direct feedback of CO_2. We find, as in Zhu et al., (2016) that 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 changes ozone concentration instead.

We have also seen that the ozone feedback generally yields a radiative feedback that mitigates the cooling, which is due to the direct forcing of CO_2, which is consistent with earlier studies such as Jonsson et al., (2004), Dietmüller et al., (2014). CFRAM is the first study that allows for calculating how much of the temperature response is due to which feedback process.

The next step would be 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. A better understanding of the effect of the increased CO_2-concentration on the middle atmosphere, will help to distinguish the effects of the changes CO_2- and O_3-concentration.

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 on the global mean temperature is currently not clear (Nowack et al., 2015, Marsh et al., 2016). A similar analysis as in this paper can be performed 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 there is a relatively large error due to the linearization in the model. 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.

To summarize you results and for having it easier with the discussion you could make a table/matrix where you mark which feedbacks are important in which altitude/latitude region.

We think that such a table would basically be a copy of Fig. 10. Instead of adding a new table, we now refer to Figure 10.

Technical corrections:

P2, 75: Add before after “in this way”. Changed
P4, L164: are -> were Changed
P4, L169: by -> with Changed
P5, L207: earth -> Earth Changed
P5, L218: SSTs forcing -> SST forcings Changed
P7, L280: delete “to” before balance Changed
P12, L505: don’t -> do not Changed
P15, L577: leads to -> leads to a Changed
P17, L619: One “the” obsolete. Changed
P18, L681: synoptic -> synoptic scale Changed
P21, L771: signifance -> significance Changed
P21, determinged -> determined Changed
P21, L776: CLOx -> ClOx Changed
P23, L824: to lead a -> to lead to a Changed
P23, L828: Delete “in WACCM”, you have already written it at the beginning of the sentence. Changed
P23, L838: aren’t -> are not Changed
P23, L841: “neither” obsolete? Or should it rather read “either”. Changed
P23, L848: found by -> simulated by or simulated with Changed
P25, L911: at a first -> as a first: This has been rephrased.
P26, L939: leads to -> leads to a Changed

Responses to Reviewer 2

General comments

The paper is now much clearer but there are still minor revisions necessary, especially concerning atmospheric chemistry.

We thank reviewer 2 for his/her comments.

Specific comments

Section 2.2, line 214ff: It is still not clear if the radiatively and chemically active gases CH$_4$ and N$_2$O (sometimes called well-mixed greenhouse gases) are kept at preindustrial levels in all scenarios at the surface (include in Table 1). It should be also mentioned that there are no CFCs in the atmosphere (I hope this is the case, otherwise major revision necessary).

Are the boundary conditions for all anthropogenic emissions kept constant in all scenarios?

Section 2.2 has been rewritten completely. We explain that in this study, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions. This compset simulates an equilibrium state, which means that it runs a perpetual year 1850. Four experiments have been performed (see Table 1).

The compset used in all experiments is still F_1850, which means that other radiatively and chemically active gases, such as ozone, will change only because of the changes in the CO_2-concentration, due to WACCM's interactive chemistry. CFCs don't play a role in our experiments.

The remark on inconsistencies in SST in the reply should be included in the text.

It has now been added that the fact that CESM doesn't include atmospheric chemistry, is not consideration in our study.

Line 305ff: It is not a good idea to use the same letters for different physical quantities (power flux and heating rate).

We use W/m^2 as the units of heating rates (per unit of area) for the air layer between two adjacent vertical levels, which is equivalent to heating rate per unit volume, instead of K/day, which is heating rate per unit mass. Because the radiative heating rates are the convergence of radiative energy fluxes entering/leaving the layer, it is rather natural and straightforward (meaning "without extra steps") to have the same units of heating rates (convergence) as the radiative energy fluxes. Throughout, the symbol S_vector (denote S with the vector on the top) represents of the convergence of short wave radiative fluxes entering each layer (thereby positive for heating), whereas R_vector represents of the divergence of long wave radiative fluxes leaving each layer (thereby positive for cooling). We use vector to represent the heating rates or a thermodynamical variable (e.g., T) in the vertical layers for mathematical convenience such as the DR/DT would be a matrix. In other words, the vector has no sense of direction in the physical world as it does not correspond to the upward or downward radiative fluxes.

In the manuscript, we have make it clear the S_vector is the vertical profile of the convergence of shortwave radiative energy fluxes, corresponding to the heating rates due to absorption of shortwaves in units of W/m^2, whereas R_vector is the vertical profile of the divergence of longwave radiative energy fluxes, corresponding to the cooling rates due to net emission of longwave radiation in units of W/m^2.

Also the constant 'g' is not defined (gravitational acceleration?).

g is indeed the gravitational acceleration, this has been added now.

Line 669: Cite the original references also here, they are in the reference list.

These references have been added. (This part has been moved to the introduction).

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (Brewer, 1949; Dobson, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (Zhu et al., 2016).

Line 708: Only non-orographic gravity waves are modulated by convection. Please rearrange paragraph.

This paragraph has been rewritten:

The warmer sea surface temperatures affect the dynamics in the middle atmosphere. It has for example been shown that higher SSTs in the tropics leads to an amplification in deep convection, which enhances the generation of quasi-stationary waves (Deckert and Dameris, 2008). Enhanced SSTs lead to an enhanced dissipation of planetary waves, as well as an enhanced dissipation of orographic and non-orographic waves in the upper stratosphere (Oberländer et al., 2013).

Line 745ff: please rewrite: HOx dominates ozone destruction in the mesosphere and lower stratosphere, NOx and Clx in the middle and upper stratosphere.

This part has been rewritten:

Ozone plays a crucial role in the chemical and radiative budget of the middle atmosphere. The distribution of ozone in the middle atmosphere is determined by both chemical and dynamical processes. Most of the ozone production takes place in the tropical stratosphere, as a result of photochemical processes, which involve oxygen. Meriodional circulation then transports ozone to higher latitudes (Langematz, 2019). The production of ozone is largely balanced by catalytic destruction cycles involving NO_x, HO_x and Cl_x radicals. HO_x dominates ozone destruction in the mesosphere and lower stratosphere, while NO_x and Cl_x dominate this process in the middle and upper stratosphere (Cariolle, 1983).

Insert in line 754 "due to the strong temperature dependence of the Chapman reaction O+O$_3$".

This has been added:

In this study, we are interested in the temperature response to changes in ozone concentration induced by the increased CO_2 concentration and/or the changes in SST in WACCM. Under enhanced CO_2 concentrations, the ratio between O_3 and O mixing ratios is generally shifted toward a higher concentration of ozone, which is caused by the strong temperature dependency of the ozone production reaction ($O + O_2 + M \rightarrow O_3 + M$).

In the preindustrial stratosphere are no aircraft NOx-emissions and no CFCs (line 755). Cl should be only from CH$_3$Cl (forest fires etc.) This is misleading here.

This sentence has been removed.

Line 773ff: Please rewrite and include something like the following: If O/O$_3$ is shifted to lower values due to cooling the catalytic ozone destruction cycles are slower since the reactions of the radicals (NO$_2$, OH, HO$_2$, ClO) with O are the rate limiting steps.
This has been rewritten (See below)

Line 783: Why? This is secondary if the scenarios are consistent. There are a lot of textbooks and review papers on stratospheric and mesospheric chemistry.

This sentence has been removed.

Thanks, we have consulted them and we refer to Schmidt et al. as they did a similar study as we did.

Line 788: This is very special and only relevant in case of perturbations by solar particles at solar maximum conditions. Chlorine should be low (in total < 0.6ppbv).

Indeed the role if Cl should not be large, we have added that now. We would like to refer to Schmidt et al., 2006. They don't use the WACCM model, but HAMMONIA, however this is coupled to the same chemical model MOZART3. In their study, they perform a similar experiment as we do in our study, however they start from 360 ppm of CO2 and double to 720, while we start from the pre-industrial level of CO2 (280 ppm). Our results are very similar. Schmidt et al. (2006) write that "The ozone decrease in the polar winter around 0.1 hPa (about 65 km) is mainly caused by the increase of NO and (to a lesser extent) Cl mixing ratios due to stronger subsidence of NO and Cl-rich air." We see a similar increase of Cl as they do.

The section about the ozone feedback has been rewritten completely. We hope that is now clear what is done.

Ozone plays a crucial role in the chemical and radiative budget of the middle atmosphere. The distribution of ozone in the middle atmosphere is determined by both chemical and dynamical processes. Most of the ozone production takes place in the tropical stratosphere, as a result of photochemical processes, which involve oxygen. Meridional circulation then transports ozone to other parts of the middle atmosphere (Langematz, 2019). The production of ozone is largely balanced by catalytic destruction cycles involving NO_x, HO_x and Cl_x radicals. HO_x dominates ozone destruction in the mesosphere and lower stratosphere, while NO_x and Cl_x dominate this process in the middle and upper stratosphere (e.g. Cariolle, 1983).

Since the 1970s ozone in the middle atmosphere began to decline globally, due to increased production of ozone depleting substances (ODSs) (Brühl and Crutzen, 1988). The Montreal Protocol, adopted in 1987 to stop this threat, eventually led to a slow recovery of the stratospheric ozone over recent two decades (WMO, 2018; Langematz, 2019). In our study, we don't consider the effect of anthropogenic ODSs since pre-industrial times (Langematz, 2019).

In this study, we are interested in the temperature response to changes in ozone concentration induced by the increased CO_2 concentration and/or the changes in SST in WACCM. Under enhanced CO_2 concentrations, the ratio between O_3 and O mixing ratios is generally shifted toward a higher concentration of ozone, which is caused by the strong temperature dependency of the ozone production reaction ($O + O_2 + M \rightarrow O_3 + M$).

Fig. 3 shows the percentage changes in O_3-concentration when the CO_2-concentration and/or the SSTs change. We observe indeed that 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. 3c and f show that changing the SSTs also has a significant impact on the ozone concentration. A 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.

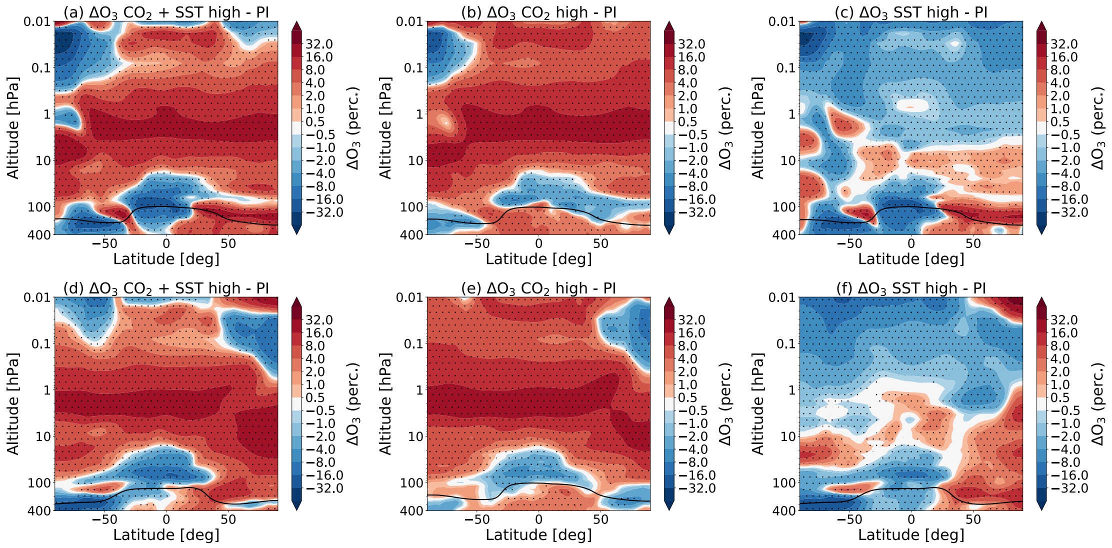

Figure 3: The percentage change in the zonal and monthly mean ozone concentration for July (top) and January (bottom) due to (a,d) combined effect of the CO_2 increase and SSTs changes (experiments C2 -B1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments B2 - B1) and the (c,f) SSTs (experiments C1 - B1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is indicated as in Fig. 1.

As we will discuss in the next section, an enhanced concentration of CO_2 also leads to changes in the dynamics in the middle atmosphere. The stratospheric Brewer-Dobson circulation is projected to strengthen, which would lead to an increase in the poleward transport of ozone. We will also see that an increase in CO_2-concentration leads to stronger summer pole-to-winter pole flow in the mesosphere.

