

Effects of Cloud Optical Property Feedbacks on the Greenhouse Warming

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ABSTRACT

Cloud optical properties, in particular the optical thickness τ , affect the earth-atmosphere radiation budget, and their potential changes associated with climate changes may induce feedback effect. A one-dimensional radiative-convective model was used to illustrate that the difference in the vertical distribution of the radiative forcing between CO_2 increase and changes of solar constant can result in a different τ feedback. Recently, Wang et al. carried out a general circulation model study of the climatic effect of atmospheric trace gases CH_4 , CFCs, and N_2O , and the model results indicate that these trace gases provide an important radiative energy source for the present climate. Because the radiative-forcing behavior of CO_2 is different from that of these other gases, the simulations also show that different radiative forcing can lead to quite different climatic effects. Consequently, increases in these trace gases may also induce different τ feedback than that due to CO_2 increases. Since no study was attempted before to address this aspect, here a one-dimensional model is used to investigate the τ feedback associated with trace gases using an updated τ scheme that relates τ to cloud liquid water content through cloud layer latent heat flux. Because of the different changes in the τ vertical distributions, the τ feedback is calculated to be a small negative value for a CO_2 increase, but much larger negative values for increases of trace gases. The strongest negative feedback is found for CFCs.

Similar experiments were also conducted using a revised version of the Somerville and Remer τ scheme, which relates τ to cloud liquid water content through cloud temperature. The results indicate that the negative feedback for CO_2 increases for a single cloud layer becomes much smaller when multiple-layer clouds are used, mainly due to the compensating effect of changes in τ values between high and low clouds. Because this scheme assumes a strong functional dependence of the local temperature, the τ feedback is also found to be sensitive to model dimensionality. In addition, the strength and sometimes even the sign of the τ feedback calculated from both schemes depend on the vertical distribution of cloud cover for the control climate, indicating the complexity of cloud-radiation interactions. Clearly, more observational and theoretical studies are needed to understand the cloud microphysics and their relation to large-scale climate variables.

1. Introduction

Because of their strong radiative effects and close interactions with the hydrologic cycle, clouds play a crucial role in the climate feedback processes influencing climate sensitivity. However, our present understanding of cloud-climate feedbacks is rather poor due to a lack of adequate observational data (see Arking 1991). Thus, climate feedbacks related to changes in clouds have been recognized to introduce major uncertainties in model studies of the greenhouse effect due to increasing atmospheric CO_2 and trace gases CH_4 , N_2O , and CFCs (DOE 1985; WMO 1987; CES 1990).

There are many studies demonstrating the importance of cloud optical properties during climate changes

(Wang et al. 1981; Charlack 1982; Somerville and Remer 1984; Liou et al. 1985; Betts and Harshvardhan 1987; Charlson et al. 1987; Coakley et al. 1987; Albrecht 1989; Mitchell et al. 1989; Platt 1989). These studies indicate that the potential changes in cloud optical properties, in particular the optical thickness τ , during climate changes may substantially modify the climate changes themselves; we refer to the feedback due to changes in τ as the τ feedback. In the present study, the τ feedback strength S (in percent) is defined as

$$S = (\Delta T_r / \Delta T - 1) \times 100,$$

where ΔT_r and ΔT are the surface temperature changes with and without the τ feedback, respectively. Previous studies indicate that for CO_2 increases, the value of S ranges from small positive values (Wang et al. 1981; Liou et al. 1985) to large negatives (Paltridge 1980; Somerville and Remer 1984; Mitchell et al. 1989). The wide range of values can be attributed to the differences

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in the cloud parameterizations (Schlesinger 1985) and the control climates (Roekner 1988) and to the consideration of phase changes in cloud droplets (Mitchell et al. 1989).

