Impact of Volcanic Aerosol on Mean Fluxes of Solar Radiation in Broken Clouds

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Introduction

One of the powerful means of detecting the volcanic effect on the earth's radiation budget and, hence, on the earth's climate is numerical modeling. However, the atmospheric processes, during and after the eruption, are too complex and frequently prohibit an adequate description of the non-unique effect of volcanic activity on the earth's energy balance and its individual components. This explains why different model results, obtained by many authors, not only disagree among themselves, but also with the data of field measurements (Kondratiev 1992). To improve the quality of radiation calculations, it is necessary to update the existing models and, in particular, the radiation codes upon which they are basically built.

Present estimates of the variability of radiative properties of the cloudy atmosphere, caused by the presence of volcanic aerosol, are based on the plane-parallel horizontally homogeneous cloud model (Pollack et al. 1976; Harshvardhan 1979; Cess et al. 1981; Pollack and Ackerman 1983; Russell et al. 1993). However, as is well known, the real cloud fields are stochastic in nature because of both fluctuating cloud optical parameters and geometrical parameters. So how important is it to take the cloud stochasticity into account when radiation calculations are made in the atmosphere containing the volcanic aerosol?

As one of the aspects of the problem, here we investigate how strongly the variations of the optical depth and single scattering albedo of volcanic aerosol influence the mean fluxes of solar radiation in the system "broken clouds - volcanic aerosol - underlying surface."

Model and Method

The model "broken clouds – volcanic aerosol – underlying surface" is defined in the height interval $0 \le z \le H^t_{atm}$, H^t_{atm} = 50 km, as a set of K atmospheric layers. A unit solar flux is incident on the atmospheric top boundary $z = H^t_{atm}$ in direction $\vec{\omega}_\oplus = (\xi_\oplus, \varphi_\oplus)$, where $\xi_\oplus$ and $\varphi_\oplus$ = 0 are zenith and azimuthal solar angles (Figure 1).

Each ith aerosol layer is assumed horizontally homogeneous, and it is characterized by the extinction coefficient $\sigma^a_{\lambda,i}$, single-scattering albedo $\omega^a_{\lambda,i}$, and scattering phase function $g^a_{\lambda,i}(\vec{\omega}, \vec{\omega}')$, i=1,2…K, λ is a wavelength.

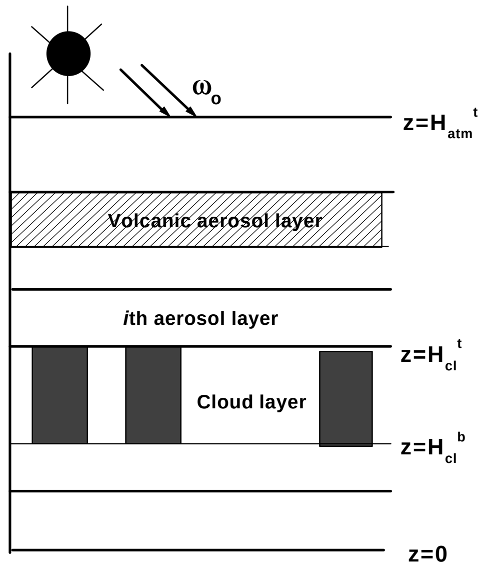

Figure 1. Model of the cloudy-aerosol atmosphere.

Clouds occupy a separate layer $H_{cl}^{b} \leq z \leq H_{cl}^{t}$, where H_{cl}^{b} and H_{cl}^{t} are cloud base and top heights, respectively. The cloud optical model is specified in terms of the random scalar fields of the cloud extinction coefficient $\sigma_\lambda \kappa(\vec{r})$, single scattering albedo $\omega_\lambda \kappa(\vec{r})$, and scattering phase function $g_\lambda(\vec{\omega}, \vec{\omega}')\kappa(\vec{r})$; the mathematical model of $\kappa(\vec{r})$ field is constructed based on the Poisson point fluxes on the straight lines (Titov 1990).

The underlying surface reflects the incident radiation according to Lambert law and has the albedo A_s.

The mean fluxes of solar radiation are calculated using the method of closed equations, developed for statistically homogeneous cloud fields by Titov (1990).

