Indirect Aerosol Forcing by Homogeneous and Inhomogeneous Clouds

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ABSTRACT

It has been hypothesized that over the past ~200 years, industrial activity has enhanced the number of cloud condensation nuclei (CCN) in the lower atmosphere thereby reducing cloud droplet effective radii $r_e$ and increasing the albedo of clouds. It is thought that in some regions, cloud albedos have increased so much that they have greatly ameliorated coincidental forcing by increased concentrations of greenhouse gases. The best estimates of this ameliorating effect come from large-scale climate/chemical transport models that assume clouds to be horizontally homogeneous at scales smaller than several hundred kilometers. It is demonstrated here that for a 2-$\mu$m reduction in $r_e$, conventional estimates of increased cloud albedo due to more CCN may be too large by up to, and possibly exceeding, 50%. The largest overestimates occur when reductions to $r_e$ are accompanied by enhancements to both cloud variability and liquid water paths. This is attributed to fundamental differences in the way homogeneous and inhomogeneous clouds transport solar radiation.

1. Introduction

Twomey (1974, 1977) has discussed how altering the number of cloud condensation nuclei (CCN) could change cloud albedo thereby affecting an external forcing of Earth’s climate. Basically, as aerosol number concentrations increase via mounting anthropogenic pollution, cloud water is distributed among more CCN thus reducing mean cloud droplet size and enhancing albedo. The Twomey effect, as it has come to be known, received relatively little attention until Charlson et al. (1987) published their Gaia-related hypothesis that links phytoplankton emission of dimethylsulfide to cloud albedo. Subsequently, the Twomey effect has been expanded into what is now referred to as indirect aerosol forcing. Indirect aerosol forcing of cloud albedo has, in turn, been invoked to help explain why global climate models (GCMs) often overestimate surface air temperature changes through the industrial era. In reality, the forcings due to both aerosols and greenhouse gases have increased over time and the climate system has responded accordingly. This makes it difficult to speak of a forcing followed by feedbacks dedicated to restoring Earth’s top of atmosphere (TOA) energy budget. In fact, very few studies have broached the topic of climatic response (feedbacks) to indirect aerosol forcing (e.g., Lohmann and Feichter 1997).

Nevertheless, the most reliable estimates of indirect forcings and feedbacks come from GCMs with large horizontal grid spacings and solar radiative transfer algorithms that assume clouds to be homogeneous at scales smaller than several hundred kilometers. As numerous studies (e.g., Cahalan et al. 1994; Oreopoulos and Barker 1999; Barker and Fu 2000) have documented systematic differences in the way realistic and homogeneous clouds transport radiation, the purpose of this note is to demonstrate that radiative transfer codes used currently in GCMs are likely to overstate indirect aerosol forcing.

2. Radiative transfer calculations

The gamma-weighted two-stream approximation (GWTSA) (Barker 1996) was used here to demonstrate how neglect of unresolved cloud variability by GCMs might impact estimates of indirect aerosol forcing. While the GWTSA is a simple, straightforward model,
it has been shown to be accurate for a wide range of conditions (Barker et al. 1996; Oreopoulos and Barker 1999; Barker and Fu 1999). It assumes that normalized distributions of cloud optical depth $\tau$ within GCM grid cells can be described approximately as

$$ p(\tau | \bar{\tau}, \nu) = \frac{1}{\Gamma(\nu)} \left( \frac{\nu}{\bar{\tau}} \right)^{\nu-1} e^{-\nu \tau / \bar{\tau}}, $$

where $\Gamma(\nu)$ is the gamma function, $\bar{\tau}$ is the mean of $\tau$, and $\nu$ represents cloud variability and depends on $\bar{\tau}$ and either $\log(\bar{\tau})$ or $\var(\tau)$ depending on how it is estimated (Barker et al. 1996). The albedo of a cloud layer for collimated irradiance is approximated as

$$ R_c(\bar{\tau}, \mu_o) = \int_0^\infty p(\tau | \bar{\tau}, \nu) R_c(\tau, \mu_o) \, d\tau, $$

where $R_c$ is from any two-stream approximation, and $\mu_o$ is cosine of solar zenith angle. For details and closed-form solutions for (2), see Oreopoulos and Barker (1999). Note that in the homogeneous limit,

$$ R_c(\bar{\tau}, \mu_o) \approx \lim_{\nu \to \infty} \int_0^\infty p(\tau | \bar{\tau}, \nu) R_c(\tau, \mu_o) \, d\tau. $$

