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O_3 and Stratospheric H_2O Radiative Forcing Resulting from a Supersonic Jet Transport Emission Scenario

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ABSTRACT

The tropospheric radiative forcing has been calculated for ozone and water vapor perturbations caused by a realistic High Speed Civil Transport (HSCT) aircraft emission scenario. Atmospheric profiles of water vapor and ozone were obtained using the LLNL 2-D chemical-radiative-transport model (CRT) of the global troposphere and stratosphere. IR radiative forcing calculations were made with the LLNL correlated k-distribution radiative transfer model. UV-Visible-Near IR radiative forcing calculations were made with the LLNL two stream solar radiation model. For the case of water vapor the IR and Near IR radiative forcing was determined at five different latitudes and then averaged using an appropriate latitudinal average to obtain the global average value. Global average values of radiative forcing were approximately $1.2 - 2.6 \cdot 10^{-3} \text{ W/m}^2$, depending on the background atmospheric water vapor profile. This result is consistent with prior published values for a similar aircraft scenario and supports the conclusion that the water vapor climate forcing effect is very small. The radiative forcing in the IR and UV-Visible spectral ranges, due to the ozone perturbation, was calculated for the globally averaged atmosphere. Global average values of the radiative forcing were 0.034 W/m^2 for the UV-Visible spectral range and 0.006 W/m^2 for the IR spectral range (0.04 W/m^2 total). This result is also consistent with the range of published values obtained for a similar HSCT scenario. As was the case for water vapor, the ozone forcing is too small to be of major consequence.

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I. INTRODUCTION

The aviation community has started to investigate the possibility of developing, marketing, and producing a fleet of High Speed Civil Transports or HSCTs (e.g., Stolarski and Wesoky, 1993; Stolarski et al., 1995). These aircraft are designed to cruise at Mach 1.6 - 2.4 with a range of 5000 to 6500 nautical miles. The primary market for this new fleet of supersonic passenger aircraft will be in the Atlantic and Pacific flight corridors, decreasing the average subsonic travel time by more than half. This HSCT fleet will cruise primarily in the low stratosphere, within an ozone rich region. Effluents from this proposed fleet will predominately be contained within the Northern Hemisphere, mid-latitude / lower stratosphere flight corridor. This study addresses the potential climate impact of emissions from a proposed fleet of HSCTs.

Emissions of trace constituents from HSCTs engines are produced from: 1) direct combustion of the kerosene based fuel components forming H_2O , CO_2 , CO , CH_4 , non-methane hydrocarbons, and soot; 2) impurities in the fuel (e.g., sulfur forming SO_2); and 3) high temperature processes, breaking down atmospheric nitrogen forming NO_x ($\text{NO} + \text{NO}_2$). Direct emissions of CO_2 , CH_4 , and non-methane hydrocarbon will perturb the ambient concentrations by a minimal amount, therefore, will not have a detectable climatic influence. However, direct emissions of H_2O are substantial and could affect climate. Emission of SO_2 and soot may affect the radiative forcing in the lower stratosphere by increasing the aerosol loadings. Indirect effects of HSCT effluent on climate may also be important. For example, the HSCT emission of nitrogen oxides has been proposed to catalytically reduce ozone abundances in the stratosphere (Johnston, 1971).

Theoretical reductions in ozone, depending on the altitude of the perturbations, are considered to be climatically important (Lacis et al., 1990). In this study, the climate forcing of both the direct emissions of H_2O and the indirect effect of NO_x emissions on ozone will be addressed.

In a recent publication Rind and Lonergan (1995, hereafter referred to as RL) modeled (using the GISS GCM) the effect of ozone stratospheric and water vapor perturbations, due to a high speed civil transport (HSCT) emission scenario, on the surface temperature of the earth. They concluded that the surface temperature change, over a 30 year period, due to the ozone perturbation was insignificant due to the conflicting influences of stratospheric ozone reduction and tropospheric ozone increase, although this latter effect is uncertain due to the fact that the photochemistry model was not validated for the troposphere. They found, for example, that the global average surface temperature decreased by approximately 0.025°C for years 16 - 20 of their run and increased by about 0.025°C for model years 26 - 30. They also found that cooling was generally maximized at

high latitudes where only ozone decreases were prescribed. For the case of stratospheric water vapor RL increased the stratospheric water vapor content from 3 to 3.2 ppm throughout the middle atmosphere and calculated the tropospheric radiative forcing as well as the surface temperature change over a 50 year time period. They obtain a value of less than 0.05 W/m^2 for the radiative forcing at the tropopause. The resulting surface temperature changes vary between 0.06°C and -0.07°C . They concluded that these temperature changes are not significant and that the direct climate effect due to the HSCT emission scenario is likely to be small for both ozone and water vapor perturbations.