The nature of the climate perturbation itself may also affect S . For example, Wang et al. (1981, hereinafter referred as W81) find that while $S = -17\%$ in the 2% solar constant increase experiment, $S = 1\%$ for CO₂ doubling. This difference in the S values is caused by the difference in the vertical distributions of the radiative perturbations between the two cases. Consequently, the different absorption characteristics between CO₂ and other greenhouse gases CH₄, N₂O, and CFCs (Ramanathan et al. 1985; Wang and Molnar 1985; Wang et al. 1991) may also yield different S values than that due to CO₂ increase. It has been shown that the contribution of these radiatively important trace gases to the total greenhouse warming is significant, augmenting the CO₂ warming by about 50%–60% around the middle of the next century (Wang and Molnar 1985). It is therefore important to study to what extent the τ feedback will affect the total greenhouse warming.

Because of their different solar albedo and greenhouse effects, low and high clouds affect S differently (see Fig. 1 of W81). For low clouds, the solar albedo effect dominates; thus, their changes will lead to a negative feedback. On the other hand, for high, optically thin clouds, the greenhouse effect dominates and their changes will induce a positive feedback. However, recent work by Stephens et al. (1990) questions this simple picture, based on model simulations of cirrus climatic effects as a function of cloud microphysical properties. At any rate, it seems unavoidable to use multiple cloud layers to study the τ feedback. These arguments lead to the focus of our present work: to study the τ feedback in connection with the increases of the trace gases in models with multilayered clouds.

In addition, the wide range of S values in the case of a CO₂ increase points to a need to address the effect of different schemes to parameterize τ feedback. Thermodynamical considerations by Betts and Harshvardhan (1987), as well as empirical evidence presented by Somerville and Remer (1984, hereinafter referred as SR84), indicate that, for certain types of clouds, a temperature-dependent parameterization for τ may be appropriate. We refer to this parameterization as the CT scheme, since it relates changes in cloud water content to changes in cloud temperature through

$$f = \frac{1}{\rho} (\partial\rho/\partial T_c),$$

where ρ is cloud liquid water content and T_c is cloud temperature. Studies also indicate that f has a strong temperature and pressure (altitude) dependence. A similar feature has been shown to exist for cirrus clouds

by Platt and Harshvardhan (1988) and Platt (1989). Thus, introducing low and high clouds may alter the τ feedback found by SR84, since they used only a single, middle cloud with $f = 0.05$. This f value implies a significant S value, about -50% to -60% (see Table 3 of SR84). Considering that the sign of τ feedback can be different for high and low clouds, the cloud-distribution adopted for the current (control) climate also may have a strong influence on S . For instance, using larger than observed low cloud amounts may introduce a bias toward negative feedback. Furthermore, the latitudinal characteristics of f through its temperature dependence can also affect the S value.

The above considerations lead us to employ the following two schemes to address the possible role of τ feedback in greenhouse warming:

- 1) the cloud/precipitation (CP) type of scheme, since it relates τ to cloud layer latent heat flux convergence (thus precipitation), and in its original form as in W81, provides a small positive S for CO₂ increase; and
- 2) the cloud/temperature (CT) type of scheme, which is local in nature, and according to its first use by SR84, can induce a large negative S value for CO₂ increase.

Note that, at least in their original form, the two schemes bracket the S values of all previous studies of the CO₂ increase. Furthermore, we have reasons not to consider other schemes here. For example, Betts and Harshvardhan (1987) have illustrated that scaling cloud liquid water content to saturation mixing ratio, as essentially Paltridge (1980) and Charlock (1982) did, implies larger f values than that derived from thermodynamical theory. This is especially true for low and middle clouds, thus leading to an overestimation of their (negative) contribution to S . The τ -feedback parameterization used by Liou et al. (1985) is similar to the CP scheme, since they scale cloud liquid water content according to saturation latent heat flux.

Similar to W81, we used a one-dimensional radiative-convective (1D RC) model. For one case in the CT scheme, however, a model with latitudinal resolution is used. Section 2 presents a brief description of the model and the important characteristics of the control climates for both the CP and CT schemes. In section 3a, we use the CP scheme to study to what extent S depends on the nature of the climate perturbation itself. Results of perturbation experiments for doubling of CO₂, N₂O, and CH₄, as well as for CFCs increases, are described and discussed. For comparison, we employ the CT scheme to conduct the same experiments, and the results are presented in section 3b. Also included in section 3 are the studies of the sensitivity of the τ feedback of the two schemes to different vertical and latitudinal cloud distribution. We summarize our conclusions in section 4.

