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Recovery of the Tropical Lower Stratospheric Ozone Layer

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Abstract.

Long-term time dependent simulations out to the year 2050 using the Goddard Space Flight Center interactive 2D chemistry-radiation-dynamics model are used to predict tropical lower stratospheric ozone. We find that despite a return to background chlorine levels, the tropical lower stratospheric ozone layer will not recover because of increasing carbon dioxide. With background aerosol amounts, tropical ozone at 50 hPa is reduced by ~2% between 1980 and 1997 and remains at that level of depletion out to 2050. The simulations also suggest that tropical tropopause temperatures will be lowered by a maximum of 0.1 K between 1980 and 2015 and that tropical lower stratospheric water vapor will be reduced by a maximum of 2% between 1980 and 2018. These results may have important implications for recovery detection strategies.

1. Introduction

The future evolution of column ozone computed by eight zonally averaged stratospheric chemistry models was compared in *WMO* [2003]. The two models which included temperature feedbacks from increasing greenhouse gases, the NOCAR [*Portmann et al.,* 1999] and the GSFC-INT [*Rosenfield et al.*, 2002] models, predicted more rapid rates of column ozone increases. The model calculations presented in *Rosenfield et al.* [2002] predicted that annually, globally averaged column ozone would recover to 1980 values in ~2035. These simulations showed that increasing carbon dioxide would speed up tropical ozone recovery above ~30 hPa due to colder temperatures and slower photochemical loss. Below 30 hPa, ozone recovery was delayed. This delay was attributed to decreased photochemical production and increased tropical upwelling. In this paper we study further the tropical lower stratospheric ozone depletion and its predicted non-recovery. The initial tropical ozone depletion in the 1980-1997 period is shown to be the result of ozone losses in the middle and high latitudes due to heterogeneous chemistry, while the non-recovery after 1997 is due solely to decreased production as a result of increasing carbon dioxide.

2. Method

The chemical model used here is the GSFC chemistry-radiation-dynamics interactive 2-D (latitude-pressure) model, which has been described by *Rosenfield* [2003] and *Rosenfield et al.* [2002]. The model includes sulfate aerosols, polar stratospheric clouds, and tropical subvisible cirrus clouds. Gas phase reaction rates and photolysis cross sections are from *Sander et al.* [2003].

Three time dependent runs between 1970 and 2050 were carried out, with the period between 1970 and 1980 allowing for model spinup. In each of the runs the time varying surface boundary conditions of methane, nitrous oxide, and chlorine and bromine species were given by Scenario A3 of *WMO* [1999, p.12.7]. Run 1 included background sulfate aerosols in the chemistry and radiation, and allowed for carbon dioxide to increase according to this same WMO scenario. Run 2 was the same as run 1 except that carbon dioxide was fixed at 1980 levels. Run 3 was the same as run 1 except that the sulfate aerosol arising from the Mt. Pinatubo eruption of 1991 was also included [see *Rosenfield*, 2003].

3. Results and Discussion

Figure 1 shows the computed change in annual mean ozone mixing ratios from 1980 as a function of time at 5^{o}N and 51 hPa. As chlorine increases between 1980 and 1997, each of the three runs show ozone depletion. There is a large 6% depletion in 1992 due to the volcanic aerosol, while the background aerosol runs show depletions of ~1% and ~2% for the fixed and increasing CO_2 cases, respectively. After the year 1997, the ozone in the scenarios with increasing CO_2 stay depleted, while in the case where CO_2 does not increase, ozone shows a recovery to 1980 values in ~2025. There are similar patterns at the 38 hPa model level. The nonrecovery computed in runs 1 and 3 occurs between the latitudes 15^{o}S and 25^{o}N.

The tropical column ozone is predicted to recover to 1980 values in ~2021 for the increasing CO_2 cases and ~2042 for the fixed CO_2 case (not shown). Computed changes in surface ultraviolet radiative fluxes in the tropics were negligible.

3.1 Changes from 1980 - 1997

To explain these results we focus first on years 1980 to 1997, when stratospheric chlorine is increasing. In the model, the ozone change is determined by chemical production, chemical loss, transport, and diffusion. In the tropical lower stratosphere the ozone production and transport terms are roughly an order of magnitude greater than the loss and diffusion terms. Thus the computed ozone amounts are largely determined by the balance between the chemical production and the upward tropical transport.