Figure 4 shows the percentage change in the zonal and monthly mean concentration of Cl, NO, O, OH, CH_3, NO_x and N_2O in July due to combined effect of the CO_2 increase and SSTs changes (experiment (C2 vs B1), as simulated by WACCM. The patterns in January look similar (not shown).

We would like to point out that the changes in these constituents are only brought about by the CO_2-concentration and/or the SSTs. We still use the F_1850 compset and the only difference between the runs is the forcing in CO_2 and SSTs. The changes in chemical constituents look very similar to those found by Schmidt et al. (2006) who performed a similar experiment as discussed in section 2.2, see their Figure 20. Note that Fig. 4 shows the changes due to both the CO_2 increase and SSTs changes, while their Figure 20 shows the percentage changes due the changes in CO_2-concentration only.

As Schmidt et al. (2006), we see an decrease in atomic oxygen (O) mixing ratio at high summer latitudes around 0.01 hPa (see Fig. 4c), which results from increased upwelling. This increase in O leads to a decrease in ozone in this region. We also see decrease of ozone concentration in the winter polar region around 0.1 hPa (approximately 65 km). This could be caused an increase of NO and for a small part by Cl mixing ratios, which result from a stronger subsidence of NO and Cl rich air, as suggested in Schmidt et al. 2006. As stated before, complete discussion of the *changes in ozone concentration is out of the scope of this paper and the changes in other constituents shown in Figure 4 are shown for reference only.*

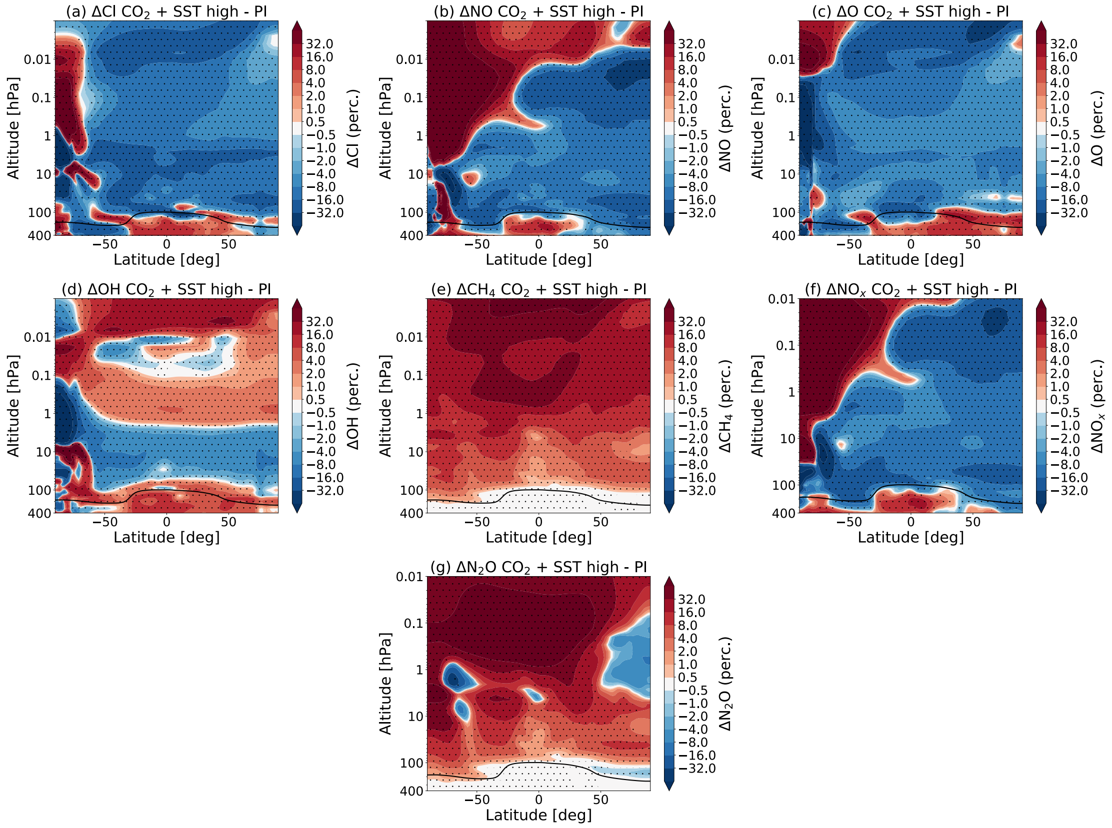

Figure 4: The percentage change in the zonal and monthly mean concentration of Cl (a), NO (b), O (c), OH (d), CH_3 (e) and NO_x (f) and N_2O (g) in July due to combined effect of the CO_2 increase and SSTs changes (experiment (C2 vs B1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is indicated as in Fig. 1.

What is new in this study, is that we can calculate the temperature responses due to the changes in ozone concentration. These temperature responses are shown in Figure 5. 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 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).

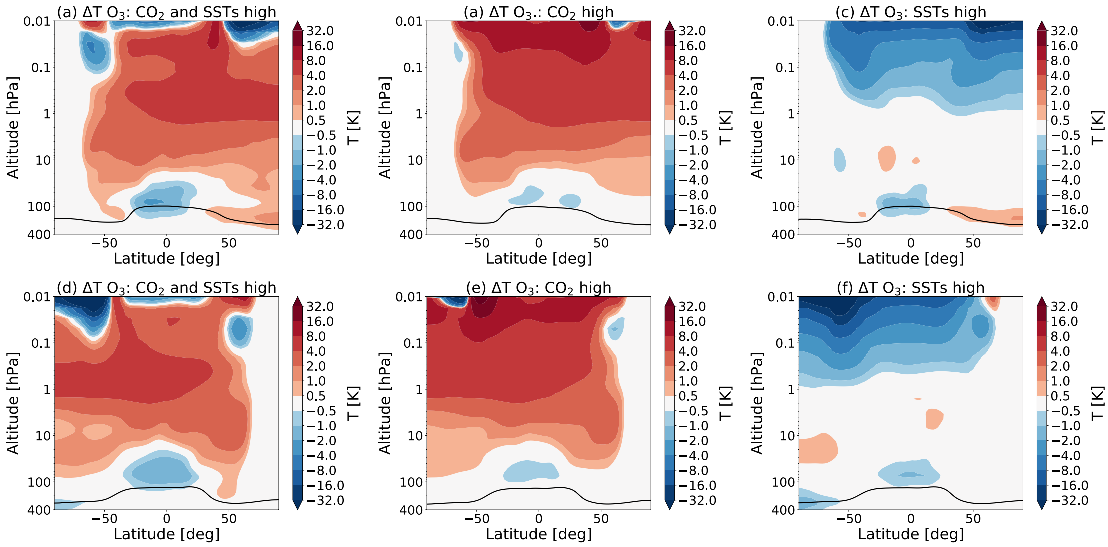

Figure 5: Partial temperature responses to changes in O_3-concentration, as calculated by CFRAM, in July (top) and January (bottom) due to the (a,d) combined effect of the CO_2 increase and SSTs changes (C1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments B2) and the (c,f) SSTs (experiments C1). The tropopause height is indicated as in Fig. 1.

Line 928: This sentence is a mess. Focus on temperature dependence of reaction rates.

This sentence has been rewritten. The aim of this paper is not the explain the changes in ozone concentration. As we only would like to calculate the temperature effect of these, so we have left this part out.

The ozone concentration changes due to changes in the CO_2-concentration as well as by changes in the SSTs. The temperature changes caused by this change in ozone concentration generally mitigate the cooling caused by the direct forcing of CO_2. However, we have also seen in that in the tropical lower stratosphere and in some regions of the mesosphere, the ozone feedback cools these regions further.

Line 944ff: This is uncertain, please include remark or leave out here.

A remark has been included.

Using CFRAM on WACCM data shows that the temperature change in the lower stratosphere is influenced by the water vapour feedback and to a lesser degree by the cloud and albedo feedback, while these feedbacks play no role in the upper stratosphere and the mesosphere. It should be noted that climate models currently have a limited representation of the processes determining the distribution and variability of lower stratospheric water vapour. This means that the temperature response to the water vapour feedback might be different using a different model.

Technical corrections

Table 1: use subscripts.

Subscripts added.

Line 710: Typo

Corrected.

Line 736: Subscripts

Subscripts added.

line 749: Typo, also include 'e.g.' in citation.

Corrected and e.g. added.

Ozone plays a crucial role in the chemical and radiative budget of the middle atmosphere. The distribution of ozone in the middle atmosphere is determined by both chemical and dynamical processes. Most of the ozone production takes place in the tropical stratosphere, as a result of photochemical processes, which involve oxygen. Meridional circulation then transports ozone to other parts of the middle atmosphere (Langematz, 2019). The production of ozone is largely balanced by catalytic destruction cycles involving NO_x*,* HO_x *and* Cl_x *radicals.* HO_x *dominates ozone destruction in the mesosphere and lower stratosphere, while* NO_x *and* Cl_x *dominate this process in the middle and upper stratosphere (e.g. Cariolle, 1983).*

References: Several entries are incomplete, e.g. page numbers and/or DOI missing.

DOI has been added for all entries. Page number added for those which had page numbers specified.

Line 1121: incomplete journal name.

This reference is no longer there.

Response to reviewer 3:

I am sorry to say that I still have reservations about the paper in its present form. I am afraid that there are some rather fundamental disagreements between the authors and the reviewer so that my comments are rather general.

We thank reviewer 3 for his/her comments.

My major point is that the paper is still missing a clear message, going beyond what is already known.

This is the first study that allows for calculating how much of the temperature response in the middle atmosphere is due to which feedback process. We have now also made clearer in the introduction what we mean with feedbacks in this paper.

Yes, in a doubled CO2 climate, the direct radiative forcing of CO2 will lead to a cooling in the upper stratosphere, but this is certainly well known (e.g. WMO 2018).

What is new in this paper, is that we calculate exactly how much of the temperature change at a certain point of the middle atmosphere is due to which processes. We can calculate the temperature response due to the direct forcing of CO2, as well as the temperature responses due to the changes in ozone concentration, water vapour concentration, clouds, albedo and dynamics. This is the first paper that applies the CFRAM method to the middle atmosphere in this way and thus first study that allows for a calculation the partial temperature responses.

The cooling in the upper stratosphere, by affecting upper stratospheric ozone chemistry (but which ozone chemistry exactly, see below), leads to an increase in ozone (referred to in the abstract as "ozone feedback", which leads to an increase in short-wave heating (i.e. yields a radiative feedback that generally mitigates this cooling) -- I agree, but again, what is new here?

What is new in this paper, is that we calculate exactly how much of the temperature change at a certain point of the middle atmosphere is due the changes in ozone concentration. More traditional methods of studying the ozone feedback in the middle atmosphere don't allow for calculating the temperature response to the changes in ozone concentration.

The abstract states that the "temperature response due to dynamical feedbacks is small in global average", but I find it difficult to understand which dynamical feedbacks are meant here -- perhaps enhanced tropical upwelling in a 2xCO2 run?

This has now been made clear in the introduction. CFRAM allow to calculate the radiative contribution to the temperature response for different feedback processes. However, changes in the dynamics also influence the temperature response and are calculate separately (see also equation A4 in the appendix and *Zhu et al*., 2016 for comparison).

This has been added to the introduction:
In this study, we investigate the effects of doubling the CO_2-concentration and the accompanying sea surface temperature change on the temperature in the middle atmosphere as compared to the pre-industrial state. We use CFRAM to calculate the radiative contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, and clouds. We refer to the changes in ozone, water vapour, albedo and clouds in response to changes in the CO_2-concentration as the ozone, water vapour, albedo and cloud feedbacks.

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (Brewer, 1949; Dobson, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (Zhu et al., 2016).

The main goal of this paper is to calculate the contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, clouds and dynamics in the middle atmosphere under a double CO_2-scenario using CFRAM. Our intention is not to give a complete account of the exact mechanisms behind the changes in ozone, water vapour, albedo, clouds and dynamics.

Further, it is stated that the "temperature change in the lower stratosphere is influenced by the water vapour feedback", but again, the processes in question here remain unclear. I suggest to state at least what the sign of the temperature change is (increase or decrease?) and what "water vapour feedback" means (increase or decrease of water vapour? chemical impact of water vapour on ozone or radiative effects of water vapour?).