2. Model

The one-dimensional RC model adopted for this study is similar to the one used in W81. The lapse rate is prescribed to be 6.5 K km^{-1} . The fixed cloud altitude parameterization is used throughout our climate perturbation experiments, and ice-albedo feedback is included. During climate perturbation calculations, the relative humidity and control climate fractional cloud cover are kept fixed.

However, the present version has an important refinement over the earlier one. We modified the lapse-rate adjustment to include a drag law parameterization that allows for the separate determination of the surface sensible (F_s) and latent heat fluxes (F_l). Therefore, the Bowen ratio, $B (= F_s/F_l)$, can be calculated explicitly, rather than being approximated as in W81.

Following Lindzen et al. (1982), the sensible and latent heat fluxes are calculated,

$$F_s = C_d u_* C_p (T_g - T_s);$$

$$F_l = LC_d u_* [q_s(T_g) - r q_s(T_s)],$$

where $C_d (=0.0024)$ is the aerodynamic drag coefficient, u_* ($=6.5 \text{ m s}^{-1}$; Oort 1982) is the surface boundary layer wind speed and $r (=75\%)$ is used as the surface relative humidity. The saturation water vapor mixing ratio q_s is calculated from the Clausius-Clapeyron relation, while T_g and T_s denote the ground and surface-air temperatures, respectively. For the latent heat of evaporation L , the temperature-dependent expression of Stone and Carlson (1979) is used.

The one-dimensional control climate cloud fraction and τ distributions are calculated according to W81, which allows cloud to form in any tropospheric model layer. As shown in Table 1, the vertical τ -distribution generally reflects the decreasing magnitude with altitude. The planetary albedo, ice line, surface temperature, and outgoing longwave radiation are all in good agreement with observations. Unless otherwise noted, this model climate will be the control climate in the perturbation experiments with both the CP and CT schemes.

3. Results

In this section results of the τ feedback using both the CP and CT schemes are presented. In addition, we show the sensitivity of the results to the vertical and latitudinal cloud distributions.

a. Cloud/precipitation scheme

As discussed in W81, the τ for a given cloud layer can be expressed as

$$\tau = H_c f_3 [f_1 L (1 + B)]^{-1}.$$

Here the cloud optical thickness is proportional to the

TABLE 1. One-dimensional model simulated characteristics of the present climate.

A. General characteristics		
Parameter	Characteristics ^a	
Planetary albedo	0.30 (0.31)	
Surface temperature (K)	287.5 (287.9)	
Ice line, $x_s = \sin(\text{lat})$	0.95 (0.95)	
Outgoing longwave radiation (W m^{-2})	236.5 (237.0)	
B. Cloud characteristics		
Altitude (km)	Cloud cover (%)	Optical thickness ($0.55 \mu\text{m}$)
High (cirrus)		
9-12	2.7	0.68
7-9	6.6	1.54
Middle (water)		
6-7	7.7	7.54
4.5-6	7.8	7.56
3-4.5	6.9	6.92
Low (water)		
2-3	5.9	15.56
1-2	5.7	15.01
0-1	2.9	7.59

^a The values in parenthesis are observations taken from Wang and Molnar (1985).

total convective heat flux heating rate (H_c) in the layer and inversely proportional to B . The precipitation rate P is related to the cloud liquid water content through the relation, $P = f_1 \rho$, while $f_1^{-1} = 8 \times 10^3 \text{ s}$ represents a conversion time of cloud droplets into rain. The f_3 parameter is altitude dependent, with values of 3.09×10^4 ($z \leq 3 \text{ km}$), 1.15×10^4 ($3 < z \leq 8 \text{ km}$), and 2.75×10^3 ($z > 8 \text{ km}$). The above relation is based on the assumption that the global and annual mean vertical cloud distribution can be related to the mean vertical precipitation distribution. In other words, the CP scheme adopts the hypothesis that higher precipitation rates can result from more vigorous cloud systems. Since for our experiments $B \geq 0$; the decrease of the Bowen ratio with increased temperature yields an increase in τ in the case of a greenhouse warming. On the other hand, even if the total convective heat flux increases for a climate perturbation, the local convective heating at a particular altitude may either increase or decrease, depending to a large extent on the radiative perturbation profile (see discussion later). The change in local τ with increasing temperature may thus have either sign. Note that cloud fractions also can change in the W81 scheme, but we keep them constant to isolate the τ feedback.