Figure 2 shows the time varying annual mean production and negative of the transport at 5^{o}N and 51 hPa for the background aerosol runs with increasing and fixed CO_2. Both the production and the upwelling increase from 1980 to 1997. The ozone losses in the upper stratosphere during this time period allow more ultraviolet radiation to reach the lower stratosphere, resulting in greater chemical production. This is an example of the well known “self-healing” effect [*Hudson,* 1977, p. 201]. The rise in production at this level is less in the increasing CO_2 case because there is less upper stratospheric ozone loss due to slower photochemistry in a relatively colder stratosphere [*Rosenfield et al.*, 2002].

The increase in tropical upwelling from 1980 to 1997 is a result of reduced ozone at middle and high latitudes due primarily to heterogeneous chemistry on aerosol surfaces. Between 45^{o}N and 65^{o}N computed O_3 losses are 3-6%, while between 45^{o}S and 65^{o}S the losses are 6-20%. These ozone losses and the subsequent decrease in solar heating lead to increased net radiative cooling. This, in turn, leads to increased downwelling in the middle and high latitudes. The global mass conservation requires a ~3% increase in tropical upwelling.

In the case of increasing CO_2 we compute a ~2% ozone depletion and upward transport always exceeds production. With fixed carbon dioxide the upward transport is greater than the ozone production from 1980 to 1990. Lower stratospheric ozone production increases more rapidly in the fixed CO_2 case due to the greater upper stratospheric ozone loss. Thus, as shown in Figure 1, the ozone does not decline as rapidly after 1990.

The mechanism described above for increasing the tropical upwelling occurs at the tropopause as well as in the lower stratosphere. This increase in tropical tropopause upwelling leads to increased adiabatic cooling and an accompaning temperature decrease in the years 1980-1997. The computed annual mean temperature changes at the model tropical tropopause are shown in Figure 3. There are temperature decreases of 0.1K and 0.25K for the increasing and fixed CO_2 cases, respectively. Increasing CO_2 tends to warm the tropopause radiatively, partially reducing the adiabatic cooling. This explains the difference between the fixed and increasing CO_2 cases.

In the volcanic run the aerosol radiative heating causes a 1.4 K temperature increase in 1992 which then reverts to the temperature decrease computed for the background aerosol case as volcanic aerosols are transported out of the stratosphere. In the fixed CO_2 case the temperature asymptotes to 1980 values out to 2050, while in the increasing CO_2 case the temperature remains below 1980 values until ~2015.

These changes in tropopause temperature impact stratospheric water vapor. Figure 4 shows annual mean water vapor reductions at 51 hPa for the fixed and increasing CO_2 background aerosol cases of 4% and 2%, respectively. The reductions in water vapor are another factor contributing to increased upwelling, as less radiative cooling due to water vapor would tend to increase the net radiative heating and result in increased upwelling. The lower stratospheric net heating increases by ~3% between 1980 and 1997. In the volcanic case water vapor increases by 6% in 1992, due to the temperature increase at the tropopause, and then falls back to a 2% depletion in 1996. In the changing CO_2 cases, aside from the temporary volcano induced increase, water vapor values remain below 1980 values until ~2018.

We also carried out steady state runs for 1980 and 2000 boundary conditions, with CO_2 fixed at the 1995 value. These runs were compared with fixed boundary condition runs of a 3D general circulation model with coupled ozone chemistry [*Gupta et al*., 2004]. In both the 2D and the 3D simulations, tropical 50 hPa ozone was ~1% lower in 2000 than in 1980, and tropical 50 hPa temperatures were ~0.25 K lower in 2000 than in 1980. At this altitude the 3D water vapor was ~2% lower in 2000 relative to 1980, while the 2D water vapor was ~3% lower. These results lend credence to the 2D model simulations. The 3D model simulations will be reported in a subsequent publication.

3.2 Changes after 1997

We turn now to an analysis of the results for the years after 1997, when stratospheric chlorine is in decline. Figure 1 shows that in the runs with increasing CO_2 ozone values stay lowered out to the year 2050, while in the run with fixed CO_2 ozone mixing ratios increase, with recovery occurring in the year 2026. It is the increasing carbon dioxide that prevents recovery. Figure 2 shows a much greater post-1997 fall-off in production for the increasing CO_2 case compared to the fixed CO_2 case. As upper stratospheric ozone recovers, ozone blocks UV radiation from penetrating to 51 hPa and reduces production, the reverse of “self-healing”. Upper stratospheric ozone recovers faster with increasing CO_2 due to the cooling and the temperature dependence of O_3 loss rates [e.g., *Rosenfield et al*., 2002]. Thus the fall-off in lower stratospheric production will be greater in the increasing CO_2 case.

