Indirect Forcing by Anthropogenic Aerosols: A Method for Testing a Parameterization

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Introduction

Atmospheric aerosols affect the radiation balance through direct and indirect effects. The direct effect refers to the scattering and absorption of radiation by the aerosols themselves. The indirect effect refers to changes in cloud optical properties by aerosols that act as cloud condensation nuclei (CCN). An increase in aerosols may result in an increase in cloud drop number concentrations which, in the absence of absorbing aerosols, leads to higher cloud reflectivity.

We have developed a parameterization for the indirect effect that is based on a mechanistic description of droplet formation and the chemical processes controlling the formation of sulfate. It has been evaluated by comparing the predicted droplet size from our coupled climate-aerosol model with those measured by satellite. However, a much more thorough test of the parameterization is needed. We propose to evaluate the parameterization using data available through the ARM program.

In order to provide a global understanding of the effects of aerosols on clouds, one must first understand the global concentrations of the different aerosol components or types. Our past work has been aimed at developing an understanding of global and regional aerosol abundances, and developing a parameterization of cloud response to aerosol abundance (Chuang and Penner 1995) to evaluate the importance of aerosol/cloud interactions to climate forcing. In order to understand whether the aerosol particles act as CCN, one needs to know the composition of hygroscopic material in the aerosol (e.g., sulfate, nitrates, ammonium) (Pruppacher and Klett 1978). This understanding requires a quantitative understanding, on a global basis, of the aerosol sources, transformation and removal processes.

Recently, Boucher and Rodhe (1994), Jones et al. (1994), Boucher and Lohmann (1995), and Jones and Slingo (1996) have each developed parameterizations relating cloud drop concentration to sulfate mass or aerosol number concentration and used them to develop estimates of the indirect forcing by anthropogenic sulfate aerosols. These parameterizations made use of measured relationships in continental and maritime clouds. However, these relationships are inherently noisy, yielding more than a factor of 2 variation in cloud drop number concentration for a given aerosol number (or for a given sulfate mass) concentration. They do not make use of information from the climate model regarding local updraft velocities, and they have had to make certain simplifying assumptions.

In contrast to previous studies, our parameterization of the effects of aerosols on cloud droplet distributions uses a more mechanistic approach. The characteristics of the cloud drop size distribution near cloud base are initially determined by the size distribution and chemical characteristics of the aerosol particles that serve as CCN and by the local updraft velocity (Lee et al. 1980; Chuang et al. 1992). Once drop concentrations at cloud base are established, measurements have shown that these remain near constant with altitude throughout the main part of the clouds, at least in the case of stratiform and stratocumulus clouds (Nicholls 1984; Bower et al. 1994; Mitchell, DRI, private communication). Thus, ideally, it should be possible to use these fundamental properties of aerosol size, chemical composition, and updraft at cloud base to predict the effects of anthropogenic aerosols on drop number concentrations in stratiform clouds in general circulation models. Chuang and Penner (1995) presented the basic parameterization, while Chuang et al. (1997) used it to explore the indirect forcing by anthropogenic sulfate aerosols. Penner et al. (1996) have also used the parameterization to explore indirect forcing by carbonaceous aerosols. Our task now is to test the parameterization.

The aerosol/cloud parameterization has the following form:

$$N_{cloud} = \frac{N_a v}{cN_a + v} \quad (1)$$

where N_{cloud} is the droplet number concentration, N_a is the aerosol number, v is the updraft velocity, and c is a correction factor, derived using a detailed microphysical model which accounts for the aerosol size distribution and composition (Ghan et al. 1993).