We have quantified this now and added the caveat that this might be model dependent.

Using CFRAM on WACCM data shows that the change in water vapour leads to a cooling of up to 2 K in the lower stratosphere. It should be noted that climate models currently have a limited representation of the processes determining the distribution *and variability of lower stratospheric water vapour. This means that the temperature response to the water vapour feedback might be different using a different model.*

In the introduction we have now also made clear what we mean with feedbacks:

In this study, we investigate the effects of doubling the CO_2-concentration and the accompanying sea surface temperature change on the temperature in the middle atmosphere as compared to the pre-industrial state. We use CFRAM to calculate the radiative contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, and clouds. We refer to the changes in ozone, water vapour, albedo and clouds in response to changes in the CO_2-concentration as the ozone, water vapour, albedo and cloud feedbacks.

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (Brewer, 1949; Dobson, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (Zhu et al., 2016).

What remains is an evaluation of CFRAM as a tool to study climate change. There could be new developments here, but if the focus were on CFRAM than the nature of the paper would be much more methodological that it is in its present form.

CFRAM is the first study that allows for calculating how much of the temperature response is due to which feedback process. The main goal of this paper is to calculate the contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, clouds and dynamics in the middle atmosphere under a double CO_2-scenario using CFRAM. Our intention is not to give a complete account of the exact mechanisms behind the changes in ozone, water vapour, albedo, clouds and dynamics. We have given a detailed account of the method, but the focus is on the results.

I see also remaining disagreements between the authors and the reviewer judging from the reply to my comments.

The authors state "We are not speaking here about the changes in O3-concentration due to the ozone hole, but rather changes in ozone concentration that are resulting from changes in the CO2 concentration." I did not talk about the "ozone hole" either, what I am talking about is the upper stratospheric ozone loss, which is also chlorine driven. And enhanced levels of stratospheric chlorine are around for many decades to come. From reading the manuscript, I assume that Cl was set to zero, or perhaps 0.6 ppb -- but I am not sure (see also below). I think this issue could be at least briefly addressed (if chlorine is not relevant for the present study it should be stated in the paper). Other chemical effects could be due to changing N2O levels. I think it is not a good idea to remove the reference to the WMO ozone assessment (WMO 2018) entirely.

In our experiment, the levels of Cl and other fields are set to pre-industrial levels (as in the F_1850 compset of WACCM) in WACCM. However, the increase in CO_2-concentration brings about different concentrations of reactive consituents affecting the ozone concentration. The changes in the above-mentioned constituents are simulated using the interactive chemistry as in the MOZART3 model (*Kinnison et al.,* 2007).

In the paper we have now given an overview of the percentage change in the zonal and monthly mean concentration of Cl (a), NO (b), O (c), OH (d), CH_4 (e) and NO_x (f) and N_2O (g) in July due to combined effect of the CO_2 increase and SSTs changes (experiment (C2 vs B1), as simulated by WACCM.

The reference to the WMO ozone assessment has been added.

I stated in my previous review that "it should be clearly said that the middle atmosphere is *not* in radiative equilibrium" -- in response the authors changed the discussion, which clearly improved the presentation of this aspect. However I find the sentence "In the absence of eddy motions the zonal-mean temperature would relax to a radiatively determined state" still misleading. This sounds a bit as this were still a possible state of the atmosphere; note that without "eddy motions", i.e. without atmospheric waves, in a radiative equilibrium there are no heating or cooling terms, i.e. no transport across isentropic surfaces, in other words no Brewer-Dobson circulation. (And I think that throughout most of the troposphere, outside of convection, the troposphere is not dynamically unstable).

This sentence has now been removed.

A few details:

l 214: WACCMs chemical model is associated here with CO2 changes? What is the CO2 chemistry in WACCM? Or is it transport and changing CO2 emissions that are relevant here?

A more detailed description of the experimental setup has now been added (see below).

We would also like to refer to Schmidt et al., 2006. They don't use the WACCM model, but HAMMONIA, however this is coupled to the same chemical model MOZART3. In their study, they perform a similar experiment as we do in our study, however they start from 360 ppm of CO2 and double to 720, while we start from the pre-industrial level of CO2 (280 ppm). Our results are very similar.

In this study, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions. This compset simulates an equilibrium state, which means that it runs a perpetual year 1850. Four experiments have been performed for this study (see Table 1).

Experiment C1 is the control run, with the pre-industrial CO_2 concentration (280 ppm) and forced with pre-industrial ocean surface conditions such as sea surface temperature and sea ice (referred to SSTs from now on). These SSTs are generated from the CMIP5 pre-industrial control simulation by the fully coupled Earth system model CESM. The atmospheric component of CESM is the same as WACCM, but does not include stratospheric chemistry (Hurrell et al., 2013). This latter aspect is not considered in this study.

Experiment C2 represents the experiment with the CO_2 concentration doubled as compared to the pre-industrial state (560 ppm) and forced with the same pre-industrial SSTs as in experiment C1. 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.

The compset used in this experiment and all the following ones is still F_1850, which means that other radiatively and chemically active gases, such as ozone, will change only because of the changes in the CO_2-concentration, due to WACCM's interactive chemistry. Chlorofluorocarbons (CFCs), which have a major impact on the ozone concentration in the real atmosphere, don't play a role in our experiments.

In experiment S1, we simulate the scenario, in which there is the SSTs forcing from the coupled CESM for double CO_2 condition, but the pre-industrial CO_2-concentration of 280 ppm. S2 represents the experiment with the CO_2-concentration in the atmosphere doubled to 560 ppm and the SSTs prescribed for the double CO_2-climate. Experiment C1, C2, S1 and S2 will be also referred to hereafter by PI, the simulation with high CO2, the simulation with high SSTs and the simulation with high CO_2 and SSTs, respectively.

The experimental setup of this study is similar to the setup performed with the Canadian Middle Atmosphere Model (CMAM) by Fomichev et al. (2007) and with the Hamburg Model of the Neutral and Ionized Atmosphere (HAMMONIA) by Schmidt et al. (2006). The HAMMONIA model is coupled to the same chemical model as WACCM: MOZART3. The setup in their study is similar, however, in their study, they double the CO_2-concentration from 360 ppm to 720 ppm, while in our study, we double from the pre-industrial level of CO_2 (280 ppm).

l. 224: I think you need more documentation here on the WACCM run. I see pre-industrial CO2 and doubled CO2 (also SSTs are mentioned), but a lot of other fields are unclear;
e.g. stratospheric chlorine, N2O, CH4. Aerosol loading? Are these compounds (and the entire setup) based on "pre-industrial"? Could tropospheric ozone be relevant (different between the runs). Perhaps you could use a citation, where all the assumptions of the WACCM run are described? Note that when you use pre-industrial with (just) CO2 doubled, these doubled CO2 runs do not describe a future

A more detailed description of the experimental setup has now been added (the response the to the previous question).

It is correct that the doubles CO2 run don't describe a realistic future scenario. However, these experiments can still teach us something. We write this in the paper as well (in the introduction);

We acknowledge that such idealized equilibrium simulation cannot reproduce the complexity of the atmosphere, in which the CO_2-concentration is changing gradually. However, simulating a double CO_2-scenario still allows us to identify robust feedback processes in the middle atmosphere.
We mention this now also in section 2.2:

Note that experiment S1 and C2 are not representing scenarios that could happen in the real atmosphere. These experiments have been used to study the effect of the SSTs separately. Experiment S2 doesn't take into account other (anthropogenic) changes in the atmosphere not caused by changes in the CO_2-concentration and the SSTs.

We would also like to refer to *Schmidt et al.,* 2006. They don't use the WACCM model, but HAMMONIA, however this is coupled to the same chemical model: MOZART3. In their study, they perform a similar experiment as we do in our study, however they start from 360 ppm of CO2 and double to 720, while we start from the pre-industrial level of CO2 (280 ppm). Our results are very similar.

In our experiment, the levels of Cl and other fields are set to pre-industrial levels (as in the F_1850 compset of WACCM) in WACCM. However, the increase in CO_2-concentration brings about different concentrations of reactive consituents affecting the ozone concentration. The changes in the above-mentioned constituents are simulated using the interactive chemistry as in the MOZART3 model (*Kinnison et al.,* 2007).

In the paper we have now given an overview of the percentage change in the zonal and monthly mean concentration of Cl (a), NO (b), O (c), OH (d), CH_3 (e) and NO_x (f) and N_2O (g) in July due to combined effect of the CO_2 increase and SSTs changes (experiment (C2 vs B1), as simulated by WACCM.

l 745-748: You state "Ozone plays a major role in the chemical and radiative budget of the middle atmosphere. The ozone distribution in the mesosphere is maintained by a balance between transport processes and various catalytic cycles involving nitrogen oxides, HOx and Clx radicals." Which transport processes with an impact on ozone are you referring to here?

This part has been rewritten:

Ozone plays a crucial role in the chemical and radiative budget of the middle atmosphere. The distribution of ozone in the middle atmosphere is determined by both chemical and dynamical processes. Most of the ozone production takes place in the tropical stratosphere, as a result of photochemical processes, which involve oxygen. Meriodional circulation then transports ozone to other parts of the middle atmosphere (Langematz, 2019). The production of ozone is largely balanced by catalytic destruction cycles involving NO_x*,* HO_x *and* Cl_x *radicals.* HO_x *dominates ozone destruction in the mesosphere and lower stratosphere, while* NO_x *and* Cl_x *in the middle and upper stratosphere (Cariolle, 1983).*

Since the 1970s ozone in the middle atmosphere began to decline globally, due to increased production of ozone depleting substances (ODSs). The Montreal Protocol, adopted in 1987 to stop this threat, eventually led to a slow recovery of the stratospheric ozone over the recent two decades. In our study, we don't consider the effect of anthropogenic ODSs since pre-industrial times (Langematz, 2019).

In this study, we are interested in the temperature response to changes in ozone concentration induced by the increased CO_2 *concentration and/or the changes in SST in WACCM. Under enhanced* CO_2 *concentrations, the ratio between* O_3 *and O mixing ratios is generally shifted toward a higher concentration of ozone, which is caused by the strong temperature dependency of the ozone production reaction (*$O + O_2 + M \rightarrow O_3 + M$*)*

Major changes made to the manuscript:

- Substantial rewriting of abstract, introduction, model and methods.
- Restructuring of section 3, 4 and 5.
- More quantification of our results in the text.
- Relating clearer to earlier studies and more emphasis on the fact that this is the first method to calculate the partial temperature changes at each point of the atmosphere.
- Clearer explanation of the atmospheric chemistry of WACCM and the changes in other chemical constituents due to changes in CO_2 and/or SSTs.

Using the climate feedback response method to quantify climate feedbacks in the middle atmosphere in WACCM

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Abstract

Over recent decades it has become clear that the middle atmosphere has a significant impact on surface and tropospheric climate. A better understanding of the middle atmosphere and how it reacts to the current increase of the concentration of carbon dioxide (CO_2) is therefore necessary. In this study, we investigate the response to a doubling CO_2 in the middle atmosphere using the Whole Atmosphere Community Climate Model (WACCM). We use the climate feedback response analysis method (CFRAM) to calculate the partial temperature changes due to an external forcing and climate feedbacks in the atmosphere. As this method has the unique feature of additivity, these partial temperature changes are linearly addable. In this study, we discuss the direct forcing of CO_2 and the effects of the ozone, water vapour, cloud, albedo and dynamical feedbacks.
As expected, our results show that the direct forcing of CO_2 cools the middle atmosphere. This cooling becomes stronger with increasing height; the cooling in the upper stratosphere is about three times as strong as the cooling in the lower stratosphere. The ozone feedback yields a radiative feedback that mitigates this cooling in most regions of the middle atmosphere. However, in the tropical lower stratosphere and in some regions of the mesosphere, the ozone feedback has a cooling effect. The increase in CO_2-concentration causes the dynamics to change. The temperature response due to this dynamical feedback is small in the global average, although there are large temperature changes due to this feedback locally. The temperature change in the lower stratosphere is influenced by the water vapour feedback and to a lesser degree by the cloud and albedo feedback. These feedbacks play no role in the upper stratosphere and the mesosphere. We find that the effects of the changed SSTs on the middle atmosphere are relatively small as compared to the effects of changing the CO_2. However, the changes in SSTs are responsible for dynamical feedbacks that cause large temperature

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In a double CO_2 climate, the direct forcing of CO_2 would lead to a cooling which increases with increasing height in
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The ozone feedback yields a radiative feedback that generally mitigates this cooling. 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 only of importance in the lower stratosphere.
CFRAM is very powerful tool to study climate feedbacks in the middle atmosphere however, there is an error term caused by the linearization in the method. ... [1]
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changes. Moreover, the temperature response to the water vapour feedback in the lower stratosphere is almost solely due to changes in the SSTs. As CFRAM has not been applied to the middle atmosphere in this way before, this study also serves to investigate the applicability as well as the limitations of this method. This work shows that CFRAM is a very powerful tool to study climate feedbacks in the middle atmosphere. However, it should be noted that there is a relatively large error term associated with the current method in the middle atmosphere, which can be for a large part be explained by the linearization in the method.