Broadband solar fluxes were computed using Fu and Liou’s (1993) parametrization of cloud droplet optical properties and Fu and Liou’s (1992) correlated-$k$ distribution method for gaseous absorption.

If $R_a$ and $R_c$ denote albedos at cloud top, let $\alpha_a$ and $\alpha_c$ denote corresponding TOA albedos. Given the preliminary nature of the experiments reported on below, just climatological values are discussed and so results are shown only for $\mu_o = 0.5$ and dependencies on $\mu_o$ are omitted hereinafter. Furthermore, because the current intention is to demonstrate relative differences between changes to $\alpha_a$ and $\alpha_c$, only overcast skies need be considered.

3. Methodology

Basically, indirect aerosol forcing begins with increased concentrations of sulfate aerosol, which increase the number of CCN and decrease cloud droplet effective radii $r_e$. With all else constant, this enhances $\bar{\tau}$ and increases $R_c$ and $\alpha_c$. Mechanisms governing the first two steps are both complex (e.g., Leaitch et al. 1992; Lohmann and Feichter 1997) and irrelevant to the point at hand, which can be made simply by reducing $r_e$. For illustrative purposes, $r_e$ was reduced in the following experiments by 2 $\mu$m to approximate the likely change over the most affected regions during the past $\sim 200$ yr (Boucher and Lohmann 1995). For simplicity, $r_e$ was assumed to be constant across GCM grid cells. In actuality, it is often correlated at smaller scales with cloud liquid water path $L$ (e.g., Szczechok et al. 2000, submitted to J. Atmos. Sci.) and may influence estimates of forcing (Feingold et al. 1997; Kogan et al. 1997). Moreover, assuming constant $r_e$ implies that distributions of $L$ also follow (1) with $\tau$ replaced by $\bar{\tau}$.

It is thought that following alterations to $r_e$ (first-order forcing), other cloud properties will change via a combination of immediate responses (second-order forcings) and feedbacks. Changes that are of particular interest here are those to $\bar{\tau}$ and $\nu$. Again, for the present arguments, the exact means by which these changes might arise are of secondary importance. What is important, however, is whether small changes to these quantities lead to significant differences in forcing estimates for homogeneous and inhomogeneous clouds.

The initial condition of the experiments reported here was

$$ \alpha_a(L_a| r_e) = \alpha_a(L_e| r_e), $$

where $L_a$ and $L_e$ are preindustrial mean liquid water paths required to equalize TOA albedos. The rationale behind (3) is that $L$ is tuned in GCMs in order that their TOA albedos and cloud fractions agree well with corresponding estimates from satellite data. In general, $L_a \sim L_e$ so if the GWTSAs were to replace a regular two-stream model in a GCM, more cloud water would be required to recover TOA albedo. Harshvardhan and Randall (1985) and Yu et al. (1997) estimated that on a global scale, $L_a \approx 0.3 L_e$ while Cahalan et al. (1994) estimated $L_a \approx 0.7 L_e$ for California marine stratuscumulus [which is probably too great a reduction at GCM time steps (Pincus et al. 1999)]. Since indirect forcing due to industrial aerosol occurs most over mid-latitude continents during early summer (Kiehl et al. 2000), where and when stratus and cumulus clouds prevail (Warren et al. 1986), the present study uses, for the most part, a representative value of $\nu = 1$ (see Barker et al. 1996; Oreopoulos and Davies 1998), which is similar to Cahalan et al.’s reduction [see appendix B of Barker (1996)].