In order to assess the effect of the vertical structure of the initial radiative perturbation on S , we have performed separate sensitivity experiments with and without τ feedback for specified increases of several

trace gases. A summary of these model runs is presented in Table 2.

The calculated surface air temperature changes without the τ feedback are consistent with our previous results (see Wang and Molnar 1985); however, the effect of the τ feedback is different for different gases. The biggest difference is between the cases of CO_2 and the CFCs, with values of -13% and -31% , respectively. The S values are insensitive to the use of ground temperature or the surface air temperature. Note that the S value for CO_2 doubling was found to be $+1\%$ in the W81 study. To check the consistency, the same experiments have been performed using the convective flux treatment of W81. The S value is a small positive one, about $+2\%$. This result implies that the current -13% S value is mainly due to the explicit calculation of the Bowen ratio. Indeed, we have found that the decrease of B is about twice the amount of that calculated with the approximate formula introduced in W81. Consequently, the change of the $(1+B)^{-1}$ factor will be greater for the current Bowen ratio parameterization. Also, since B is assumed to be the same at each cloud layer, the decrease of f_3 with altitude will yield the albedo effect of low clouds to be more dominant when the change of B is larger.

Figure 1 shows that changes in the vertical τ distribution are different for different gases. For CFCs increases, the τ increases in each cloud layer, leading to a rather large negative S value. In the CO_2 doubling case, however, the τ values of the low and middle clouds, especially the uppermost low cloud, decrease considerably, causing a significant positive contribution to the overall τ feedback. In addition, the increases of the cirrus τ values are not negligible, providing a larger greenhouse contribution than that for the CFCs' case. As for CH_4 and N_2O , changes of the vertical τ distributions are very similar to the CFCs above ~ 5 km. This means that the greenhouse effect due to increases of cirrus τ is small. On the other hand, the cloud τ values in the middle troposphere change very little, yielding S values in between the two extreme cases.

Figure 2 illustrates that the quantitative differences in the S values can be attributed mainly to the differences in the initial radiative forcing profiles: the shapes

TABLE 2. The effect of τ feedback on the surface air temperature ΔT_s ($^{\circ}\text{C}$). The feedback strength S is included in the parenthesis.

Case	No τ feedback	τ Feedback	
		CP scheme	CT scheme
CO_2 : 330 \rightarrow 660 ppmv	2.88	2.49 (-13)	1.68 (-42)
F-11: 0 \rightarrow 2 ppbv	0.57	0.39 (-31)	0.34 (-41)
F-12: 0 \rightarrow 2 ppbv			
N_2O : 0.3 \rightarrow 0.6 ppmv	0.72	0.55 (-22)	0.44 (-39)
CH_4 : 1.7 \rightarrow 3.4 ppmv	0.48	0.38 (-20)	0.33 (-30)