Calculation Results

Calculations are made for optical characteristics of the cloudy-aerosol atmosphere, corresponding to a wavelength λ = 0.55 μm (in what follows, the index "λ" is omitted for simplicity).

The extinction coefficient and single-scattering albedo are chosen in accordance with LOWTRAN-5 (Air Force Geophysics Laboratory 1980) specification. Because we consider only variations of mean radiative fluxes caused by variability of optical depth and single-scattering albedo of volcanic aerosol, within the aerosol atmosphere the scattering phase function is simply specified such as to correspond to the Haze *H* model within the lower stratosphere and to the Haze *L* model elsewhere (Deirmendjian 1969).

The scattering phase function of water clouds (layer 1 km to 2 km) is calculated from Mie theory (Deirmendjian 1969) for the C_1 cloud. It is assumed that within the cloudy layer the absorption is absent and the single-scattering albedo $\omega = 1$. The aspect ratio $\gamma = H \mid D$ (where H is the thickness, and D the characteristic horizontal size of cloud elements) varies in the range $0 \le \gamma \le 2$.

Surface albedo was varied in the range $0.0 \le A_s \le 0.8$.

Below we analyze the mean albedo, calculated at the level of the top of the atmosphere (TOA) $z = H^t_{atm}$ (A), and mean diffuse transmittance at the level z = 0 (Q_s) at $A_s = 0.2$. For convenience, the term "mean" will not be used in the following.

Occurrence of volcanic aerosol in the atmosphere modifies boundary conditions for the cloudy layer. Specifically, unscattered flux $S\left(H^t_{cl}, \xi_\oplus, \varphi_\oplus\right)$ and diffuse flux $Q_s\left(H^t_{cl}\right) = \int_0^{2\pi} \int_{-1}^{0} I^{\downarrow}\left(H^t_{cl}, \xi, \varphi\right) \cos\xi \, d\xi \, d\varphi$ of solar radiation are now incident on the cloud top boundary $z = H^t_{cl}$. The greater $S\left(H^t_{cl}, \xi_\oplus, \varphi_\oplus\right)$, the closer the mechanism of formation of cloud radiative properties to that in the case with a collimated source of radiation as described extensively by Zuev and Titov (1995). Conversely, the larger the optical thickness of volcanic aerosol and the solar zenith angle, the less the fraction of unscattered radiation $S\left(H^t_{cl}, \xi_\oplus, \varphi_\oplus\right)$, and the more complex processes are involved in the formation of radiative properties.

In this work, we intend to answer two questions discussed below.

1. How strongly do albedo and diffuse transmittance of the system "broken clouds - volcanic aerosol - underlying surface" change in response to variations of τ_{VA} and ϖ_{VA} in cumulus and equivalent stratus clouds? (Cumulus and equivalent stratus clouds only differ in the aspect ratio γ, being $\gamma << 1$ for stratus and $0.5 \le \gamma \le 2$ for cumulus.)

Changes in A and Q_s, due to increase of stratospheric aerosol optical depth from $\tau_{VA} = 0.005$ to $\tau_{VA} = 0.3$, will be characterized by the quantity

$$\Delta_{VA} R = \frac{\left(R\left(\tau_{VA} = 0.3\right) - R\left(\tau_{VA} = 0.005\right)\right)}{R\left(\tau_{VA} = 0.005\right)}, R = A, Q_s .$$

We now let $\omega_{VA} = 1$. As calculations show, significant (>5%) relative increases of albedo and transmittance, $\Delta_{VA}A$ and $\Delta_{VA}Q_s$, occur for small and intermediate cloud fractions $N \le 0.7$ at particular solar zenith angles. For $\xi_\oplus \le 30°$, $\Delta_{VA}A$ is small (5% to 8%) for both stratus and cumulus clouds (Figure 2a). Influence of cloud type on $\Delta_{VA}A$ increases with growing $\xi_\oplus$; for instance, at $\xi_\oplus = 75°$ and N = 0.1, the relative increase of stratus albedo $\Delta_{VA}A_{St}$ exceeds that of cumulus albedo $\Delta_{VA}A_{Cu}$ by approximately a factor of 2 (Figure 2c).