The major purpose of this study is to calculate the radiative forcing due to tropospheric and stratospheric changes in ozone and changes in stratospheric water vapor caused by aircraft emissions from a realistic HSCT scenario. In addition, estimates will be made for surface temperature changes caused by the ozone and water vapor changes so that comparison to the results of RL can be made.

HSCT aircraft emission induced ozone and water vapor changes were derived using the LLNL 2-D CRT (Kinnison et al., 1994a,b). The results of this exercise were published in Stolarski et al., 1995. The annual average change in ozone and water vapor are shown in Figures 1 and 2. The 2-D model input scenario used in the RL study differed from this study in three major ways: 1) the emission scenario used in this study is a more realistic representation of both subsonic and HSCT fleet size; 2) the LLNL 2-D CRT model typically derives a larger indirect ozone change relative to the GSFC 2-D CRT model which was used for the RL study; 3) this study focuses primarily on the radiative forcing change from both the direct water vapor and the indirect effect of NO_x emissions on ozone (i.e., a GCM was not used to derive temperature change).

Radiative transfer calculations were made based on the LLNL 2-D model derived HSCT impact on ozone and water vapor. This procedure is similar to a recent report, Wuebbles et al. (1994), which derived the indirect radiative forcing at the tropopause for a change in tropospheric and stratospheric ozone as well as a stratospheric water vapor perturbation resulting from a methane change scenario. The LLNL 2-D CRT model plus a correlated k-distribution IR radiative transport model (Grossman et al., 1994) were used in the calculations. Indirect global warming potential (GWP) calculations due to the ozone and water vapor changes compared quite well with other indirect methane GWP calculations (IPCC, 1990, 1994). A radiative transfer model has also been developed (Grossman et al., 1993) to calculate UV and visible fluxes and radiative forcing at altitudes between 0 - 80 km. Ozone radiative forcing calculations using both this model and the IR model were used in a model intercomparison study (Shine et al., 1995) and the results compared favorably with other model calculations.

II. IR RADIATIVE FORCING MODELS

A. Infrared Radiative Transfer Model

We have used an infrared radiative transfer model based on the correlated k-distribution algorithm (Grant et. al., 1992, Grossman and Grant, 1992) to calculate the radiative forcing at the tropopause. Calculations were done for a globally averaged atmosphere for ozone. The stratospheric radiative forcing of H₂O, caused by the aircraft emissions, was calculated at five different latitudes ($\pm 60^\circ$, $\pm 30^\circ$, 0°). An appropriate latitudinal average was calculated to determine the globally-averaged radiative forcing at the tropopause. The parameters of the calculations are:

- 1) O₃ forcing, wave number range of 500-1650 cm⁻¹, 25 cm⁻¹ subintervals;
- 2) H₂O forcing, wave number range of 25-3000 cm⁻¹, 25 cm⁻¹ subintervals;
- 3) Tropopause of globally averaged atmosphere at 134 mb;
- 4) Ground Temperature of 291K for global average atmosphere (Peixoto and Oort, 1992).

The radiative forcing is defined as the difference between the net infrared flux at the tropopause in the background, globally averaged atmosphere, and the net flux at the tropopause in the atmosphere with changed H₂O and O₃ profiles calculated for the aircraft scenario. Other trace gases included in the radiative transfer model but unchanged by the aircraft scenario are CH₄, CO₂, and N₂O.

Two interesting aspects of the use of a 2-D CRT latitudinally averaged model atmosphere are the question of the variation of the tropospheric radiative forcing as a function of latitude and the difference between the latitudinally averaged value of the tropospheric radiative forcing and the tropospheric radiative forcing calculated for the globally averaged model atmosphere. Wuebbles et al. (1994) calculated the direct CH₄ and indirect O₃ radiative forcing for five different latitudes between 71°S—71°N for a CH₄ change scenario. The results indicate that the radiative forcing values calculated for the globally averaged atmosphere differ by at most ~5 percent for CH₄ and ~9 percent for O₃, from the average of the forcing values for each latitude. These differences are relatively small and thus the use of the globally averaged atmosphere to determine the average radiative forcing was a reasonable approach to the calculation. The latitude variation of the O₃ radiative forcing was approximately a factor of two, pole to equator. Inspection of Figure 1 indicates that the latitude variation of the ozone perturbation due to the HSCT scenario should produce a radiative forcing variation, with latitude, of the same order as that encountered in the Wuebbles et al. (1994) study. Therefore, we make the same assumption that the ozone radiative forcing in the globally averaged atmosphere will be approximately equal to the latitude average of the radiative forcing.