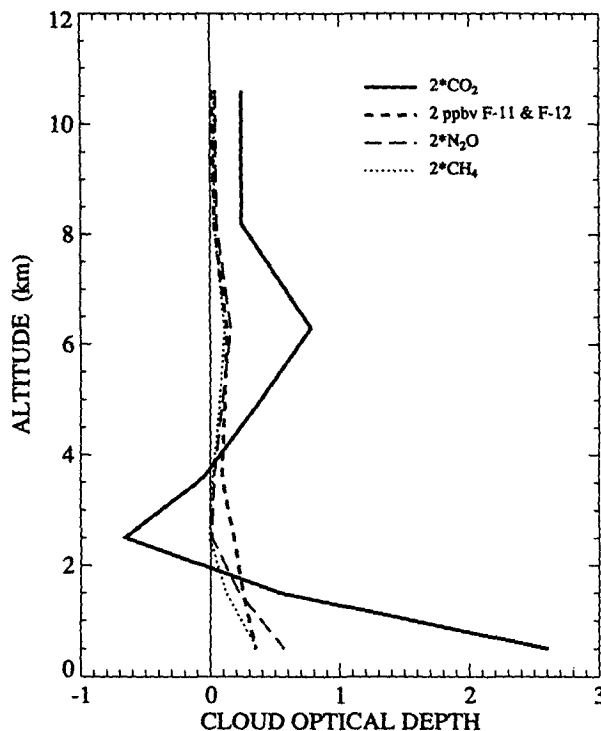


FIG. 1. Changes of vertical distribution of cloud optical depth calculated with the CP scheme.

of their curves are similar to their corresponding changes in the τ distributions. What happens is that within the convective adjustment scheme, essentially the change of the total convective heating rate has to "compensate" for the initial radiative perturbation change in order to reach a new equilibrium climate, thus leading to different S values for different trace gases.

b. Cloud/temperature scheme

This parameterization can be implemented into a wide variety of climate models in which the temperature is a prognostic variable. SR84 and Schlesinger (1988) evaluated this scheme using a 1D RC model with one cloud layer. However, as discussed earlier, the use of a single cloud layer may lead to biased S values. Here we repeat the sensitivity experiments employing the CT scheme.

For multiple cloud layers, the CT scheme needs to include the vertical distribution of the f parameter, which is provided in the theoretical study of Betts and Harshvardhan (1987). Assuming that clouds are formed by ascent along a moist adiabat, they calculate the f parameter for a wide range of atmospheric temperatures and pressures. The calculated values were in good agreement with those observed and used by SR84. To be consistent, we have used the f values calculated

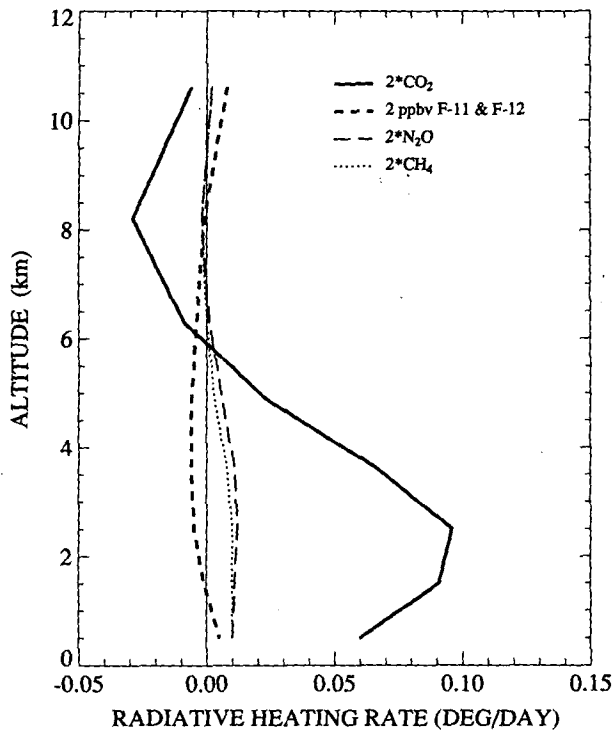


FIG. 2. Changes of vertical distribution of initial radiative forcing with the CP scheme.

using the Betts and Harshvardhan (1987) formulas. Since they considered only water clouds, the observational findings by Platt and Harshvardhan (1988) were used to calculate τ values for cirrus. Thus, the τ feedback scheme, referred to as the CT scheme, is essentially a combination of a simple thermodynamical theory for water clouds and observed relationships for ice clouds. For consistency, we have replaced the high-cloud τ values of the CP scheme control climate by τ values corresponding to the cirrus τ parameterization of Platt and Harshvardhan (1988). It is interesting to note that these τ values are quite close to those obtained with the CP scheme, that is, the 0.68 and 1.54 cirrus τ values in Table 1 are changed to 0.60 and 1.79, respectively.