Once $L_a$ and $L_e$ have been solved for in (3), $r_e$, $\bar{\tau}$, and $\nu$ are changed by $\Delta r_e$, $\Delta \bar{\tau}$, and $\Delta \nu$ thus yielding new (i.e., present day) values of TOA albedo. The ratio of forcings due to these changes

$$ R = \frac{\Delta \alpha_a}{\Delta \alpha_c} $$

$$ = \frac{\alpha_a(L_a + \Delta L, r_e + \Delta r_e) - \alpha_a(L_a, r_e)}{\alpha_c(L_a + \Delta \bar{\tau}, r_e + \Delta r_e) - \alpha_c(L_a, r_e)} $$

can be interpreted as the fractional albedo change error due to the homogeneous cloud assumption. Changes to $r_e$ only are discussed in the next section. This is followed by changes to both $r_e$ and $\bar{\tau}$, and finally by additional changes to $\nu$. 


4. Results

a. Changes to \( r_e \)

Figure 1 shows two sets of \( \alpha_\infty \) and \( \alpha_{\infty,1} \) as functions of \( \overline{L} \): one for \( r_e \) changing from 7 to 5 \( \mu \)m (continental) and another for \( r_e \) changing from 12 to 10 \( \mu \)m (oceanic) (Han et al. 1994). The cloud was situated near 1 km in the midlatitude summer atmosphere and the surface was Lambertian with constant albedo of 0.1. To estimate indirect aerosol forcings, begin by selecting a TOA albedo and find the corresponding values of \( \overline{L} \). Then, reduce \( r_e \) by 2 \( \mu \)m (arrows denoted by “1” and “2”) or reduce \( r_e \) and increase \( \overline{L} \) (arrows denoted by “3” and “4”) and find the albedo enhancements. Differences between final albedo values for the homogeneous and inhomogeneous models are interpreted as the error for indirect aerosol forcing made by the homogeneous model.

b. Changes to \( \overline{L} \)

As a direct consequence of reducing \( r_e \), it is suspected that precipitation may be suppressed thereby extending cloud life and increasing \( \overline{L} \) (Fouquart and Isaka 1992). Using a GCM, Lohmann and Feichter (1997) estimated that since the industrial revolution, \( \overline{L} \) might have increased anywhere from 2% to 10%. Thus, Fig. 1 shows albedo changes for the examples just discussed but in addition, \( \overline{L} \) was increased, nominally, by 8 g m\(^{-2}\) for both radiation models. As expected, albedo enhancements can be much larger than those for changes to \( r_e \) only (the more so the smaller \( \overline{L} \); see Table 1) with \( \Delta \alpha_{\infty} \) now exceeding \( \Delta \alpha_{\infty,1} \) by 30%–60% for the three examples. Again, this is primarily because \( \overline{L} \) and

\[
\left( \frac{\partial \alpha_{\infty}}{\partial \overline{L}} \right)_{r_e} \gg \left( \frac{\partial \alpha_{\infty,1}}{\partial \overline{L}} \right)_{r_e},
\]

as inhomogeneous clouds approach their asymptotic albedo slower than do homogeneous clouds (e.g., Coley and Jonas 1997). For the GWTSA, increases to \( \overline{L} \) were assumed to result from all \( L \) increasing by a constant fraction. This leaves \( \nu \) unchanged. Had preference been given to increase large \( \overline{L} \) (i.e., precipitating areas), the variance of \( \overline{L} \) would have increased thus reducing \( \nu \), suppressing \( \Delta \alpha_{\infty} \), and enhancing \( R \). The impact on \( R \) due to cloud structural alterations are discussed in more detail in the next section.