Because H₂O is a very strong greenhouse gas, the H₂O radiative forcing will strongly depend on the height of the tropopause relative to that of the hygropause because of the amount of background (unperturbed) water vapor above the tropopause. As will be

discussed below, some of the water vapor profiles considered in this study, particularly at high latitudes, have a hygropause which is higher in altitude than the tropopause. This profile structure allows the tropospheric water vapor profile to coincide with the region of the HSCT water vapor perturbation, introducing a variable amount of background water vapor above the tropopause. The lower the tropopause altitude, the larger the amount of background water vapor and thus the smaller the downward radiative forcing. In order to compare the latitude averaged radiative forcing for H₂O with the radiative forcing in a globally averaged atmosphere, Wuebbles et al. (1994) found that the tropospheric radiative forcing for H₂O in a globally averaged atmosphere (166 mb tropopause) was approximately 30 percent less than the globally averaged radiative forcing. Since the location of the tropopause is a critical factor in the radiative forcing it was recommended, for the calculation of H₂O forcing, that the radiative forcing should be calculated as a function of latitude and then averaged as opposed to the calculation of the radiative forcing in the globally averaged atmosphere.

B. UV - Visible - Near IR Radiative Transfer Model

The solar model covers the spectral region from 0.175 μm to 4 μm . It has nine bands covering the UV-visible region from 0.175 μm to 0.700 μm and seven bands resolving H₂O absorption in the near-IR between 0.400 and 0.700 μm . The solution method is a two-stream algorithm using an adding technique to calculate the fluxes at layer edges similar to that published by Grant and Hunt (1969). A spherical atmosphere treatment of the path of the direct solar beam is done for large solar zenith angles. A plane parallel treatment of multiple scattering is used. Sources of diffuse radiation are calculated using the delta-Eddington approximation (King and Harshvardhan, 1986). Reflection and transmission of diffuse radiation are calculated using the Sagan-Pollack model (Sagan and Pollack, 1976). In a manner similar to Chou (1992), coefficients for absorption by ozone and molecular oxygen and for Rayleigh scattering were calculated for each of nine UV-visible bands. Molecular oxygen absorbs only in the first two of these bands. Because the spectral dependence is much smoother and has much less variance than in the near-IR, only single extinction coefficients are needed for each interaction per band. The best coefficient for each interaction and band was estimated by a least squares fit over the expected range of extinction path lengths. Bands were chosen to facilitate overlapping and to keep the range of absorber variation within a band small. The treatment of the near-IR absorption by water vapor was the most complicated, due to the wide range of absorption coefficients within each band. The original flux-weighted k-distribution parameterization for the seven near-IR bands was taken from Table 1 of Chou (1986). While providing better spectral resolution than previous one-band near-IR parameterizations, direct use of Table 1 was undesirable since it contains 199 separate absorption terms. We economized this to a total of 38 absorption terms over the entire seven bands via a combination of correlated k-distribution formalism combined with least squares fitting.