A summary of the results with the CT scheme is also given in Table 2. For CO_2 doubling, the surface temperature change is 1.68 K, indicating a significant negative feedback. However, as opposed to the approximately $S = -60\%$ obtained by SR84 for a middle-layer cloud, the value is smaller: -42% . The reason for this lies mainly in the inclusion in our study of high clouds and, to a less extent, low clouds. For high clouds, the increase in τ leads to a positive feedback because the solar albedo effect (negative feedback) is more than compensated by the thermal greenhouse effect (positive feedback). In addition, the f parameter is larger for high clouds than for middle clouds, which further en-

hances this effect. Low clouds, on the other hand, contribute to a negative feedback, but the relative magnitude of the effect is small because the f parameter is smaller for these clouds than for middle ones (see Betts and Harshvardhan 1987).

A similar -41% S value is calculated for the case of a 2 ppbv CFCs increase. The total τ change is 1.09, versus the 5.56 calculated for the CO_2 case. It is interesting to note that the surface temperature changes linearly with the changes of total τ . Contrary to the CP scheme, the magnitude of the negative feedback in the CT scheme appears to be less dependent on the type of climate perturbation. This finding can be explained in terms of the basic assumptions of the CT scheme. On one hand, f can be regarded as independent of the nature of climate perturbation for the relatively small perturbations considered here. On the other hand, the local temperature change is uniform throughout the troposphere due to the prescribed, altitude-independent lapse rate. These arguments are illustrated in Fig. 3, which shows the vertical distributions of the τ changes for these sensitivity experiments. The curves are of the same shape, and the optical thickness increases for every cloud layer. In summary, even with a multilayered cloud distribution, the CT scheme still leads to a significant negative feedback on the surface temperature. However, the feedback magnitude is smaller than that obtained with a single, middle-level effective cloud layer as found in SR84.

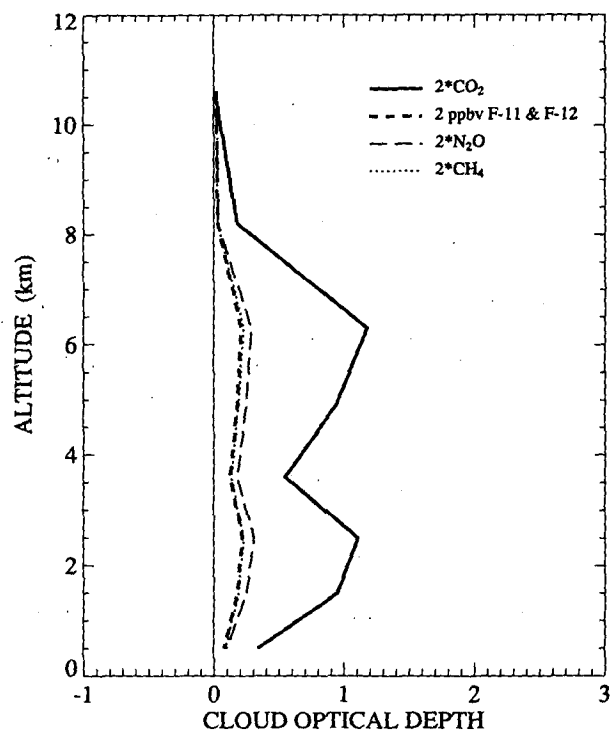


FIG. 3. Same as in Fig. 1 except for the CT scheme.