In stratus clouds, when τ_{VA} grows, the $\Delta_{VA}Q_{s,St}$ value may amount to tens of a percent. In cumulus, a significant (≈40%) gain $\Delta_{VA}Q_{s,Cu}$ occurs only for small cloud fractions (N = 0.1) and small solar zenith

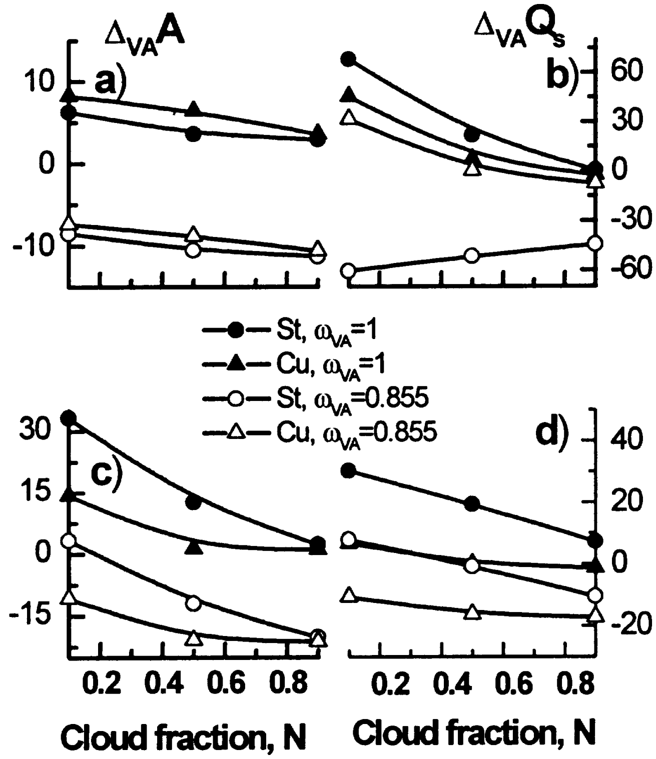

Figure 2. Relative changes in atmospheric albedo and transmittance caused by the presence of volcanic aerosol, calculated for different input model parameters: $\gamma_{Cu} = 2$; (a, b) $\tau = 10$, $\xi_\oplus = 3030°$; (c, d) $\tau = 40$, $\xi_\oplus = 75°$.

angles ($\xi_\oplus = 30°$) (Figure 2 b,d). This can be explained as follows. The bottom boundary of the layer containing volcanic aerosol represents the source of diffusely transmitted photons. In the presence of stratus clouds in the atmosphere, most of these photons reach the surface through gaps between the clouds, whereas in cumulus they are trapped by cloudy elements and suffer extra attenuation.

How will A and Q_s change when going from the Background Stratospheric Model to the Extreme Volcanic Model? (The latter model describes the atmospheric state several weeks after a large volcanic eruption, when volcanic aerosol not only scatters but also absorbs radiation: $\omega_{VA} < 1$). Let us compare A and Q_s, calculated for $\tau_{VA} = 0.005$, $\omega_{VA} = 1$ and $\tau_{VA} = 0.3$, $\omega_{VA} = 0.855$. When $\omega_{VA} < 1$, radiation exiting from the cloud top boundary $z = H_{cl}^{t}$ passes through the layer of absorbing and scattering volcanic aerosol. Thus, a reduction in albedo of the system "broken clouds - volcanic aerosol - underlying surface" is expected (Figure 2 a,c). Behavior of ΔQ_s with decreasing single-scattering albedo is more complex and depends on N, $\xi_\oplus$, and cloud type (Figure 2 b,d).

2. As is well known, mean fluxes in cumulus and equivalent stratus clouds, in the case when a collimated flux of solar radiation is incident on the cloud top, may differ by as much as tens of a percent (Zuev and Titov 1995). Will this difference change due to the presence of a layer of volcanic aerosol above the cloudy layer?