4. Summary and Conclusions

In conclusion, model simulations predict that increasing carbon dioxide will prevent the recovery of tropical lower stratospheric ozone out to the year 2050. These results suggest that the tropical lower stratosphere would be a poor place in which to look for signs of chemical ozone recovery. The model results also indicate a decline in tropical tropopause temperatures, and an associated decline in tropical lower stratospheric water persisting until ~2015-2018.

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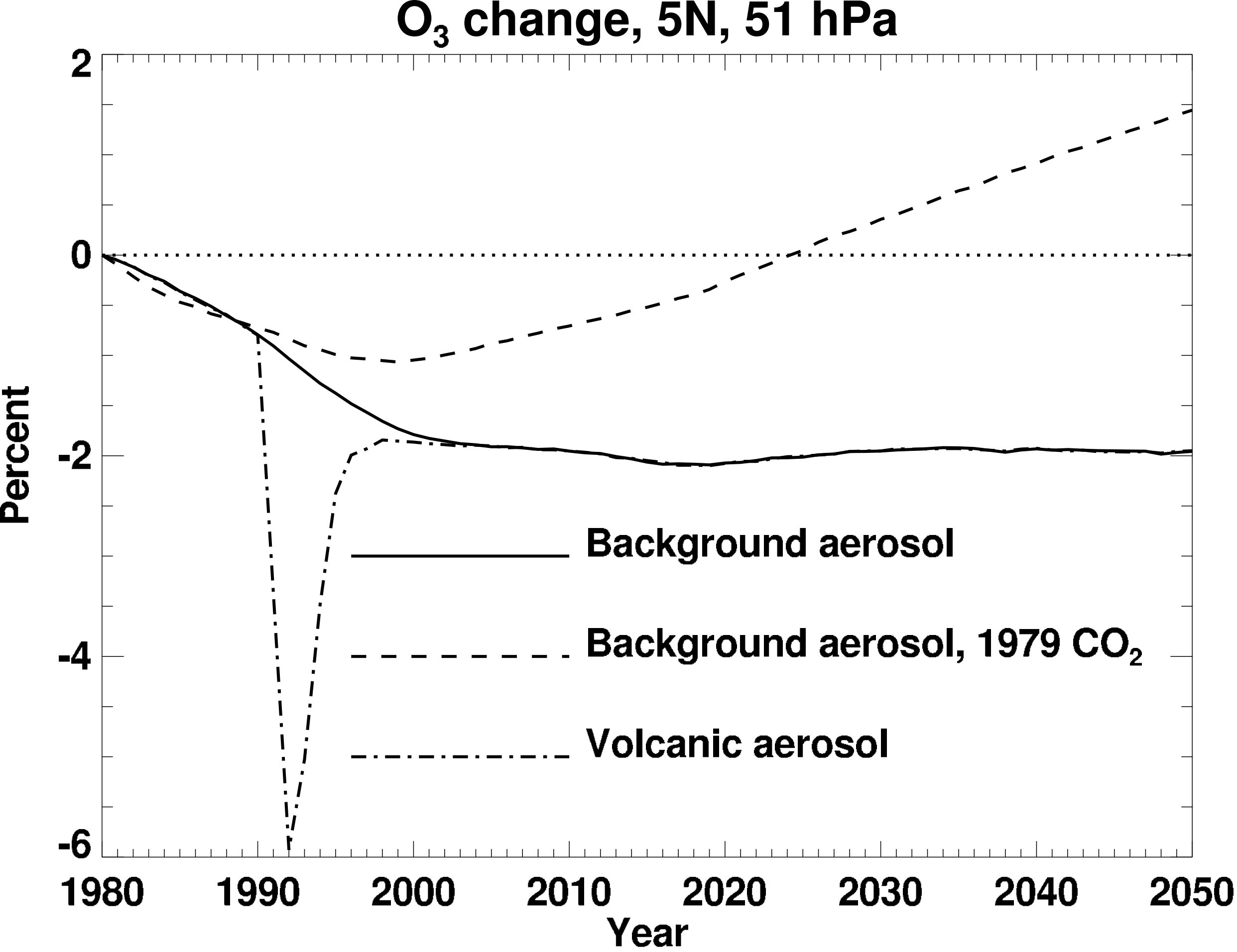
Figure Captions