III. RESULTS AND DISCUSSION

Figures 3a — 3e show the temperature-pressure profiles for the background (ambient plus subsonic aircraft) atmosphere at latitudes of 60°S (a), 30°S (b), 0° (c), 30°N (d), and 60°N (e). Horizontal lines indicate the tropopause pressure at each latitude, 322mb at 60°NS, 172mb at 30°NS, and 113mb at the equator. Figures 4a — 4e show the water vapor mixing ratio—pressure profiles at latitudes of 60°S (a), 30°S (b), 0 (c), 30°N (d), and 60°N (e). Horizontal lines indicate the tropopause pressure at each latitude. The LLNL 2-D CRT model determines the water vapor mixing ratio profiles in the following manner. A seasonally varying water climatology (Oort, 1983) is used to determine the profile from the surface to a pressure of ~300mb. The water vapor mixing ratio is interpolated from 300mb to the prescribed minimum water vapor mixing ratio (approximately 3.3 ppmv). Above the prescribed minimum, the water vapor is derived from both methane oxidation and transport from the troposphere. Since the downward radiative forcing critically depends on the background water vapor abundance between the tropopause and the water vapor minimum, the determination of the model prescribed profile between 300mb (climatological cutoff) and the prescribed water vapor minimum (typically 1.5-3.0 km above the tropopause) is very important. Setting the minimum water vapor mixing ratio in the LLNL 2-D model a few kilometers above the tropopause was done to represent the hygropause. A recent study (Abbas et al., 1996), primarily using water vapor measurements from the Atmospheric Trace Molecule Spectroscopy (ATMOS) Fourier transform spectrometer on the ATLAS-3 Shuttle flight (in November 1994) suggest that between 8°N and 49°N, the minimum water vapor mixing ratio is at or very near the tropopause. Therefore, to assess the uncertainty in forcing due to the uncertainty in the water vapor profile, a modified background water vapor profile was constructed by arbitrarily holding the mixing ratio at a constant value between the model prescribed hygropause and the tropopause. The resulting profiles are shown in Figures 5a - 5e. The aircraft perturbation was applied to both profiles and the radiative forcing was calculated. Table 1 gives the radiative forcing due to the water vapor perturbation at each latitude and the latitudinally averaged value for both the original LLNL 2-D model interpolated water vapor profile ("Normal") and the water vapor profile with constant mixing ratio above the tropopause ("Modified"). An estimate of the near IR (0.4 - 0.7 μ m) contribution to the forcing was made using the near IR model on the globally averaged water vapor profiles. This estimate was added to the IR forcing values to obtain a net radiative forcing for the water vapor perturbation and amounts to approximately 20 - 37 percent of the total forcing depending on the water vapor profile used. Inspection of Figure 2 shows that the maximum water vapor perturbation occurs between 30°N and 60°N latitude. The variation of the radiative forcing for the 100 - 300 mb interpolated water vapor profile, shown in Table 1, peaks at the equator and decreases toward the poles for the northern hemisphere in contrast to the perturbation distribution shown in Figure 2. This effect is due to the variation of the background water vapor between the hygropause and the tropopause at latitudes between the equator and 60°N. To a first approximation, the downward radiative forcing is proportional to $\exp(-\tau_{\text{H}_2\text{O}})$, where $\tau_{\text{H}_2\text{O}}$ is the optical depth of the background water vapor. Figure 4 shows the maximum amount of background water vapor between the hygropause and the tropopause at a latitude of 60°NS, and the minimum at the equator. This is the cause of the decrease in the forcing at the higher latitudes. When the amount of background water vapor between the hygropause and the tropopause is

essentially equalized at all latitudes, as is the case for the modified water vapor profile, then the resulting radiative forcing variation shown in Table 1 agrees quite well with the strength dependence of the perturbation shown in Figure 2. In a recent publication Shine and Sinha (1991) calculated the surface temperature changes due to radiative forcing inputs at various altitudes. They found that a radiative forcing input of 0.45 W/m^2 applied at a pressure of 134.4mb (the troposphere of the global average atmosphere used in this study) produced a surface temperature change of $\sim 0.15 \text{ K}$. This gives a climate forcing factor of 1/3. Using this value to estimate the surface temperature change due to the radiative forcing values for water vapor given in Table 1 gives values of $3.92 - 8.76 \times 10^{-3} \text{ C}$, depending on the water vapor profile. These values are shown in Table 2 along with the values published by RL. As shown there, the estimated values calculated in this study lie within the range of values given by RL. RL concludes that the range of surface temperature changes calculated for the water vapor perturbation is too small to be of meaningful significance. The radiative forcing and estimated surface temperature changes calculated in this study support the RL conclusions.

The radiative forcing calculated for the globally averaged ozone perturbation shown in Figure 1 is given in Tables 3 (IR), 4 (UV-Vis.), and 5 (Net). For the ozone perturbation the short wave radiative forcing is the dominant effect, amounting to approximately 85 percent of the total radiative forcing. The surface temperature change associated with the net radiative forcing at the tropopause (134.4 mb) is estimated using a climate forcing factor of $1/3.5 \text{ K/W/m}^2$ (Lacis et al., 1990). The surface temperature change thus obtained is 0.011 C . This value is shown in Table 6 along with the values published by RL. It is seen that the value calculated in this study lies within the range of values given by RL. As was the case for the water vapor forcing, RL concludes that the range of surface temperature changes calculated for the ozone perturbation is too small to be of meaningful significance. The ozone radiative forcing and estimated surface temperature change calculated in this study tends to support the RL conclusions.