1. Introduction

The increase of concentration of carbon dioxide in the atmosphere forms a major perturbation to the climate system. It is commonly associated with lower-atmospheric warming. However, in the middle atmosphere, the increase of CO_2 leads to a cooling of the region instead. This cooling has been well documented and is found by both model studies and observations (e.g. *Manabe and Wetherald*, 1975; *Ramaswamy et al.*, 2001; *Beig et al.*, 2003).

The middle atmosphere is not only affected by the increase in CO_2-concentration, but also by the decrease in ozone-concentration. The depletion of ozone (O_3) also effects the temperature in the stratosphere and leads to a cooling (*Shine et al*, 2003). A better understanding of the effect of the increased CO_2-concentration on the middle atmosphere, will help to distinguish the effects of the changes CO_2- and O_3-concentration.

Another major motivation for this study is the emerging evidence that the middle atmosphere has an important influence on surface and tropospheric climate (*Shaw and Shepherd*, 2008). It has, for example, been shown that cold winters in Siberia are linked to changes in the stratospheric circulation (*Zhang et al.*, 2018).

Nowack et al. (2015) has found that there is an increase in global mean surface warming of about 1°C when the ozone is prescribed at pre-industrial levels, as compared with when it is evolving in response to an abrupt 4xCO_2 forcing. It should be noted that 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).

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

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Deleted: 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. ¶

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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.

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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 climate 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).

The climate feedback-response analysis method (CFRAM) is an alternative method which takes into account that the climate change is not only determined by the energy balance at the top of the atmosphere, but is also influenced by the energy flow within the Earth's system itself (*Cai and Lu, 2009, Lu and Cai*, 2009). 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. This makes it possible to calculate the partial temperature changes due to an external forcing and these internal feedbacks in the atmosphere. It has the unique feature of additivity, such that these partial temperature changes are linearly addable.

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). CFRAM has been applied to study the middle atmosphere climate sensitivity as well (*Zhu et al.,* 2016). In their study, Zhu et al. *(2016)* have adapted CFRAM and applied it to both model output, as well as observations. The atmospheric responses during solar maximum and minimum were studied and it was found that the variation in solar flux forms the largest radiative component of the middle atmosphere temperature response.

In the present work, we apply CFRAM to climate sensitivity experiments performed with 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 middle atmosphere response to CO_2-doubling. We acknowledge that such idealized equilibrium simulation cannot reproduce the complexity of the atmosphere, in which the CO_2-concentration is changing

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gradually. However, simulating a double CO_2-scenario still allows us to identify robust feedback processes in the middle atmosphere.

There are two aspects of the middle atmosphere response to CO_2-doubling: there is the effect of the changes in CO_2-concentration directly, as well as the changes in sea surface temperature (SST) which are in itself caused by the changes in CO_2-concentration. It is useful to investigate these aspects separately, as former should be robust, while the effect of the changed SST depends on the changes in tropospheric climate, which can be expected to depend more on the model.

In this study, we investigate the effects of doubling the CO_2-concentration and the accompanying sea surface temperature change on the temperature in the middle atmosphere as compared to the pre-industrial state. We use CFRAM to calculate the radiative contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, and clouds. We refer to the changes in ozone, water vapour, albedo and clouds in response to changes in the CO_2-concentration as the ozone, water vapour, albedo and cloud feedbacks.

The circulation in the middle atmosphere is driven by waves. Wave forcing drives the temperatures in the middle atmosphere far away from radiative equilibrium. In the mesosphere, there is a zonal forcing, which yields a summer to winter transport. In the polar winter stratosphere, there is a strong forcing that consists of rising motion in the tropics, poleward flow in the stratosphere and sinking motion in the middle and high latitudes. This circulation is referred to as the 'Brewer-Dobson circulation' (*Brewer*, 1949; *Dobson*, 1956).

Dynamical effects make important contributions to the middle-atmosphere energy budget, both through eddy heat flux divergence and through adiabatic heating due to vertical motions. It is therefore important that we also consider changes to the middle-atmosphere climate due to dynamics. We refer to this as the 'dynamical feedback' (*Zhu et al.*, 2016).

The main goal of this paper is to calculate the contribution to the temperature change due to changes in carbon dioxide directly as well as due to changes in ozone, water vapour, albedo, clouds and dynamics in the middle atmosphere under a double CO_2-scenario using CFRAM. Our intention is not to give a complete account of the exact mechanisms behind the changes in ozone, water vapour, albedo, clouds and dynamics.

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 vertical resolution, which ranges from ~1.1 km in the troposphere,

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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 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 (MOZART3). 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, the F_1850 compset (component set) of the model is used, i.e. the model assumes pre-industrial (PI) conditions. This compset simulates an equilibrium state, which means that it runs a perpetual year 1850. Four experiments have been performed for this study (see Table 1).

Experiment C1 is the control run, with the pre-industrial CO_2 concentration (280 ppm) and forced with pre-industrial ocean surface conditions such as sea surface temperature and sea ice (referred to SSTs from now on). These SSTs are generated from the CMIP5 pre-industrial control simulation by the fully coupled Earth system model CESM. The atmospheric component of CESM is the same as WACCM, but does not include stratospheric chemistry (*Hurrell et al.,* 2013). This latter aspect is not considered in this study.

Experiment C2 represents the experiment with the CO_2 concentration doubled as compared to the pre-industrial state (560 ppm) and forced with the same pre-industrial SSTs as in experiment C1. 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.

The compset used in this experiment and all the following ones is still F_1850, which means that other radiatively and chemically active gases, such as ozone, will change only because of the changes in the CO_2-concentration, due to WACCM's interactive chemistry. Chlorofluorocarbons (CFCs), which have a major impact on the ozone concentration in the real atmosphere, don't play a role in our experiments.

In experiment S1, we simulate the scenario, in which there is the SSTs forcing from the coupled CESM for double CO_2 condition, but the pre-industrial CO_2-

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concentration of 280 ppm. S2 represents the experiment with the CO_2-concentration in the atmosphere doubled to 560 ppm and the SSTs prescribed for the double CO_2-climate. Experiment C1, C2, S1 and S2 will be also referred to hereafter by PI, the simulation with high CO2, the simulation with high SSTs and the simulation with high CO_2 and SSTs, respectively.

The experimental setup of this study is similar to the setup performed with the Canadian Middle Atmosphere Model (CMAM) *by Fomichev et al.* (*2007*) and with the Hamburg Model of the Neutral and Ionized Atmosphere (HAMMONIA) by *Schmidt et al.* (2006). The HAMMONIA model is coupled to the same chemical model as WACCM: MOZART3. The setup in their study is similar, however, in their study, they double the CO_2-concentration from 360 ppm to 720 ppm, while in our study, we double from the pre-industrial level of CO_2 (280 ppm).

Note that experiment S1 and C2 are not representing scenarios that could happen in the real atmosphere. These experiments have been used to study the effect of the SSTs separately. Experiment S2 doesn't take into account other (anthropogenic) changes in the atmosphere not caused by changes in the CO_2-concentration and the SSTs.

All the simulations are run for 50 years, of which the last 40 years are used for analysis. In the all results shown, we have used the 40 year mean of our model data.

Table 1. Set-up of the model experiments.

Experiment	CO_2	SSTs from CESM equilibrium run
C1	280 ppm	PI control
C2	560 ppm	PI control
S1	280 ppm	Double CO_2 run
S2	560 ppm	Double CO_2 run

2.3 Climate feedback-response analysis method (CFRAM)

In this study, we aim to quantify the different climate feedbacks that may play a role in the middle atmosphere in a double CO_2-climate. For this purpose, we apply the climate feedback-response analysis method (CFRAM) (*Lu and Cai*, 2009).

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 consideration 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

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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.

We refer to the Appendix for the complete formulation of CFRAM diagnostics using outputs of WACCM. Based on the linear decomposition principle, we can solve Eq. (A12) for each of the terms on its right-hand side. This yields the partial temperature changes due to each specific process namely:

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

$$\Delta \vec{T}_{O_3} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{O_3} \quad (2)$$

$$\Delta \vec{T}_{H_2O} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{H_2O} \quad (3)$$

$$\Delta \vec{T}_{albedo} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{albedo} \quad (4)$$

$$\Delta \vec{T}_{cloud} = \left(\frac{\partial \vec{R}}{\partial \vec{T}}\right)^{-1} \Delta\left(\vec{S} - \vec{R}\right)_{cloud} \quad (5)$$

The factors $\Delta(\vec{S} - \vec{R})_{CO_2}$, $\Delta(\vec{S} - \vec{R})_{O_3}$, $\Delta(\vec{S} - \vec{R})_{H_2O}$, $\Delta(\vec{S} - \vec{R})_{albedo}$ and $\Delta(\vec{S} - \vec{R})_{cloud}$ 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.

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Similarly, to equations (1)-(5), we also calculate the temperature change due to non-LTE processes and the dynamical feedback. We calculate the terms $\Delta(\mathcal{S}-\mathcal{R})_{non-LTE}$ and Δdyn in Eq. (A4) and (A7).

$$\Delta T_{non-LTE} = \left(\frac{\partial \mathcal{R}}{\partial T}\right)^{-1} \Delta(\mathcal{S}-\mathcal{R})_{non-LTE} \quad (6)$$

$$\Delta T_{dyn} = \left(\frac{\partial \mathcal{R}}{\partial T}\right)^{-1} \Delta dyn \quad (7)$$

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

$$\Delta T_{CFRAM} = +\Delta T_{O_3} + \Delta T_{H_2O} + \Delta T_{albedo} + \Delta T_{cloud} + \Delta T_{non-LTE} + \Delta T_{dyn} \quad (8)$$

The linearization done for equations (A9) and (A10) 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.

$$\Delta T_{CFRAM} = \Delta T_{WACCM} - \Delta T_{error} \quad (9)$$

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

Note that the method used in this study differs 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}). In other words, we use Wm^{-2} as the units of heating rates for the layer between two adjacent vertical levels. Because the radiative heating rates are the net radiative energy fluxes entering the layer, it is rather natural and straightforward (i.e., without dividing the mass in the layer to convert it to units of Ks^{-1}) to have the same units of heating rates (convergence) as the radiative energy fluxes. 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. Temperature responses in a double CO_2 scenario

As described in section 2.2, four experiments were performed with WACCM: a simulation with pre-industrial conditions (experiment C1), a simulation with changed SSTs only (experiment S1), a simulation with only a changed CO_2-concentration (experiment C2) and a final simulation with both changed SSTs and CO_2-concentration (experiment S2).

Figure 1 shows the zonal mean temperature changes for the different experiments with respect to the pre-industrial state, as modelled by WACCM. The results reach a statistical significance of 95% for the whole middle

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atmosphere domain in the experiments S2-C1 and C2-C1, and most of the middle atmosphere for experiment S1-C1. For this figure, as well as for all the results shown in this paper, we have used the 40 year mean of our data.

In line with what was shown in earlier studies (e.g. *Akmaev*, 2006; *Fomichev et al.*, 2007), 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. 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.

As found previously by *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 (experiment C2 and S1) is approximately equal to the changes observed in the combined simulation (experiment S2). *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 radiative and dynamic processes are not independent, these processes are seen to be approximately additive (*Sigmond et al.*, 2004, *Schmidt et al.*, 2006, *Fomichev et al.,* 2007).