To give a broader impression of the potential impact of increasing \( \overline{L} \) brought on by indirect aerosol forcing, \( R \) were computed [using \( R_{\infty} \) and \( R_{\infty,1} \) in (4) rather than \( \alpha_{\infty} \) and \( \alpha_{\infty,1} \) at 0.55 \( \mu \)m above a black surface. This is a good proxy for the broadband as most of the changes in TOA albedo due to changes in \( r_e \) come from the visible where midlatitude, snow-free surface albedos are low (e.g., Tucker 1979). Following the same procedure...
as in (3) and (4), Fig. 2 shows \( \mathcal{R} \) as a function of initial values of \( r_e \) and \( \overline{L} \) when \( \Delta r_e = -2 \mu m, \nu = 1, \) and \( \Delta \overline{L} = 0 \) and 0.06 \( \overline{L}^* \) (i.e., a 6\% increase in \( \overline{L} \); cf. Lohmann and Feichter 1997). Values of \( \mathcal{R} \) derived from the curves shown in Fig. 1 and listed in Table 1 agree well with corresponding points on Fig. 2 (values in Table 1 are slightly smaller due to surface albedo of 0.1). Figure 2a shows that when \( \Delta \overline{L} = 0, \mathcal{R} \) is typically between 1.2 and 1.3. However, when liquid water paths are increased by 6\% (Fig. 2b), \( \mathcal{R} \) increases by roughly 0.15 implying \( \sim 40\% \) overestimation of cloud albedo increase by the homogeneous model.

Figure 3 shows \( \mathcal{R} \) as a function of \( \overline{L}^* \) and percentage increase in \( \overline{L}^* \) for \( \nu = 1, \) initial \( r_e = 9 \mu m, \) and \( \Delta r_e = -2 \mu m. \) At the low end of Lohmann and Feichter’s estimate (2\% increase in \( \overline{L}^* \)), \( \mathcal{R} \) is fairly constant near 1.3 regardless of \( \overline{L}^* \). Near their upper estimate, however, \( \mathcal{R} \) depends much on \( \overline{L}^* \) ranging from 1.25 for thin clouds to 1.5 for thick clouds. This suggests that the larger \( \overline{L}^* \), the more crucial it is to obtain good estimates of the increase in \( \overline{L}^* \) over the industrial era.

c. Changes to \( r_e, \overline{L}, \) and \( \nu \)

In addition to feedbacks involving the more familiar cloud properties, their seems little reason not to expect a feedback involving horizontal variability of unresolved cloud. Thus, how important might a conventional feedback term like

\[
\frac{\partial \alpha_c}{\partial \nu} \frac{d\nu}{dT_s} \quad (5)
\]

be, where \( T_s \) is surface temperature? Within the present study, this is the most convenient way to characterize cloud variability feedback; by no means is it the only way. While the first term in (5) is positive and of known magnitude, both magnitude and sign of the second term are uncertain; especially when this is meant to represent \( \nu \) for the distribution of \( \tau \) integrated through the entire atmospheric column (folding in the effects of overlapping cloud whose variability might vary with height).

There are two ways to realize cloud feedback involving \( \nu. \) First are direct changes to cloud structure as a result of more CCN (a kind of second-order forcing).\(^1\)

Second are changes to frequencies of occurrence of clouds with different typical values of \( \nu. \) This includes not only changing frequencies of occurrence but frequencies of multiple-layer cloud systems. For a short discussion on this issue, and feedback analysis in general, see appendix A.

Figure 4 shows \( \mathcal{R} \) as a function of \( \overline{L}^* \) and \( \Delta \nu \) when \( \nu = 1 \) and 2, \( r_e = 9 \mu m, \Delta r_e = -2 \mu m, \) and \( \Delta \overline{L} = 0. \) Though this range of \( \Delta \nu \) may seem small, it is not known whether it is small to the climate system. For \( \nu = 1 \) (Fig. 4a), \( \mathcal{R} \) increases significantly as \( \Delta \nu \) decreases; exceeding 1.5 by \( \Delta \nu = -0.1 \) for most \( \overline{L}^* \). This is because \( \Delta \alpha_c \) is suppressed on account of \( \partial \alpha_c / \partial \nu \Delta \nu < 0 \) and so errors due to the homogeneous assumption increase. It is possible that this may be what has happened over the industrial era given that the frequency of occurrence of convective clouds, with relatively small \( \nu \) (Barker et al. 1996), is roughly proportional to \( T_s \) (cf. Warren et al. 1986). This would make \( d\nu/dT_s < 0 \) implying a positive feedback involving cloud variability. On the contrary, in the upper-right corner of Fig. 4a, \( \mathcal{R} \) < 1 indicating that inhomogeneous clouds have become sufficiently less variable (\( \Delta \nu > 0 \)) upon reducing \( r_e \) that \( \Delta \alpha_c \) actually exceeds \( \Delta \alpha_c. \)