The radiative calculations presented in this paper were done under the assumption that the stratospheric temperature distributions were unaffected by the ozone perturbations. Stratospheric temperature changes resulting from the ozone perturbation were included in the RL study. In general, the atmospheric response to the ozone perturbation will be such as to reduce the magnitude of the radiative forcing perturbation. Thus the radiative forcing resulting from the atmosphere in which the stratospheric temperature is allowed to readjust to the ozone perturbation should be smaller than that calculated for the case of no temperature readjustment. Since the radiative forcing calculated in this paper, assuming no stratospheric temperature readjustment due to the HSCT perturbation, was not significant in terms of large scale climate change, then the values that would result if stratospheric temperature readjustment were allowed would also not produce significant climate changes.

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Table 1.Latitude Variation Of HSCT IR H₂O Radiative Forcing (W/m²) [25 - 3000 cm⁻¹]

LATITUDE (DEG)	H ₂ O Forcing	
	Normal H ₂ O	Modified H ₂ O
-60 (322mb)	2.732 (-3)	1.845 (-2)
-30 (172mb)	1.255 (-2)	2.129 (-2)
Eq. (113mb)	3.085 (-2)	3.257 (-2)
+30 (172mb)	2.613 (-2)	4.594 (-2)
+60 (322mb)	4.842 (-3)	4.829 (-2)
IR Avg.	1.876 (-2)	3.327 (-2)
NET*	1.176 (-2)	2.627 (-2)

*Estimate of Near IR radiative forcing $\Delta F \sim -0.007$ **Table 2.**

H ₂ O NET SURFACE TEMPERATURE CHANGE (°C) *		
	Normal H ₂ O	Modified H ₂ O
LLNL	3.92 (-3)	8.76 (-3)
RL	0.06 < ΔT > -0.07	

* ESTIMATE OF ΔT (°C) AT SURFACE $\sim \Delta F / 3.0$ (Shine and Sinha, 1991)**Table 3.**

O ₃ IR RADIATIVE FORCING 500–1650 cm ⁻¹ (W/m ²)	
Global Av. (134mb)	
LLNL	0.00585

Table 4.

O ₃ RADIATIVE FORCING 175–735 nm (W/m ²)	
Global Av. (134mb)	
Diurnal Factor of 0.5	
LLNL	0.034

Table 5.

O ₃ NET RADIATIVE FORCING (W/m ²)	
Global Av. (134mb)	
Diurnal Factor of 0.5	
LLNL	0.040

Table 6.

O ₃ NET SURFACE TEMPERATURE CHANGE (°C) *	
Global Av. (134mb)	
Diurnal Factor of 0.5	
LLNL	0.011
RL	0.025 < ΔT > -0.025

* Surface Temperature Change ~ Radiative Forcing / 3.5 (Lacis et al., 1990)

FIGURE CAPTIONS

- Figure 1. The annual and zonal average change in ozone mixing ratio due to the HSCT scenario as a function of latitude and altitude.
- Figure 2. The annual and zonal average change in water vapor mixing ratio due to the HSCT scenario as a function of latitude and altitude.
- Figure 3. The temperature-pressure profiles for the background (ambient plus subsonic aircraft) atmosphere at latitudes at of 60°S (a), 30°S (b), 0° (c), 30°N (d), and 60°N (e). Horizontal lines indicate the tropopause pressure at each latitude, 322mb at 60°NS, 172mb at 30°NS, and 113mb at the equator.
- Figure 4. The water vapor mixing ratio-pressure profiles for the background (ambient plus subsonic aircraft) atmosphere at latitudes at of 60°S (a), 30°S (b), 0° (c), 30°N (d), and 60°N (e). Horizontal lines indicate the tropopause pressure at each latitude, 322mb at 60°NS, 172mb at 30°NS, and 113mb at the equator.
- Figure 5. The modified water vapor mixing ratio-pressure profiles for the background (ambient plus subsonic aircraft) atmosphere at latitudes at of 60°S (a), 30°S (b), 0° (c), 30°N (d), and 60°N (e). Horizontal lines indicate the tropopause pressure at each latitude, 322mb at 60°NS, 172mb at 30°NS, and 113mb at the equator.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

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