c. Sensitivity of τ feedback to vertical cloud distribution

Roeckner et al. (1987) analyzed a GCM simulation of the changes in radiative flux at the surface for solar constant changes and concluded that S is a negative value for the CT scheme. This study of the τ feedback has been criticized by Schlesinger (1988), who demonstrated that the sign of the τ feedback is determined by the sign of the net radiative forcing at the tropopause. Additional simulation of the τ feedback by Roeckner (1988) shows that S is +9%. This value is rather far from the value of -42% that we obtained. In addition to the differences in model dimensionality (3D versus 1D), there are two factors that cause such a difference. First, Roeckner (1988) considered only water clouds since they used the Betts and Harshvardhan (1987) relationships exclusively in their parameterization (Roeckner, personal communication 1989). However, Platt and Harshvardhan (1988) have shown that this can lead to an overestimation of the f parameter for high (cold) clouds by about 20% to 75% (depending on latitude). This causes an overestimation of the positive τ feedback of the high clouds. In fact, when Schlesinger (1988) simulated the effects of the Roeckner et al. (1987) τ feedback with a single effective cloud layer, he had to use a very low τ value (1.1), which is characteristic for cirrus. The S value in Roeckner (1988) would most likely be negative if the Platt and Harshvardhan (1988) cirrus was used. A more recent GCM simulation by Mitchell et al. (1989), which allows for phase changes of the cloud droplets, calculated a significant negative S value for CO₂ doubling. In addition, the fractional cover of high clouds in our control climate is probably too low (0.093), which will also cause a relatively small positive feedback contribution from the high clouds.

The above considerations suggest that the vertical cloud distribution can significantly affect S . In order to qualitatively estimate such effects, we adopt a three-cloud model atmosphere documented in Wang et al. (1984) to run the CO₂-doubling experiment with the CT scheme. This three-cloud model, in which the cirrus fractional cover is twice the amount used in the study described above, calculates an S value of -33%. As expected, the increase of the fractional cover of cirrus causes a reduction in the magnitude of the negative τ feedback. A more detailed evaluation of the τ feedback is summarized in Table 3; the changes of the net radiative forcing at the tropopause for the corresponding changes of the individual τ are also shown. In line with Schlesinger's (1988) findings for the high-cloud τ change, the forcing at the tropopause is positive, indicating a positive contribution to the overall τ feedback. Nevertheless, the negative forcings for low and middle clouds overwhelm the positive cirrus radiative forcing, leading to a considerable negative S value.

TABLE 3. Changes in cloud τ , the net radiative forcing at tropopause F (W m^{-2}), and cloud temperature T_c (K) due to CO₂ doubling. The CT scheme is used in the 1D RC Model with three cloud layers.

Parameter	Cloud type		
	High	Middle	Low
Total τ	1.78	5.18	11.84
$\Delta\tau$	0.22	0.58	0.87
ΔF	0.25	-0.70	-0.84
ΔT_c	1.93	1.93	1.93

Thus, in the case of the CT scheme, within a 1D model framework, the high clouds seem unable to counteract the negative τ -feedback contribution due to the middle and low clouds. In fact, in a subsequent experiment we gradually increase the τ changes for high clouds up to the point when the albedo effect started to overwhelm the greenhouse effect (similarly to Fig. 1 of W81), but the positive cirrus τ feedback component never reached a magnitude large enough to result in a total positive τ feedback. In summary, with a single effective cloud layer we can arrive to a large negative (SR84) or small positive (Schlesinger 1988) τ feedback, depending on the control climate cloud optical thickness. However, in the case of more realistic multilayered clouds, it is difficult to obtain a small negative τ feedback, let alone a positive value. It appears that the latitudinal aspect may also contribute to the differences in the τ feedbacks between the CP and CT schemes in the CO₂ doubling case.

d. Sensitivity of τ feedback to latitudinal cloud distribution

Studies by Betts and Harshvardhan (1987) and Platt and Harshvardhan (1988) clearly indicate that the f parameter of the CT scheme is strongly dependent on temperature. Therefore, it is necessary to consider the latitudinal aspect in the CT scheme. For example, the vertical f profile will be different between the tropical and high latitudes since the cloud temperatures are different. To estimate the effect of the latitudinal variation of f and cloud temperature on the global τ feedback, the CT scheme is implemented into the annual, coupled high- and low-latitude radiative-dynamical model described in Wang et al. (1984). For CO₂ doubling, the value of S is calculated to be -13.5% (versus -33% from the 1D model), a value very close to the -13% calculated with the CP scheme. The values of S are -11% for the low-latitude zone and -16% for the high-latitude zone. The surface temperature changes with the τ feedback included are smaller than the 1D model result, 1.2°C for the low latitudes and 1.32°C for the high latitudes. Consequently, to explain the large difference in S values between the two model results, the relative changes of τ values, shown in Table 4, need

TABLE 4. Changes in cloud τ and cloud temperature T_c (K) due to CO₂ doubling. The CT scheme is used in the high- and low-latitude model of Wang and Molnar (1985).