In contrast, Fig. 4b shows that when \( \nu = 2, \mathcal{R} \) is much less sensitive to \( \Delta \nu \) than for the case of \( \nu = 1. \) This is because both

\[
\left| \frac{\partial \alpha_c}{\partial \nu} \right|_{\nu = 1} \gg \left| \frac{\partial \alpha_c}{\partial \nu} \right|_{\nu = 2},
\]

and clouds with \( \nu = 2 \) are substantially less variable than those with \( \nu = 1 \) and so errors due to the homogeneous assumption are smaller.

Figure 5 shows \( \mathcal{R} \) as a function of initial \( \nu \) and \( \overline{L}^* \) for \( r_e \) going from 9 to 7 \( \mu m \) with \( \Delta \overline{L} = 0.06 \overline{L}^* \) for both models, and \( \Delta \nu = \pm 0.05. \) It is clear from these plots that if \( \nu \) is much larger than 2, there is little error in albedo change (<20\%) due to the homogeneous mod-

\(^1\) By relaxing the assumption of constant \( r_e \) across grid cells but assuming constant droplet number density \( N \) and \( \tau \sim \overline{L}^{0.7}N^{0.7} \) (Feingold et al. 1997), increasing \( N \) would slightly reduce \( \nu. \) This formulation complicates the use of the GWTS model and its assumptions are only marginally more valid than constant \( r_e. \)
Fig. 2. Fractional differences \( \Delta \) between albedo changes for homogeneous (\( \nu = \infty \)) and inhomogeneous (\( \nu = 1 \)) clouds as functions of initial droplet effective radii \( r_e \) and when \( r_e \) are decreased by 2 \( \mu m \). (a) Cloud liquid water paths and variability are held fixed (\( \Delta \overline{L} = \Delta \nu = 0 \)). (b) Cloud variability is fixed but liquid water paths increase by 6%. Here \( \Delta > 1 \) signify overestimation of forcing by the homogeneous model.

Even by \( \nu = 3 \), \( \Delta \) is almost insensitive to \( \Delta \nu \) implying minimal cloud variability feedback. However, when \( \nu = 1 \), substantial errors due to the homogeneous assumption are incurred regardless of the sign of \( \Delta \nu \).

All in all, assuming that the most likely ranges of initial \( \overline{L}_e \) and \( r_e \) are approximately within (50, 100) \( g \ m^{-2} \) and (8, 12) \( \mu m \), and \( \Delta r_e \) across the last 200 yr is roughly \( -2 \mu m \), results presented here suggest that homogeneous solar radiative transfer models are prone to overestimate albedo changes due to indirect aerosol forcing by anywhere from 10% to 50%.

5. Summary and conclusions

This study employed the gamma-weighted two-stream approximation to demonstrate that the most comprehensive estimates of both indirect forcing and ensuing cloud–shortwave feedbacks may be overestimated significantly due to the assumption made within climate models that clouds are homogeneous for horizontal scales less than several hundred kilometers. Using reasonable estimates for the variance of \( L \) and assuming that \( r_e \) has decreased on average by \( \sim 2 \mu m \), albedo overestimation by regular two-stream models may range anywhere from 15% to 30% (for smaller changes to \( r_e \), relative errors increase but absolute errors decrease). When \( \overline{L} \) is increased following reduction of \( r_e \), these overestimations increase significantly; comfortably reaching 40%–50% depending on the fractional increase in \( \overline{L} \). Going a step further, when cloud variability is allowed to change (via feedbacks), by seemingly small amounts, these overestimations can be diminished (fortuitously) if clouds become less variable, or enhanced (up to 100%) if they become more variable. Though the latter scenario may seem more logical, what actually happens is unknown at present. Moreover, there are many aspects of this problem that were beyond the scope of this simple study such as: shielding of low cloud by...
Fig. 4. Fractional differences $R$ between albedo changes for homogeneous ($\nu = \infty$) and inhomogeneous clouds as functions of $L_\nu$ and change to the variability parameter of inhomogeneous cloud $D_n$ for droplet effective radii changing from 9 to 7 $\mu$m. Liquid water paths are fixed at initial values. Initial values of $\nu$ are 1 and 2 in (a) and (b), respectively.