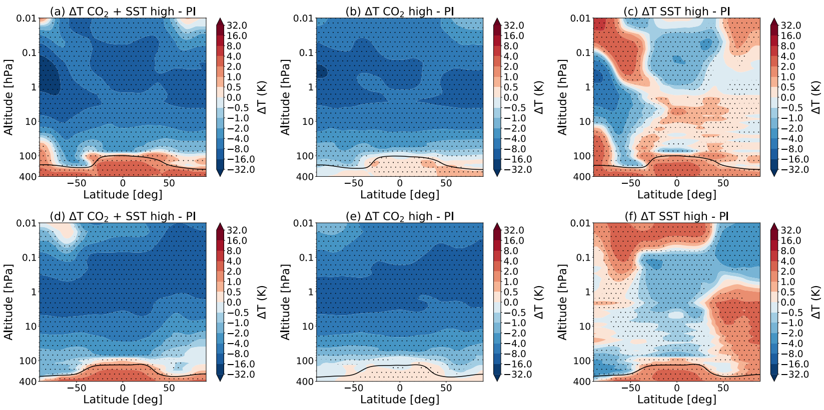

Figure 1: The total change in temperature in July (top) and January (bottom) for (a,d) the simulation with high CO_2 and SSTs, (b,e) the simulation with high CO_2, (c,f) the simulation with high SSTs, all as compared the pre-industrial control simulation. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The black line indicates the tropopause height for the experiments S2 (a,d), C2 (b,e) and S1 (c,f).

4. Meridional-vertical profiles of partial temperature changes

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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. The response to dynamical changes is calculated in a separate term.

This can be understood as we use the Fu-Liou radiative transfer model (*Fu and Liou*, 1992, 1993) to do offline calculations of the total local thermal equilibrium (LTE) radiative heating rate perturbation fields between the control experiment C1 and one of the other three experiments (i.e, C2, or S1, or S2). We use the standard outputs of atmospheric compositions (e.g., CO_2 and O_3) and thermodynamic fields (e.g., pressure, temperature, water vapour, clouds, surface albedo) as well as partial LTE radiative heating rate perturbation fields due to perturbations in individual atmospheric composition or thermodynamic fields (e.g., the terms on the right hand side of (A.9) except the first term).

We use the difference between the offline calculation of the total LTE radiative heating rate perturbations and the original total LTE radiative heating rate perturbations derived directly from the standard WACCM outputs as the error term of our offline LTE radiative heating perturbations. We note that the standard WACCM output fields also include non-LTE radiative heating fields, but do not include non-radiative heating rates. Therefore, we use the sum of the total LTE radiative heating rate perturbations and non-LTE radiative heating fields derived from the standard WACCM output fields to infer non-radiative heating rate perturbations under the equilibrium condition, namely Eq. (A.8).

We should also note that, 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.

In this section, we discuss the meridional-vertical profiles of the temperature responses to the direct forcing and the various feedbacks during July and January. In the next section, we will discuss regional and global means of partial temperature changes due to feedbacks.

4.1 Direct temperature response to CO_2

Figure 2 shows the zonal mean temperature change due to the increase in CO_2. We see that increasing CO_2 leads to a cooling almost everywhere in the middle atmosphere, except at the high latitudes in the cold summer upper mesosphere, where we see a warming instead. The higher the temperature, the more cooling due to the increasing CO_2-concentration (*Shepherd,* 2008). The reason for this is that the outgoing longwave radiation strongly depends on the Planck blackbody emission (*Zhu et al.,* 2016).

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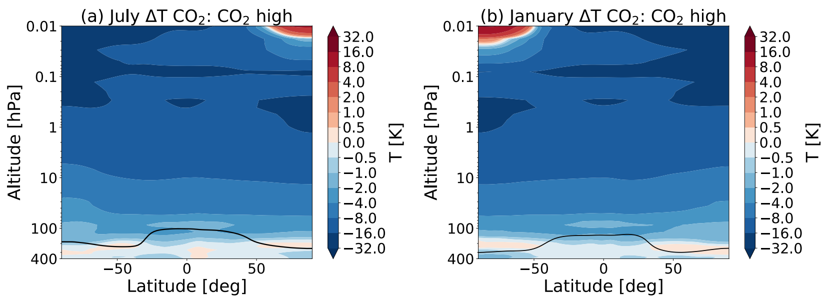

Figure 2: Partial temperature change due to the direct forcing of CO_2 for July (top) and January (bottom) due to the doubling of the atmospheric CO_2-concentration, as calculated by CFRAM, using experiment C2 and C1. The black line indicates the tropopause height for the C2 runs (with double CO_2-concentration).

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).

4.2 Ozone feedback

Ozone plays a crucial role in the chemical and radiative budget of the middle atmosphere. The distribution of ozone in the middle atmosphere is determined by both chemical and dynamical processes. Most of the ozone production takes place in the tropical stratosphere, as a result of photochemical processes, which involve oxygen. Meridional circulation then transports ozone to other parts of the middle atmosphere (*Langematz*, 2019). The production of ozone is largely balanced by catalytic destruction cycles involving NO_x, HO_x and Cl_x radicals. HO_x dominates ozone destruction in the mesosphere and lower stratosphere, while NO_x and Cl_x dominate this process in the middle and upper stratosphere (e.g. *Cariolle*, 1983).

Since the 1970s ozone in the middle atmosphere began to decline globally, due to increased production of ozone depleting substances (ODSs) (*Brühl and Crutzen*, 1988). The Montreal Protocol, adopted in 1987 to stop this threat, eventually led to a slow recovery of the stratospheric ozone over the recent two decades (*WMO*, 2018; *Langematz*, 2019). In our study, we don't consider the effect of anthropogenic ODSs since pre-industrial times (*Langematz*, 2019).

In this study, we are interested in the temperature response to changes in ozone concentration induced by the increased CO_2 concentration and/or the changes in SST in WACCM. Under enhanced CO_2 concentrations, the ratio between O_3 and O mixing ratios is generally shifted toward a higher concentration of ozone, which is caused by the strong temperature dependency of the ozone production reaction ($O + O_2 + M \rightarrow O_3 + M$).

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Fig. 3 shows the percentage changes in O_3-concentration when the CO_2-concentration and/or the SSTs change. The results reach a statistical significance of 95% for the whole middle atmosphere domain in the experiments S2-C1 and C2-C1, and most of the middle atmosphere for experiment S1-C1.

We find, as expected, that an increase in CO_2, leads to an increase of ozone in most of the middle atmosphere. The increase of O_3 is about 20% around 2 hPa in the tropical region for experiment S2 with respect to C1. This corresponds with what is seen by *Fomichev et al., (2007)*, however they find that the increase in ozone in January is a bit lower in this region (around 15%, see their Figure 7).

There are some regions where the O_3-concentration is decreasing. In the tropical lower stratosphere, a decrease of about 20% is seen, in the summer polar mesosphere (around 0.01 hPa) ozone decreases by 3%, while in the mesosphere (around 0.02 hPa), ozone decreases by over 30%. Fig. 3c and f show that changing the SSTs also has a significant impact on the ozone concentration. A complete account of the ozone changes is out of the scope of this paper.

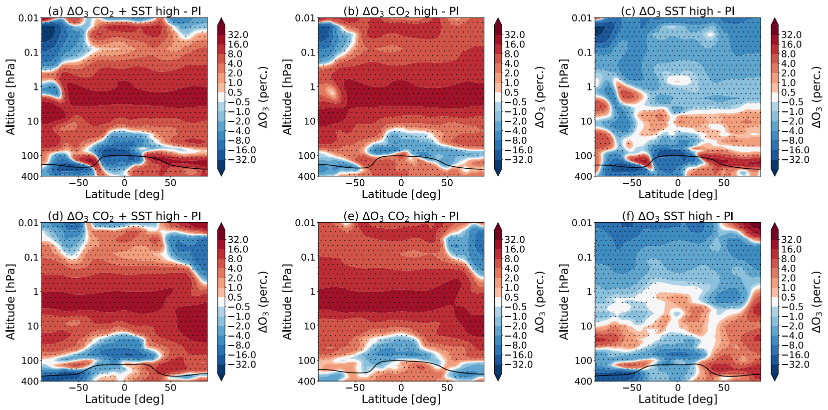

Figure 3: The percentage change in the zonal and monthly mean ozone concentration for July (top) and January (bottom) due to (a,d) combined effect of the CO_2 increase and SSTs changes (experiments S2 -C1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2 - C1) and the (c,f) SSTs (experiments S1 - C1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is indicated as in Fig. 1.

As we will discuss in the next section, an enhanced concentration of CO_2 also leads to changes in the dynamics in the middle atmosphere. The stratospheric Brewer-Dobson circulation is projected to strengthen, which would lead to an increase in the poleward transport of ozone. We will also see that an increase in CO_2-concentration leads to stronger summer pole-to-winter pole flow in the mesosphere.

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Figure 4 shows the percentage change in the zonal and monthly mean concentration of Cl, NO, O, OH, CH_3, NO_x and N_2O in July due to combined effect of the CO_2 increase and SSTs changes (experiment S2 vs C1), as simulated by WACCM. The patterns in January look similar (not shown). These results reach a statistical significance of 95% for the whole middle atmosphere domain.

We would like to point out that the changes in these constituents are only brought about by the CO_2-concentration and/or the SSTs. We still use the F_1850 compset and the only difference between the runs is the forcing in CO_2 and SSTs. The changes in chemical constituents look very similar to those found by *Schmidt et al.* (2006) who performed a similar experiment as discussed in section 2.2, see their Figure 20. Note that Fig. 4 shows the changes due to both the CO_2 increase and SSTs changes, while their Figure 20 shows the percentage changes due the changes in CO_2-concentration only and also only above 1 hPa.

As in *Schmidt et al.* (2006), we see an decrease in atomic oxygen (O) mixing ratio at high summer latitudes around 0.01 hPa (see Fig. 4c), which results from increased upwelling. This increase in O leads to a decrease in ozone in this region. We also see decrease of ozone concentration in the winter polar region around 0.1 hPa (approximately 65 km). This could be caused an increase of NO and for a small part by Cl mixing ratios, which result from a stronger subsidence of NO and Cl rich air, as suggested in Schmidt et al. *2006*. As stated before, complete discussion of the changes in ozone concentration is out of the scope of this paper and the changes in other constituents shown in Figure 4 are shown for reference only.

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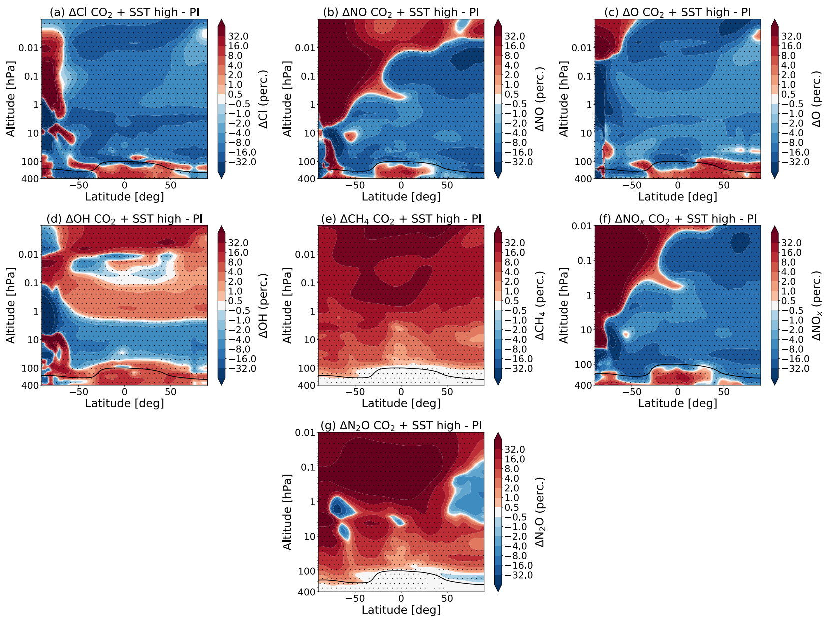

Figure 4: The percentage change in the zonal and monthly mean concentration of Cl (a), NO (b), O (c), OH (d), CH_4 (e) and NO_x (f) and N_2O (g) in July due to combined effect of the CO_2 increase and SSTs changes (experiment (S2 vs C1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is indicated as in Fig. 1.

What is new in this study, is that we can calculate the temperature responses due to the changes in ozone concentration. These temperature responses are shown in Figure 5. 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 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)*.