Given this droplet number concentration, cloud optical depth may be estimated from (Twomey 1977):

$$\tau = \pi r_e^2 Q_{ext} N_{cloud} h$$

where r_e is the droplet effective radius, Q_{ext} is the extinction efficiency (effectively $Q_{ext} = 2$ for visible radiation) and h is the depth of the cloud. This expression can also be used to express τ in terms of the aerosol number concentration using:

$$w_L = \frac{4}{3}\pi r_e^3 N_{cloud} \rho_w$$

where w_L is the liquid water content of the cloud and ρ_w is the density of water. Thus, we have the equivalent expression for cloud optical depth:

$$\tau = h\left[\frac{9\pi w_L^2 N_{cloud}}{2\rho_w^2}\right]^{\frac{1}{3}}. \quad (2)$$

The general circulation model uses these calculated optical depths with a δ-Eddington method to predict reflected radiation from the top of the cloud. The albedo of the cloud may be calculated from:

$$\alpha_c = \frac{\beta(\mu_0)\,\tau / \mu_0}{1 + \beta(\mu_0)\,\tau / \mu_0}$$

where μ_0 is the cosine of the solar zenith angle and β_0 is the fraction of sunlight incident to the cloud (at the angle corresponding to μ_0) that is scattered in the upward direction for single-particle scattering. Cloud albedo may be used to predict the reflected solar radiation.

To test the parameterization, we are beginning to examine data from ARM. Above-cloud reflected radiation can be sorted as a function of aerosol number concentration along lines of constant $p = h[9\pi w_L^2/2\rho_w^2]^{1/3} = \tau_- / N_{cloud}^{1/3}$ (see Equation 2). Figure 1 shows the reflected radiation for different values of ρ showing the variation with updraft velocity.

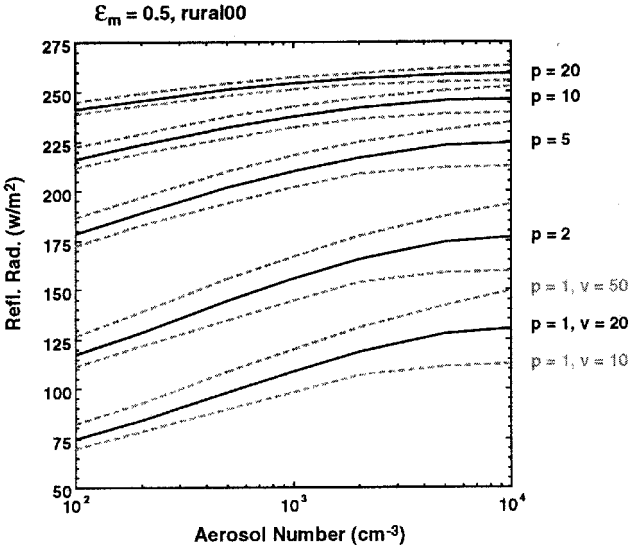

Figure 1. Predicted reflected radiation from cloud top as a function of aerosol number concentration and the parameter $\rho = \tau / N_{cloud}^{1/3}$ (from Equation 2). The variation for each value of ρ for vertical velocities ranging from 10 to 50 cm s^{-1} is also shown.

This velocity variation derives from its presence in Equation 1 above. We note from this figure, that the predicted reflected radiation shows a distinct response to the aerosol number concentration. Furthermore, although we are unlikely to have estimates available for the parameter v as a routine, the predicted cloud response is not very sensitive to v, at least over the normal range of observed vertical velocities in stratiform clouds. The unique character of this response can be used together with data at the ARM site to test the cloud droplet/albedo response to aerosol changes predicted by the parameterization.

We have also examined the response of the reflected radiation to variations in aerosol size distribution and composition. Again, for a reasonable range of variation, the reflected radiation is not very sensitive. Thus, we might hope to examine the response of reflected radiation using data from ARM.

To use the ARM data, we plan to use readily available data for cloud top and cloud base (inferred from satellite, ceilometer and micropulse lidar instruments) and cloud liquid water

content (inferred from the satellite- and radar-measured liquid water path and soundings when available) to sort ARM data by the parameter $p = \tau\_/ N_{cloud}^{1/3}$ (from Equation 2). By plotting the dependence of measured reflected solar radiation against measured aerosol concentration we will determine whether the measured relationship corresponds to that given in Figure 1.

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