Cloud type	Tropics			Extratropics		
	Total τ	$\Delta\tau$	ΔT_c	Total τ	$\Delta\tau$	ΔT_c
Low	9.10	0.26	1.35	6.05	0.49	1.56
Middle	4.55	0.31	1.92	2.75	0.29	1.67
High	1.90	0.75	4.04	1.30	0.30	2.34

to be examined. In the tropical zone the change of cirrus τ relative to the changes for low and middle clouds is much larger than in the 1D case, thus leading to a significant decrease of the negative feedback. This is due partly to the large high cloud temperature change caused by the use of the moist-adiabatic lapse rate in the low-latitude zone. On the other hand, for the warmer low and middle clouds, the f parameter is small, reducing the negative feedback contribution from these clouds. For the high-latitude zone, the change of τ is small for high clouds but is larger for low clouds, leading to an increased τ feedback. However, the relative change of cirrus τ value is still much greater than that in the 1D simulation, thus reducing considerably the magnitude of the negative feedback. These calculations illustrate that the CT scheme is very sensitive to the local conditions.

4. Conclusions

Previous studies of the effects of cloud optical property feedback considered the increases of CO₂. Here we investigate the τ -feedback effect associated with increases of trace gases CH₄, N₂O, and CFCs, which have a different radiative-forcing perturbation than CO₂. Highlights of the findings are summarized below.

1) For the CP scheme that parameterizes cloud τ with precipitation, the results suggest that the vertical distribution of radiative perturbations caused by the increase of the trace gases lead to a stronger τ feedback than that caused by CO₂ increase. The difference is particularly large between the CO₂ and CFCs, with the former giving a small negative feedback and the latter a much larger negative feedback. Thus, the τ feedback may be sensitive to the nature of the climate changes.

2) For comparisons, we have also investigated the feedback effect using the CT scheme, which relates τ directly to cloud temperature. This scheme, which yields a strong negative τ feedback for a single cloud layer, calculates a much smaller negative feedback in a multilayer cloud atmosphere. The model simulations also indicate that the τ feedback depends strongly on the convective parameterization. For example, when a moist-adiabatic lapse rate is used to simulate the vertical heat transport, the large upper-tropospheric temperature changes and thus large changes of τ for the

high clouds result in a larger compensation to the negative feedback due to changes in τ of the low clouds. The results also suggest that the presence of a relatively large cirrus fractional cover in the control climate may even lead to a strong positive τ feedback (see Somerville and Jacobellis 1989).

The satellite microwave measurements tend to confirm the correlation between precipitation and τ used in the CP scheme (Curry et al. 1990). On the other hand, recent analyses of satellite-retrieved cloudiness (Tselioudis et al. 1990; Rossow 1990) show that τ decreases with temperature for low and high clouds over low-latitude oceanic regions, which implies a deviation from adiabatic behavior. Nevertheless, the large sensitivity of the τ feedback to the vertical distributions of cloud cover and τ for the control climate highlight the importance of improving existing cloud climatologies and measuring cloud water content. Explicit consideration of cloud microphysical properties and the three-dimensional geometry of clouds may also be necessary to successfully relate τ to model variables. Furthermore, as pointed out by Wang et al. (1991), the different climate responses between CO₂ and other trace gases caused by the difference in their spatial distribution of atmospheric opacity suggest a strong coupling between radiation and large-scale dynamics. Much more observational and theoretical research studies such as those of the ARM (1990) and ISCCP are needed to better understand the cloud microphysics and optical properties as well as their relation to large-scale climate variables.

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