Figure 3 presents relative differences in albedo A and diffuse transmittance Q_s, calculated from the formula

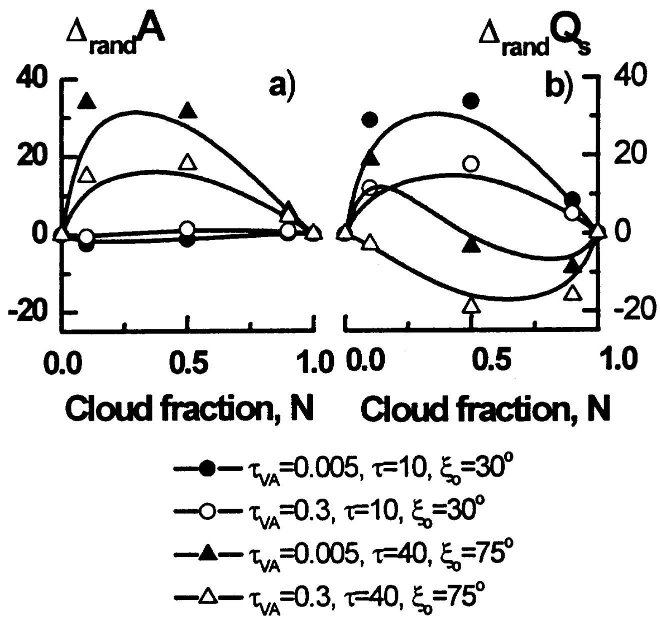

Figure 3. Influence of stochastic cloud geometry on albedo and transmittance of the atmosphere with and without volcanic aerosol present (γ_{Cu} = 2).

$$\Delta_{rand}R = \frac{R_{Cu} - R_{St}}{R_{St}}, R = A, Q_s$$

For small solar zenith angles $\xi_\oplus \leq 30°$, $\Delta_{rand}A$ does not exceed 2% while optical depth of the lower stratosphere varies in the range $0.005 \leq \tau_{VA} \leq 0.3$. When $\xi_\oplus$ is large, the influence of stochastic cloud geometry is approximately a factor of 2 more significant under background conditions $\tau_{VA} = 0.005$ than at $\tau_{VA} = 0.3$ (Figure 3a).

Zuev and Titov (1995) showed that, when $\xi_\oplus \leq 60°$, $Q_{s,Cu} \geq Q_{s,St}$ and, hence, $\Delta_{rand}Q_s \geq 0$. As τ_{VA} increases, Q_s grows for all cloud types (Figure 2), while the relative difference $\Delta_{rand}Q_s$ correspondingly decreases (Figure 3b). At still larger $\xi_\oplus$ ($\xi_\oplus$ = 75°), the fraction of scattered radiation in cumulus remains larger than in stratus:

$$A_{Cu} + Q_{s,Cu} \geq A_{St} + Q_{s,St} \tag{1}$$

When τ_{VA} has a background value and $N \leq 0.5$, i.e., when the fraction of unscattered radiation at the level z = 0 is quite large, inequality (1) is so strong that the inequality $Q_{s,Cu} \geq Q_{s,St}$ holds in addition to the inequality $A_{Cu} \geq A_{St}$. As the fraction of unscattered radiation decreases either due to N increase at $\tau_{VA} = 0.005$, or to τ_{VA} growth up to 0.3, inequality (1) becomes weaker. As a consequence, $\Delta_{rand}Q_s \leq 0$ when $\tau_{VA} = 0.005$ and $N \geq 0.5$, and when $\tau_{VA} = 0.3$ and $0 \leq N \leq 1$ (Figure 3b).

With decreasing single-scattering albedo of volcanic aerosol, values of $\Delta_{rand}A$ and $\Delta_{rand}Q_s$ change insignificantly: $\Delta_{rand}R(\omega_{VA} = 1) \approx \Delta_{rand}R(\omega_{VA} = 0.855)$, R = A, Q_s.

Conclusions

Using numerical simulation techniques, we studied how optical depth and single-scattering albedo of volcanic aerosol influence the albedo and transmittance of low-level water clouds. It is shown that

- at $\omega_{VA} = 1$, as τ_{VA} increases, A and Q_s both increase, by a comparable or larger amount in stratus than in cumulus. This means that the neglect of stochastic cloud geometry may lead to overestimation of the mean albedo and diffuse transmittance of the system "broken clouds - volcanic

- for $\omega_{VA} < 1$, A decreases, while the variability of Q_s depends on cloud type, $\xi_{\oplus}$, and N.

- albedo difference between cumulus and stratus clouds is decreased by the presence in the atmosphere of a volcanic aerosol layer.

The presented estimates are based on the assumption of horizontal homogeneity of volcanic aerosol, which in reality remains highly inhomogeneous for a long period after eruption. The horizontal inhomogeneity effects of volcanic aerosol will be the subject of our future study.

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