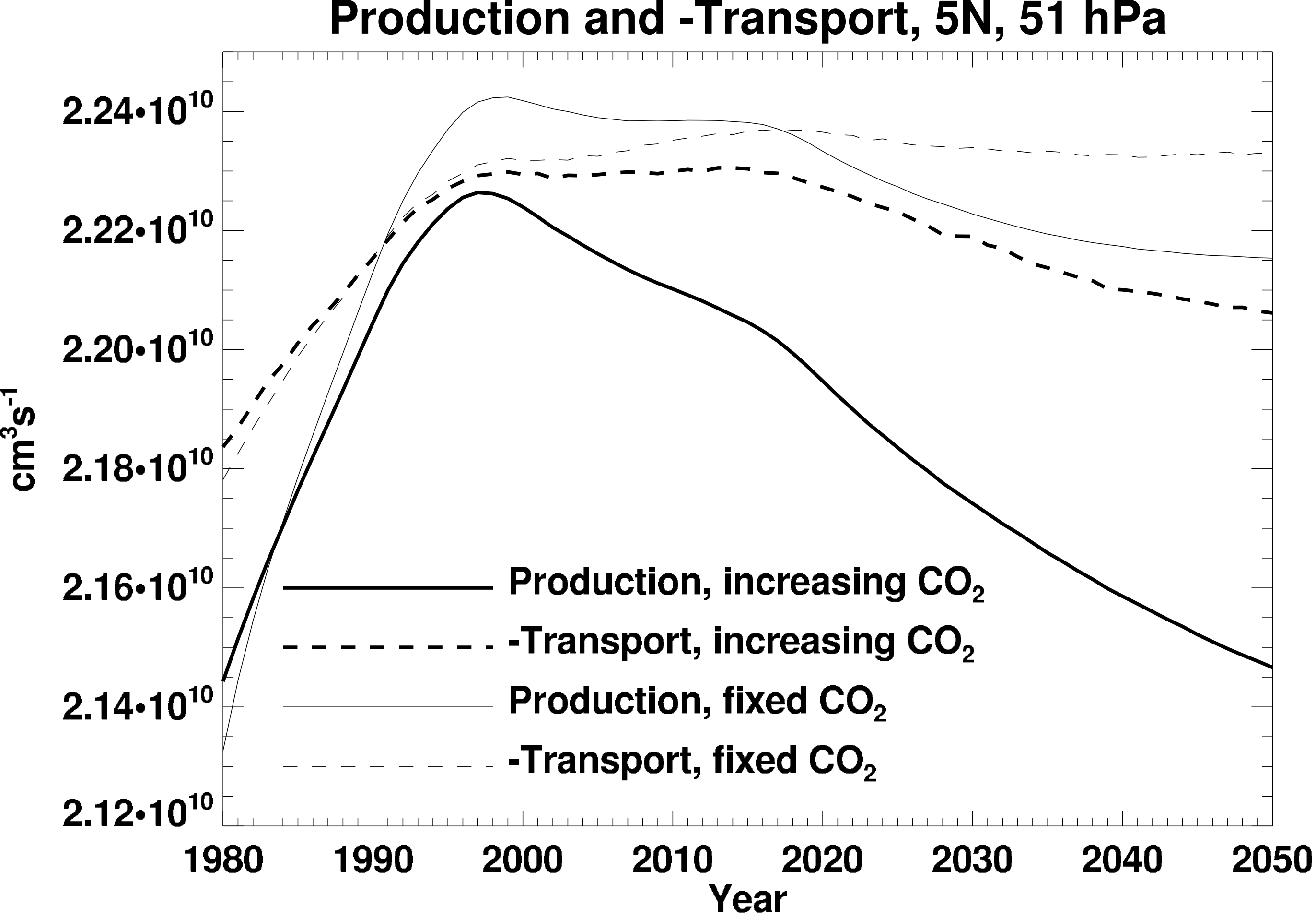
Figure 1. Percent ozone change from 1980 at 5^{o}N and 51 hPa. The solid line refers to the background aerosol case with increasing CO_2, the dashed line to the background aerosol case with fixed CO_2, and the dash-dot line the volcanic aerosol case with increasing CO_2.