Fig. 5. Fractional differences $R$ between albedo changes for homogeneous ($\nu = \infty$) and inhomogeneous clouds as functions of $L_\nu$ and initial $\nu$ when droplet effective radii decrease from 9 to 7 $\mu$m and mean liquid water paths increase by 6%. Changes to $\nu$ are $+0.05$ and $-0.05$ in (a) and (b), respectively.

high cloud, changing cloud overlap patterns, possible nonlinear sun angle effects, variable surface albedos, and complex changes to conditional probability distributions involving cloud structure (i.e., $\nu$) and other cloud variables.

To conclude, results presented here concur with, and extend, the warning issued by other studies (e.g., Stephens 1988; Rossow 1989; Davis et al. 1990; Barker et al. 1999): radiative transfer for the earth–atmosphere system is dictated by the nature of the medium, not by assumptions that make for tractable models. Thus, when mean radiative fluxes are to be computed for GCM-size domains, cloud morphology (beyond fractional amount) must be accounted for.

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APPENDIX

Linear Feedback Analysis Using Conditional Probability Distribution Functions

Global climate models are often analyzed in terms of linear feedback theory. As such, the change to Earth's surface air temperature $\Delta T_s$ due to a radiative perturbation at the tropopause $\Delta F$ is expressed as

$$
\Delta T_s = \frac{-\Delta F}{S_0 \frac{\partial \alpha_p}{\partial T_s} + \frac{\partial I}{\partial T_s} + \sum_k \left( \frac{S_0 \frac{\partial \alpha_p}{\partial x_k} + \frac{\partial I}{\partial x_k}}{4} \right) dx_k},
$$

(A1)

where $S_0$ is the solar constant, $\alpha_p$ is planetary albedo, $I$ is net longwave flux at the tropopause, and $x_k$ are climate variables. Using solar radiation and cloud optical depth $\tau$ as an example, feedback terms are normally expressed as

$$
f_{\tau,\nu} = \int \cdots \int \left[ \sum_{k=1}^n p_k(x_k | x_1, \ldots, x_{k-1}, x_{k+1}, \ldots, x_n) \left( \frac{\partial \alpha_p}{\partial x_k} \frac{dx_k}{\partial T_s} \right) \right] dx_1 \cdots dx_n.
$$

(A4)

As an example, if cloud feedback involved only $\nu$ and $\tau$ [see Eq. (1)], and $q(\nu | \tau)$ and $p(\tau | \nu)$ are conditional distributions, the cloud feedback parameter is

$$
f_{\nu,\tau} = \int_0^\infty \int_0^\infty p(\tau | \nu) \left( \frac{\partial \alpha_p}{\partial \tau} \right) d\tau d\nu + q(\nu | \tau) \left( \frac{\partial \alpha_p}{\partial \nu} \right) d\nu.
$$

(A5)

This acknowledges that there are distribution functions that describe the instantaneous distribution of clouds in model grid cells. Note also that the radiative transfer model is no longer the regular two-stream as used in current climate models, but rather the GWTSA (to facilitate use of current climate models, but rather the GWTSA (to facilitate use of current climate models).

Though expressions (A3)–(A5) are much more detailed than (A2), they still do not recognize the possibility that the distribution of a climate variable may not change its form but may shift diurnally or even seasonally. This points to another subtlety: conditional distributions associated with all solar-related feedbacks should be filtered by the diurnal solar cycle as changes to nocturnal cloud properties are irrelevant to cloud–albedo feedbacks.

**REFERENCES**


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