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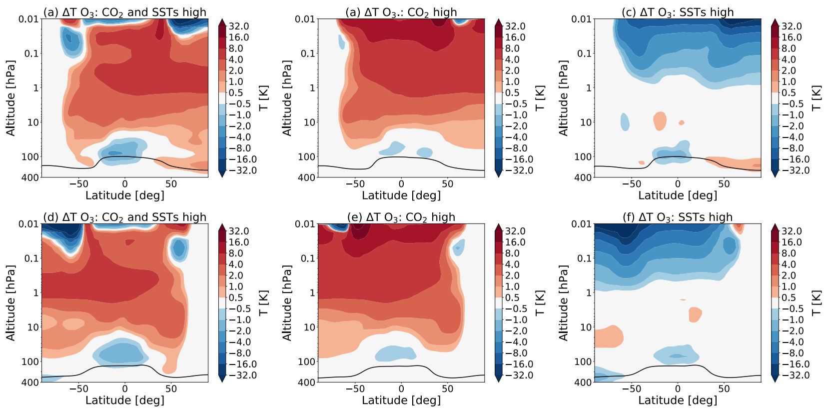

Figure 5: Partial temperature responses to changes in O_3-concentration, as calculated by CFRAM, in July (top) and January (bottom) due to the (a,d) combined effect of the CO_2 increase and SSTs changes (S1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2) and the (c,f) SSTs (experiments S1). The tropopause height is indicated as in Fig. 1.

4.2 Dynamical feedback

The zonal mean residual circulation forms an important component of the mass transport by the Brewer-Dobson circulation (BDC). It consists of a meridional ($\bar{v}^*$) and a vertical ($\bar{w}^*$) 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.

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 scale 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

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stratosphere, which leads to a strengthening 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 differences in the meridional component of the residual circulation ($\bar{v}^*$) between the different simulations are shown in Fig. 6. These data are averaged over the 40 years of data. The results reach a statistical significance of 95% almost the whole area above 1 hPa for the experiments C2-C1, for the experiment S1-C1 the results reach a statistical significance of 95% in most of the area below this level. The experiments S2-C1 show the largest region of statistical significance, apart from some regions below 1 hPa.

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 affect the dynamics in the middle atmosphere. It has for example been shown that higher SSTs in the tropics leads to an amplification in deep convection, which enhances the generation of quasi-stationary waves (*Deckert and Dameris, 2008*). Enhanced SSTs lead to an enhanced dissipation of planetary waves, as well as an enhanced dissipation of orographic and non-orographic waves in the upper stratosphere (*Oberländer et al., 2013*).

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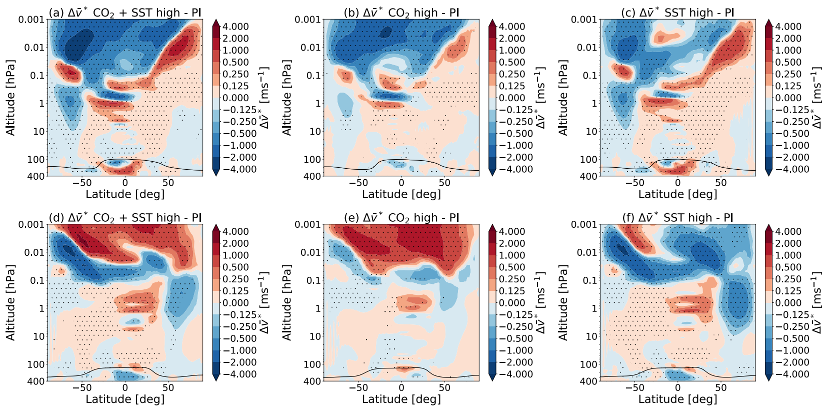

Figure 6: Changes in the zonal and monthly mean transformed Eulerian-mean residual circulation horizontal velocity $\bar{v}^*$ for July (top) and January (bottom) due to (a,d) combined effect of the CO_2 increase and SSTs changes (experiments S2 -C1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2 - C1) and the (c,f) SSTs (experiments S1 -

C1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is as indicated in Fig. 1.

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.

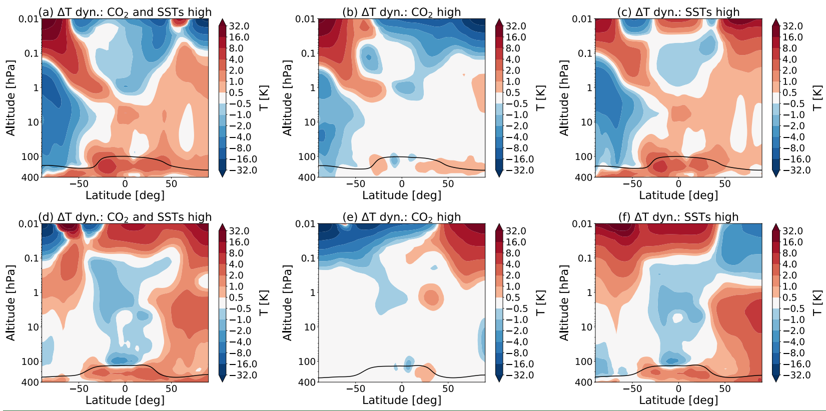

Figure 7: Partial temperature responses to changes in dynamics, as calculated by CFRAM, in July (top) and January (bottom) due to the (a,d) combined effect of the CO_2 increase and SSTs changes (S1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2) and the (c,f) SSTs (experiments S1). The tropopause height is indicated as in Fig. 1.

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.

4.4 Water vapour feedback

Figure 8 shows how the water vapour is changing in the middle atmosphere if the CO_2-concentration is increased and/or the SSTs are changed with respect the pre-industrial control run. In WACCM, increasing the CO_2-concentration alone leads to a decrease of water vapour in most of the middle atmosphere (Fig. 8b and f). The results reach a statistical significance of 95% for the

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Water vapour plays a secondary but not negligible role in determining the middle atmosphere climate sensitivity. In Figure 3, we saw the

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whole middle atmosphere domain in the experiments S2-C1 and S1-C1, and most of the middle atmosphere for experiment C2-C1, apart from the winter hemisphere region around 0.1 hPa.

The amount of water vapour in the stratosphere is determined by transport through the tropopause as well as by the oxidation of methane in the stratosphere itself. The transport of the water vapour in the stratosphere is mainly a function of the tropopause temperature (*Solomon et al.,* 2010). In WACCM, we see a decrease in temperature in the tropical tropopause for the double CO_2 experiment 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 using the SSTs from the doubled CO_2-climate leads to an increase in water vapour almost everywhere in the middle atmosphere as compared to PI (Fig. 8c and f). In WACCM, forcing with SSTs from a double CO_2-climate is observed to lead to a higher and warmer tropopause, which can explain this increase of water vapour. However, it should be noted that models currently have a limited representation of the processes determining the distribution and variability of lower stratospheric water vapour. Minimum tropopause temperatures are not consistently reproduced by climate models (*Solomon et al.,* 2010; *Riese et al.,* 2012). At the same time, observations are not completely clear about whether there is a persistent positive correlation between the SST and the stratospheric water vapour (*Solomon et al.*, 2010).

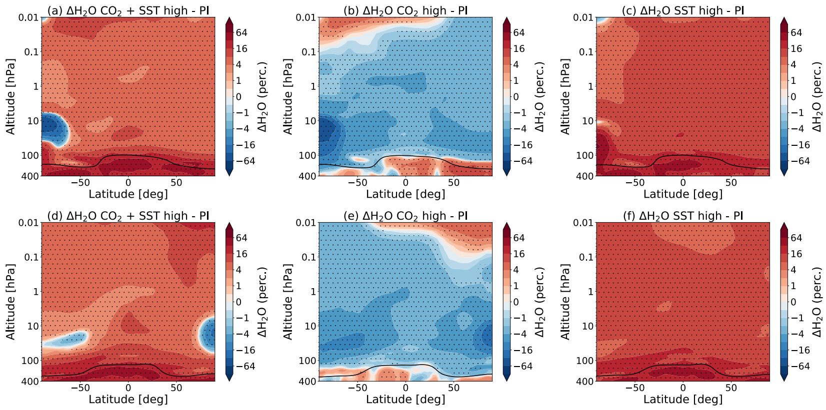

Figure 8: The percentage changes in the zonal and monthly mean water vapour mixing ratio for July (top) and January (bottom) due to (a,d) combined effect of the CO_2 increase and SSTs changes (experiments S2 -C1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2 - C1) and the (c,f) SSTs (experiments S1 - C1), as simulated by WACCM. The dotted regions indicate the regions where the data reaches a confidence level of 95%. The tropopause height is as indicated in Fig. 1.

Figure 9 shows the temperature responses due to the changes in water vapour as calculated by CFRAM. It can be seen that the regions where there

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is an increase in the water vapour, there is a cooling, and vice versa. This can be understood as increasing the water vapour in the middle atmosphere leads to an increase in longwave emissions in the mid and far-infrared by water vapour. This in turns leads to a cooling of the region. Similarly, a decrease in water vapour leads to a warming of the region (*Brasseur and Solomon,* 2005). Higher up in the atmosphere, there are large percentage changes in water vapour. However, the absolute concentration of water is small there, which explains why there is no temperature response to these changes.

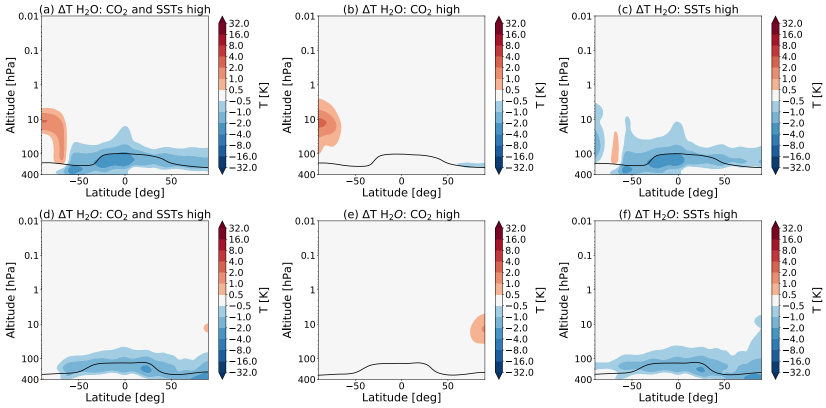

Figure 9: Partial temperature responses to changes in water vapour, as calculated by CFRAM, in July (top) and January (bottom) due to the (a,d) combined effect of the CO_2 increase and SSTs changes (S1), (b,e) the doubling of the atmospheric CO_2-concentration (experiments C2) and the (c,f) SSTs (experiments S1). The tropopause height is indicated as in Fig. 1.

Water vapour plays a secondary but not negligible role in determining the middle atmosphere climate sensitivity. In the lower stratosphere, H_2O contributes considerably to the cooling in this region. Above 30 hPa, the water vapour contribution to the energy budget is negligible, as also seen by *Fomichev et al.* (2007).

4.5 Cloud and albedo feedback

Forcing the model with SSTs from the double CO_2-climate (as in experiment S1 and S2) yields 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 (as in experiment C2). Figure 10 shows the temperature responses to changes in cloud (left) and albedo (right) in July (top) and January (bottom) for experiment S1, as calculated by CFRAM.

Fig. 10 shows 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 10a). In January, the pattern looks slightly different (see Figure 10c). These temperature changes are due to changes in the balance between the

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Changes in SSTs yield
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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 10b and d). The cooling in the summer polar stratosphere shown in Figure 10b and d is due to radiative changes. We suggest that this cooling is due to a decrease in surface albedo, which would lead to less shortwave radiation being reflected. However, more research is needed.