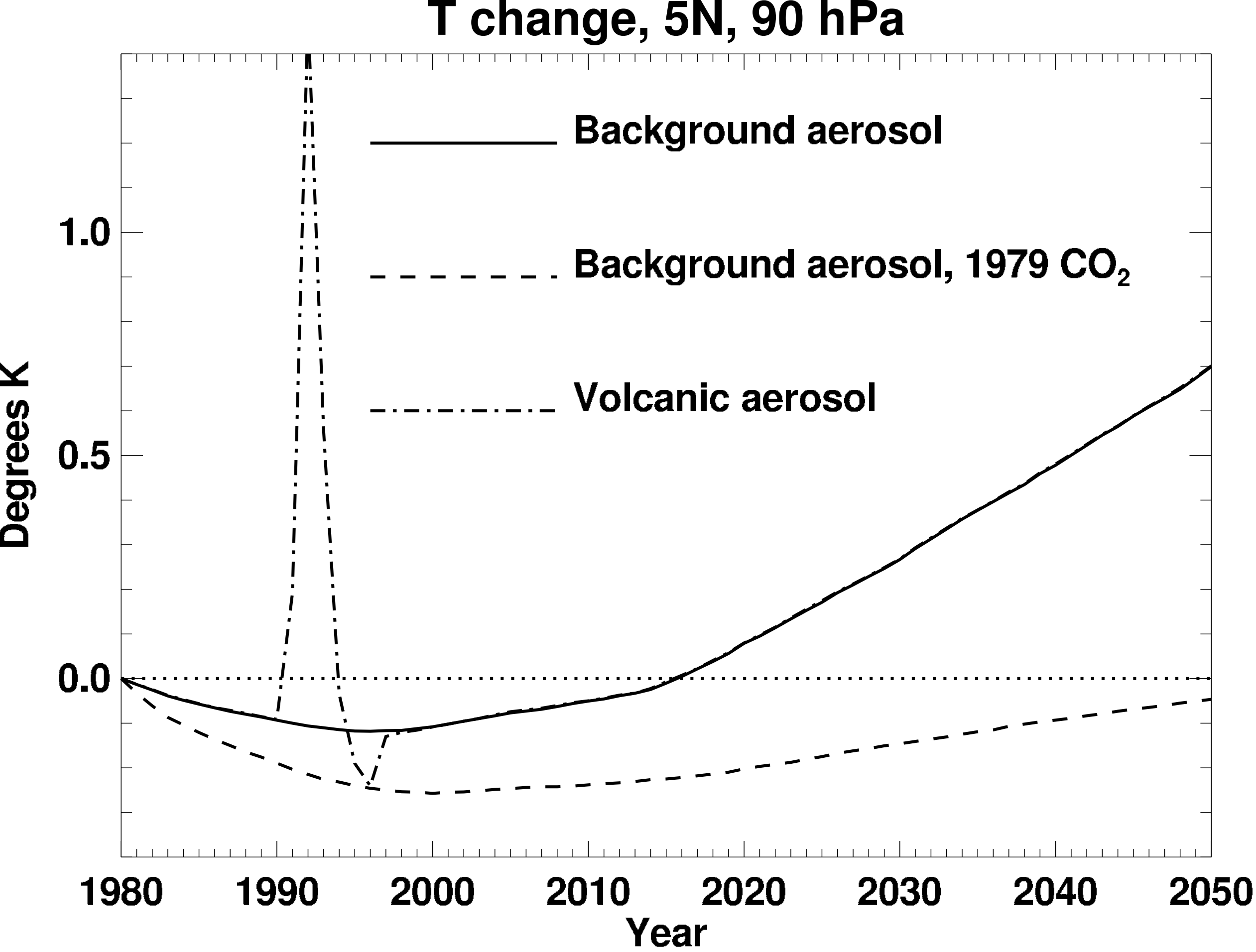
Figure 2. Ozone chemical production and the negative of the ozone vertical transport at 5^{o}N and 51 hPa for the runs with background aerosol. The thick (thin) solid line refers to the production in the increasing (fixed) CO_2 run. The thick (thin) dashed line refers to the negative of the transport in the increasing (fixed) CO_2 run.