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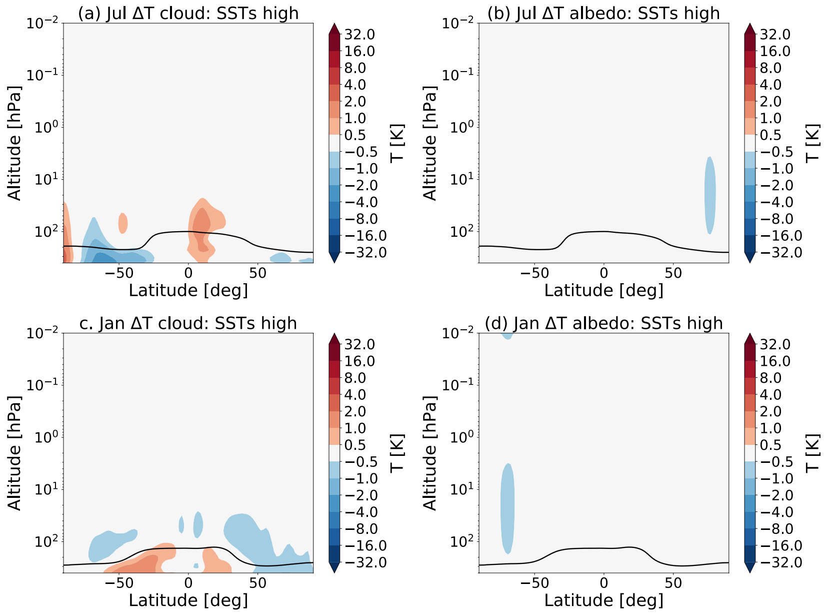

Figure 10: Partial temperature responses to changes in clouds (left) and albedo (right), as calculated by CFRAM, in July (top) and January (bottom) due to the SSTs (experiment S1). The tropopause height is indicated as in Fig. 1.

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Cloud and albedo 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). However, it is clear that these feedbacks play only a very small role in the middle atmosphere temperature response to the doubling of CO_2 and SSTs.

5. Regional and global means of partial temperature changes due to feedbacks

To study the relative importance of the different feedback processes globally we show the average change in global mean temperature for the lower stratosphere, the upper stratosphere and the mesosphere for the S2 experiment with the changed CO_2-concentration and changed SSTs in Figure 11. We also show the average change in temperature in the polar regions

(90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) for the lower and upper stratosphere and the mesosphere.

In order to calculate the lower stratosphere temperature changes, we take the average value of the temperature change from the tropopause until 24 hPa. The pressure level of the tropopause is simulated in WACCM for each latitude and longitude, we use this pressure level to demarcate between the troposphere and stratosphere. We consider 24 hPa as a crude estimate for the boundary between the lower and upper stratosphere.

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The tropopause is not exactly at the same pressure level in the perturbation experiments as compared to the pre-industrial control run (C1). We always take the tropopause of the perturbation experiment which is a bit higher at some latitudes, to make sure that we do not use values from the troposphere. We add the values for each latitude up and take the average. This average is not mass weighted. By taking the in this way, we can directly compare the vertical values in different regions of the atmosphere. The temperature changes in the upper stratosphere and in the mesosphere are calculated in the same way, but then for the altitudes 24 hPa-1 hPa and 1 hPa-0.01 hPa respectively.

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Figure 11 shows the radiative feedbacks due to ozone, water vapour, clouds, albedo and the dynamical feedback, as well as the small contribution due to the Non-LTE processes, as calculated by CFRAM. The 'total'-column shows the temperature changes in WACCM, while the column 'error' shows the difference between temperature change in WACCM and the sum of the calculated temperature responses in CFRAM. Note that the range of values on the y-axis is not the same for the different subplots.

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Figure 11 shows that the temperature change in the lower stratosphere due to the direct forcing of CO_2 is around 3 K in the global mean. There is a stronger cooling in the tropical region of about 4 K in July and 3.5 K in January. We also observe that there is a cooling of about 1 K due to ozone feedback in the tropical region while there is a slight warming taking place in the summer hemispheres in both January and July. We also see that the temperature change in the lower stratosphere is influenced by the water vapour feedback. There is a cooling of about 0.5 K in the lower stratosphere, apart from in the southern polar area. There is some small influence from the cloud and albedo feedback, which can be negative or positive (see also Fig. 9).

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In the upper stratosphere, the cooling due to the direct forcing of CO_2 is with about 9 K in the global mean considerably stronger than in the lower stratosphere. The cooling is stronger in the summer polar regions, where the cooling due to the direct forcing of CO_2 reaches 11K. In the winter polar region, this cooling is only about 8K.

The water vapour, cloud and albedo feedback play no role in the upper stratosphere nor in the mesosphere. The ozone feedback results in the positive partial temperature changes in the upper stratosphere, of about 2 K in the global mean. The changes in ozone don’t result in temperature changes in the winter hemisphere, as discussed in section 4.2.

The picture in the mesosphere is similar as in the upper stratosphere. The main difference is that the temperature changes are larger. The global temperature change due to direct forcing of CO_2 is about 15 K. The O_3-feedback results in a partial temperature changes of about 3 K in the mesosphere in the global mean. The temperature change due to ozone in the equatorial mesosphere is about 4 K, while the warming due to ozone in the summer polar region a bit smaller: around 3K. Just like in the upper stratosphere water vapour, cloud and albedo feedback play no role.

We see, that the ozone feedback generally yields a radiative feedback that mitigates the cooling, which is due to the direct forcing of CO_2. This has been suggested in earlier studies, such as *Jonsson et al.*, *2004, Dietmüller et al., 2014*. With CFRAM, it is possible to quantify this effect and to compare it with the effects of other feedbacks in the middle atmosphere. Note that no other method before has been able to quantify how much of the temperature change in the middle atmosphere is due to the different feedback processes before.

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The temperature response due to dynamical feedbacks is small in global average: less than 1 K. This can be understood as waves generally do not generate momentum and heat, but redistribute these instead (*Zhu et al.*, 2016). However, the local responses to dynamical changes in the high latitudes are large, as we have seen in section 4.2. There are some very small temperature responses due to non-LTE effects as well, which contribute to the temperature change in the mesosphere mostly.

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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 transfer model used in the offline CFRAM calculations is different than the radiative transfer model used in the climate simulations with WACCM.

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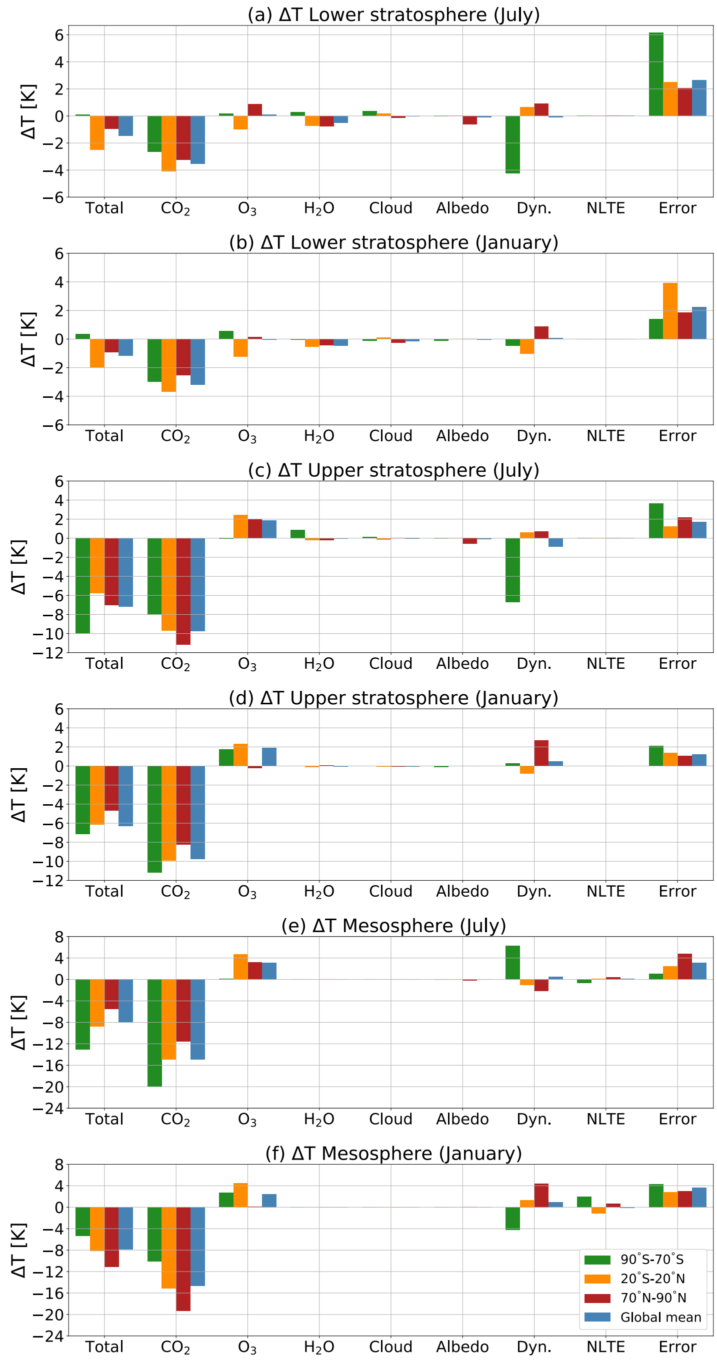
(a) ΔT Lower stratosphere (July)
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(f) ΔT Mesosphere (January)
ΔT [K]
Total
CO_2
O_3
H_2O
Cloud
Albedo
Dyn.
NLTE
Error
90˚S-70˚S
20˚S-20˚N
70˚N-90˚N
Global mean

Figure 11: The mean temperature responses to the changes in CO_2 and various feedback processes in the lower stratosphere from the tropopause height until 24 hPa (a,b), upper stratosphere from 24-1 hPa (c, d) and in the mesosphere from 1-0.01 hPa (e,f) in July (a, c, e) and January (b, d, f) in the polar regions (90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) and the global mean, for S1 experiment (double CO_2 and changed SSTs). Note that the range of values on the y-axis is not the same for the different subplots.

In addition, the vertical profiles of the temperature responses to the direct forcing of CO_2 and the feedbacks are shown in Figure 12. 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. The changes in ozone concentration in response to the doubling of CO_2 lead 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.

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There is also a relatively large temperature response to the changes in dynamics. In Fig. 12, 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 11. The Non-LTE effects are also small, but start to play a small role above 0.1 hPa, the exact mechanism of which is outside the scope of this paper.

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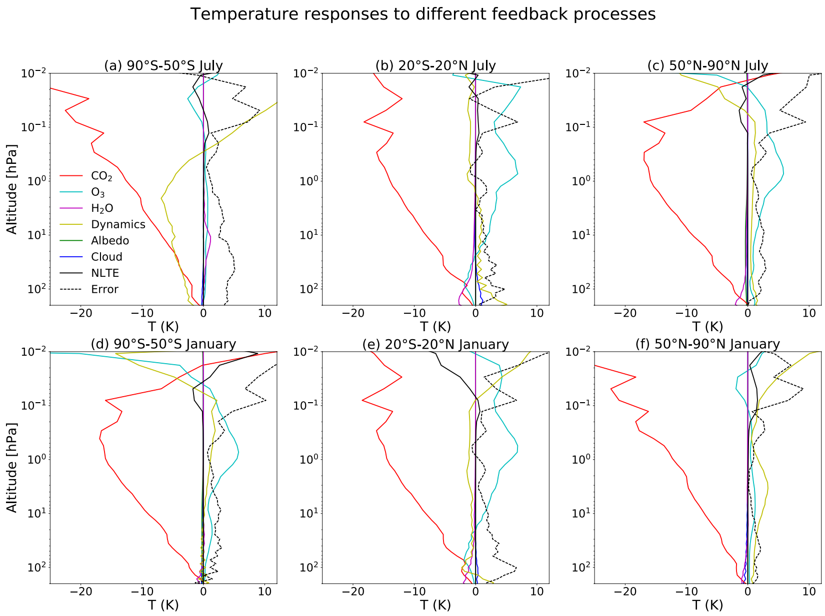

Figure 12: Vertical profiles of the temperature responses to the changes in CO_2 and various feedback processes in July (top) and January (bottom) for

due to double CO_2 and changed SSTs in the atmosphere between 200 and 0.01 hPa, for regions from 50° N/S poleward and the tropics (20°S-20°N), as calculated by CFRAM.