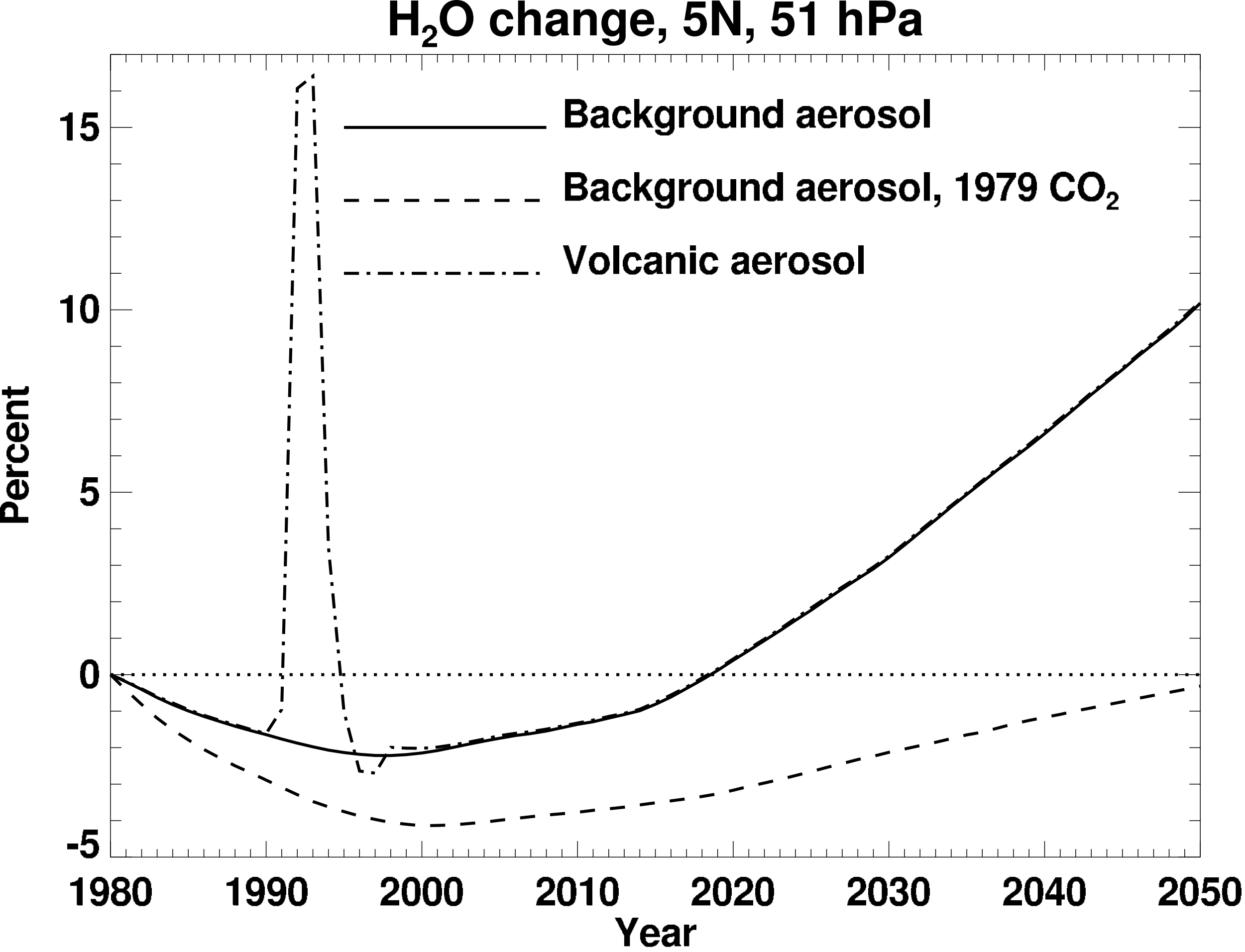
Figure 3. Temperature change from 1980 at 5^{o}N and 90 hPa (the model tropical tropopause). The line styles are the same as in Figure 1.

Figure 4. Percent water vapor change from 1980 at 5^{o}N and 51 hPa. The line styles are the same as in Figure 1.


O_3 change, 5N, 51 hPa
Percent
2
0
-2
-4
-6
1980
1990
2000
2010
2020
2030
2040
2050
Year
Background aerosol
Background aerosol, 1979 CO_2
Volcanic aerosol


Production and -Transport, 5N, 51 hPa
cm^{3}s^{-1}
2.24•10^{10}
2.22•10^{10}
2.20•10^{10}
2.18•10^{10}
2.16•10^{10}
2.14•10^{10}
2.12•10^{10}
1980
1990
2000
2010
2020
2030
2040
2050
Year
Production, increasing CO$_2$
-Transport, increasing CO$_2$
Production, fixed CO$_2$
-Transport, fixed CO$_2$


T change, 5N, 90 hPa
Background aerosol
Background aerosol, 1979 CO_2
Volcanic aerosol
Degrees K
1.0
0.5
0.0
1980
1990
2000
2010
2020
2030
2040
2050
Year


H$_2$O change, 5N, 51 hPa
Background aerosol
Background aerosol, 1979 CO$_2$
Volcanic aerosol
Percent
15
10
5
0
-5
1980
1990
2000
2010
2020
2030
2040
2050
Year