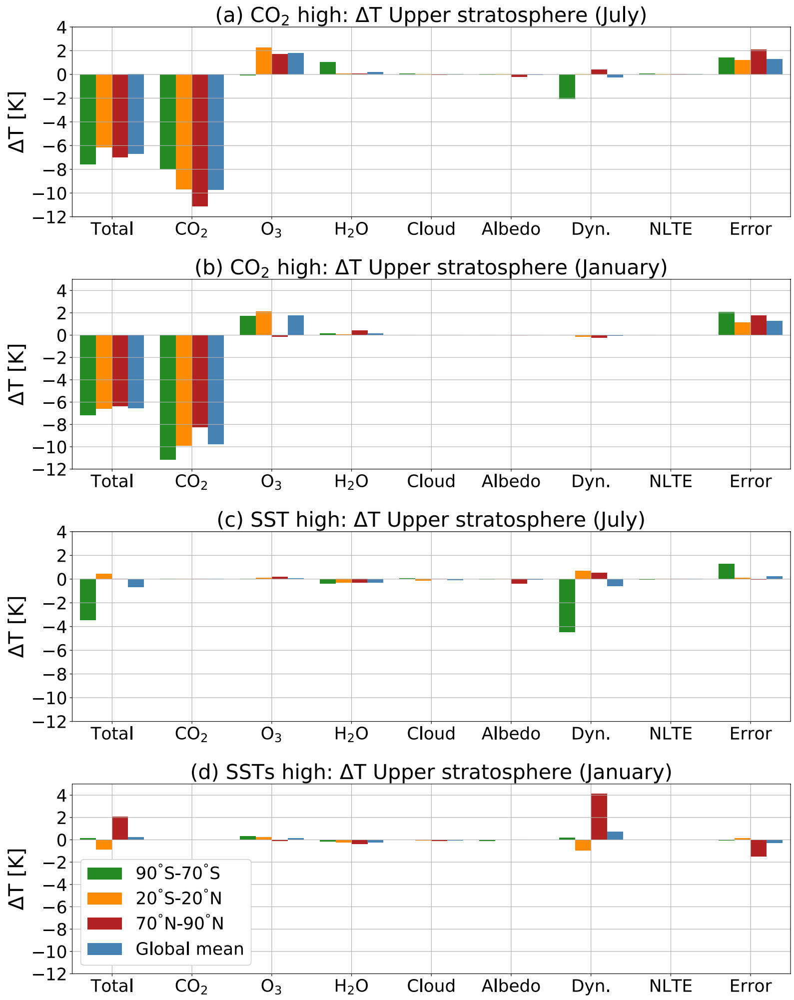

Figure 13: The mean temperature responses to the changes in CO_2 and various feedback processes in July (a,c) and January (b,d) in the upper stratosphere between 24 and 1 hPa, for polar regions (90°S-70°S and 70°N-90°N), the tropics (20°S-20°N) and the global mean for the experiment with double CO_2 (C2) (a,b) and changed SSTs (S1) (c,d) separately.

Figure 13 shows temperature responses in the upper stratosphere for the experiment with double CO_2 (a,b) and changed SSTs (c,d) separately. This has been done to give insight in the temperature response of the CO_2 and the SST separately. These temperature changes were calculated in the same way as for Fig. 11. Again also, the 'total'-column shows the temperature changes as simulated 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.

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We learn from this figure that the effects of the changed SSTs on the upper stratosphere are relatively small as compared to the effects of changing the CO_2. We show the temperature changes for the upper stratosphere as an example. In the lower stratosphere and the mesosphere, we see the same pattern: the effect of the CO_2 on the temperature is generally much larger than the effect of the SSTs on the temperature. This finding is consistent with the study of *Fomichev et al.* (2007), where it is concluded that the impact of changes in SSTs on the middle atmosphere is relatively small and localized as compared to the combined response of changing the CO_2-concentration and the SSTs.

The changes in SSTs are, however, responsible for large temperature changes as a result of the dynamical feedbacks, especially in the winter hemispheres, where there is a temperature response of 4K. A similar figure for the lower stratosphere (not shown) shows that the temperature response to the water vapour feedback is almost solely due to changes in the SSTs and not the direct forcing of CO_2.

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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.

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5. Discussion and conclusions

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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 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 is important to note that no other method before has been able to quantify how much of the temperature change in the middle atmosphere is due to the different feedback processes before.

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It was found 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.

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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 have given an overview of the mean temperature responses to the changes in CO_2 and various feedback processes in the lower stratosphere, upper stratosphere and in the mesosphere in January and July. We find that the temperature change due to the direct forcing of CO_2 increases with increasing height in the middle atmosphere. The temperature change in the lower stratosphere due to the direct forcing of CO_2 is around 3 K. There is a stronger cooling in the tropical lower stratosphere of about 4 K in July and 3.5 K in January.

In the upper stratosphere, the cooling due to the direct forcing of CO_2 is about 9 K, which is considerably stronger than in the lower stratosphere. The cooling is stronger in the summer polar regions, where the cooling reaches a value of 11K, than in the winter polar region, where the cooling is only about 8K. In the mesosphere, the cooling due to the direct forcing of CO_2 is even stronger: 15 K.

The ozone concentration changes due to changes in the CO_2-concentration as well as by changes in the SSTs. The temperature changes caused by this change in ozone concentration generally mitigate the cooling caused by the direct forcing of CO_2. However, in the tropical lower stratosphere and in some regions of the mesosphere, the ozone feedback cools these regions further. In the tropical lower stratosphere, for example, there is a cooling of 1K due to the ozone feedback.

We also have seen that the global mean temperature response due to dynamical feedbacks is small in the global average in all regions: less than 1 K. However, local responses to the changes in dynamics can be large. Doubling the CO_2 leads to a stronger summer-to-winter-pole flow, which leads to a cooling of the summer mesosphere and a warming of the winter mesosphere. Changing the SSTs weakens this effect in the mesosphere, but affects the temperature response in the stratosphere and lower mesosphere.

Using CFRAM on WACCM data shows that the change in water vapour leads to a cooling of up to 2 K in the lower stratosphere. It should be noted that climate models currently have a limited representation of the processes determining the distribution and variability of lower stratospheric water vapour. This means that the temperature response to the water vapour feedback might be different using a different model. We have also seen a small effect of the cloud and albedo feedback on the temperature response in the lower stratosphere, while these feedbacks play no role in the upper stratosphere and the mesosphere.

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Ozone responds to changes in respond to changes in CO_2 and/or SSTs due to changes in chemical reaction rate constants and due to the strength of the up- and downwelling. The temperature changes caused by these changes…

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The results seen in this study are consistent with earlier findings. As in *Shepherd et al.*, (2008), we find that the higher the temperature at a region in the atmosphere, the more cooling there is seen due to the direct feedback of CO_2. We find, as in *Zhu et al.*, (2016) that 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 changes in ozone concentration instead.

We have also seen that the ozone feedback generally yields a radiative feedback that mitigates the cooling, which is due to the direct forcing of CO_2, which is consistent with earlier studies such as *Jonsson et al., (2004), Dietmüller et al., (2014)*. CFRAM is the first study that allows for calculating how much of the temperature response is due to which feedback process.

The next step would be 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. A better understanding of the effect of the increased CO_2-concentration on the middle atmosphere, will help to distinguish the effects of the changes CO_2- and O_3-concentration.

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 on the global mean temperature is currently not clear (*Nowack et al.*, 2015, *Marsh et al.*, 2016). A similar analysis as in this paper can be performed 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 there is a relatively large error due to the linearization in the model. 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.

Appendix: Formulation of CFRAM diagnostics using outputs of WACCM

The mathematical formulation of CFRAM is based on the conservation of total energy (*Lu and Cai*, 2009). At a given location in the atmosphere, the energy balance in an atmosphere-surface column can be written as:

$$\vec{R} = \vec{S} + \vec{Q}^{conv} + \vec{Q}^{turb} - \vec{D}^{v} - \vec{D}^{h} + \vec{W}^{fric} \qquad \text{(A1)}$$

$\vec{R}$ represents the vertical profile of the net long-wave radiation emitted by each layer in the atmosphere and by the surface. $\vec{S}$ is the vertical profile of the solar

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radiation absorbed by each layer. Q^{turb} is the convergence of total energy fluxes in each layer due to turbulent motions, Q^{conv} is convergence of total energy fluxes into the layers due to convective motion. D^{v} is the large-scale vertical transport of energy from different layers to others. D^{h} is the large-scale horizontal transport within the layers and W^{fric} is the work done by atmospheric friction. All terms in (A1) have units of Wm^{-2}.

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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:

$$\Delta R = \Delta F^{ext} + \Delta S + \Delta Q^{conv} + \Delta Q^{turb} - \Delta D^{v} - \Delta D^{h} + \Delta W^{fric} \qquad \text{(A2)}$$

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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 balance the non-temperature induced energy flux perturbations.

Equation (A2) can also be written as:

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$$\Delta(S - R)_{total} + \Delta dyn = 0 \qquad \text{(A3)}$$

The term $\Delta(S - R)$ can be calculated as the longwave heating rate and the solar heating rate are output variables of the model simulations. We take the time mean of the WACCM data and perform the calculations for each grid point of the WACCM data. This means that in the end, we will have the temperature changes at each latitude, longitude and height.

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We then calculate the difference in these heating rates for the perturbation simulation and the control simulation.

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

$$\Delta dyn = -\Delta(S - R)_{total} \qquad \text{(A4)}$$

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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.

$$\Delta(S - R) = \Delta(S - R)_{(WACCM)} * mass_k * c_p \qquad \text{(A5)}$$

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In which $\Delta(\mathcal{S}-\mathcal{R})$ is the difference in the shortwave radiation ($\mathcal{S}$) and longwave radiation ($\mathcal{R}$) between the perturbation run and the control run as a flux in Wm^{-2}, while $\Delta(\mathcal{S}-\mathcal{R})_{(WACCM)}$ is this difference as heating rate in Ks^{-1} in WACCM, with $mass_k = \frac{p_{k+1}-p_k}{g}$ with p in Pa, $c_p = 1004 \ \ J\,kg^{-1}\,K^{-1}$ the specific heat capacity at constant pressure and *g* the gravitational acceleration 9.81 ms^{-2}.

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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 $\Delta(\mathcal{S}-\mathcal{R})_{total}$ in an LTE and a non-LTE term:

$$\Delta(\mathcal{S}-\mathcal{R})_{total} = \Delta(\mathcal{S}-\mathcal{R})_{LTE} + \Delta(\mathcal{S}-\mathcal{R})_{non-LTE} \qquad \text{(A6)}$$

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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 $\Delta(\mathcal{S}-\mathcal{R})_{non-LTE}$ as well, where we again need to convert our result from Ks^{-1} to Wm^{-2}:

$$\Delta(\mathcal{S}-\mathcal{R})_{non-LTE} = \Delta(\mathcal{S}-\mathcal{R})_{non-LTE(WACCM)} \ \ mass_k * c_p \qquad \text{(A7)}$$

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This term can be inserted in equation (3):

$$\Delta(\mathcal{S}-\mathcal{R})_{LTE} + \Delta(\mathcal{S}-\mathcal{R})_{non-LTE} + \Delta dyn = 0 \qquad \text{(A8)}$$

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

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

$$\Delta\mathcal{R}_{LTE} = \frac{\partial\mathcal{R}}{\partial T}\Delta T + \Delta\mathcal{R}_{CO_2} + \Delta\mathcal{R}_{O_3} + \Delta\mathcal{R}_{H_2O} + \Delta\mathcal{R}_{albedo} + \Delta\mathcal{R}_{cloud} \qquad \text{(A9)}$$

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

For equation (A9), 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).

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The term $\frac{\partial\mathcal{R}}{\partial T}\Delta T$ represents the changes in the IR radiative fluxes related to the temperature changes in the entire atmosphere-surface column. The matrix $\frac{\partial\mathcal{R}}{\partial T}$

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:

$$\Delta S_{LTE} = \Delta S_{CO_2} + \Delta S_{O_3} + \Delta S_{H_2O} + \Delta S_{albedo} + \Delta S_{cloud} \qquad \text{(A10)}$$

Similarly, to equation (A9), we perform a linearization.

Substituting (A9) and (A10) in equation (A8) yields:

$$\Delta(S-R)_{CO_2} + \Delta(S-R)_{O_3} + \Delta(S-R)_{H_2O} + \Delta(S-R)_{albedo} + \Delta(S-R)_{cloud} - \frac{\partial R}{\partial T}\Delta T + \Delta(S-R)_{non-LTE} + \Delta dyn = 0 \qquad \text{(A11)}$$

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This can be written as:

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$$\Delta T = \left(\frac{\partial R}{\partial T}\right)^{-1} \{\Delta(S-R)_{CO_2} + \Delta(S-R)_{O_3} + \Delta(S-R)_{H_2O} + \Delta(S-R)_{albedo} + \Delta(S-R)_{cloud} + \Delta(S-R)_{non-LTE} + \Delta dyn\} \qquad \text{(A12)}$$

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Competing interests

The authors have no competing interests to declare.

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