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PERTURBATION OF THE ZONAL RADIATION BALANCE BY A STRATOSPHERIC AEROSOL LAYER

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ABSTRACT

The effect of stratospheric aerosols on the earth's monthly zonal radiation balance is investigated using a model layer consisting of 75% H_2SO_4, which is the primary constituent of the background aerosol layer.

INTRODUCTION

The role of stratospheric aerosols in enhancing the reflection of solar radiation has been investigated recently by numerous authors (e.g., Pollack, et al. 1976). Model calculations have shown that the dominant climatic effect of these aerosols is enhanced solar reflection with increased infrared opacity playing a smaller role. However, Herman, et al., (1976) have also pointed out that the degree of albedo enhancement is a strong function of the albedo of the underlying surface and the zenith angle of the incident solar radiation.

In this work, the effect of the aerosol layer at solar wavelengths will first be presented in terms of the albedo sensitivity, which may be defined as $d\alpha/d\tau$, where α is the unperturbed system albedo and τ is the optical depth computed at some reference wavelength, here $0.55\mu m$. The zonal variation of albedo sensitivity by month, together with the seasonal variation of solar irradiation yields the solar perturbation. With the inclusion of the effect of the stratospheric aerosol layer on the infrared radiation leaving the top of the atmosphere, the net energy deficit on a zonal basis is obtained.

AEROSOL LAYER

There is a persistent tenuous layer of submicron size aerosols in the stratosphere that is composed primarily of an aqueous solution of sulfuric acid. The number concentration of this aerosol layer increases after major volcanic events, and as the residence time of these aerosols is of the order of a year or more, radiative perturbations may be expected. Initially this enhanced concentration is limited to the vicinity of the explosion but in a matter of months may spread globally.

One result of this severe enhancement is the additional reflection of solar energy back to space. A radiative model of the aerosol is needed to compute the reflectivity of the layer. The aerosol composition is assumed to be 75% H_2SO_4 and a modified gamma size distribution is adopted. Since the optical depth of the aerosol layer is small ($\sim$0.1) even after major volcanic events, the optically thin approximation is used to compute the reflectivity of the layer. Use of the Henyey-Greenstein phase function yields analytical expressions for the planar albedo, $a(\mu_o)$, and global albedo, $\bar{a}$, which are used here. The altered albedo of the earth-atmosphere system with the addition of a non-absorbing aerosol layer is

$$\alpha(\mu_0) = a(\mu_0) + \frac{\alpha'(\mu_0)\ [1 - a(\mu_0)]\ (1 - \bar{a})}{1 - \bar{a}\ \alpha'(\mu_0)} \tag{1}$$

where $\alpha'(\mu_0)$ is the unperturbed system albedo. A similar relationship exists for absorbing aerosols.

The albedo sensitivity averaged over the solar spectrum for the model aerosol layer is presented in Figure 1. Use has been made of the monthly zonal albedo given by Ellis and Vonder Haar (1976) to compute the sensitivity. The isopleths are in units of change in albedo in percent of incident solar energy for an optical depth perturbation of $\Delta\tau = 0.01$. In the optically thin limit, the response is linear and the effect of other layer thicknesses or non-uniform spread can be easily deduced. The dominant feature is seen to be the zenith angle dependence translated here in a monthly and latitudinal dependence. Over high latitudes the effect of surface reflectivity may also be noticed, such as the lower values of $\Delta\alpha/\Delta\tau$ in late winter and early spring when there is snow cover on the ground.

RADIATION BALANCE

The net zonal radiation balance is

$$D(\phi) = Q(\phi)\ [1 - \alpha(\phi)] - I(\phi) \tag{2}$$

where $Q(\phi)$ is the solar insolation at latitude ϕ, $\alpha(\phi)$ the albedo, $I(\phi)$ is the outgoing terrestrial infrared radiation. The presence of a stratospheric aerosol layer modifies α through increased reflectivity and I through an increased greenhouse effect acting in the opposite direction. The reduction in solar energy absorbed, $Q\Delta\alpha$, caused by a uniform stratospheric layer of $\tau_{vis} = 0.1$ is plotted in Figure 2 which follows from Figure 1 and the distribution of $Q(\phi)$ for each month.

The net loss in a column of the earth-atmosphere system, however, is $Q\Delta\alpha + \Delta I$ (note that ΔI is a gain, hence negative), which is the change in the radiation balance and it is necessary to compute the infrared greenhouse sensitivity of the aerosol layer. This is done for five model atmospheres using an atmospheric radiation model similar to the one by Harshvardhan and Cess (1978) in which an emissivity formulation has been used to calculate the infrared flux. When the infrared effect is combined with the solar perturbation, the net reduction in zonal radiation balance is obtained and is plotted in Figure 3.

The perturbation in the radiation balance can also be expressed as a change in the effective black-body radiative temperature of the atmosphere. This change is given by

$$\Delta T = -\frac{Q\Delta\alpha + \Delta I}{4\sigma T^3} \tag{3}$$

where σ is the Stefan-Boltzmann constant, T the effective black-body radiative temperature of the atmospheric column and $(Q\Delta\alpha + \Delta I)$ the decrease in the radiation balance. Figure 4 is a plot of ΔT in °K for $\tau_{vis} = 0.1$. It can be seen that the major radiative effect of the stratospheric aerosol layer occurs in the spring and fall when there is a pronounced change in the equator to pole radiative energy gradient. As the results pertain to the entire atmospheric column, it is not possible to estimate the change in the diabatic heating at various levels of the atmosphere or estimate surface temperature changes. However, it appears from this analysis that a uniform layer of stratospheric aerosols would have only a small effect on the long term radiative regime equatorwards of 50° - 60°, if the change in turbidity corresponds to that caused by the Agung eruption.

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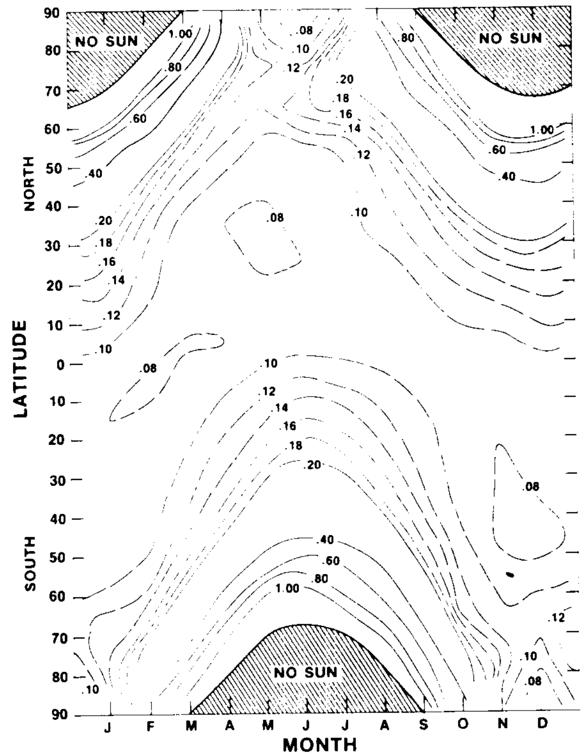

Fig. 1-Monthly zonal albedo sensitivity with solar averaged reflectivity.

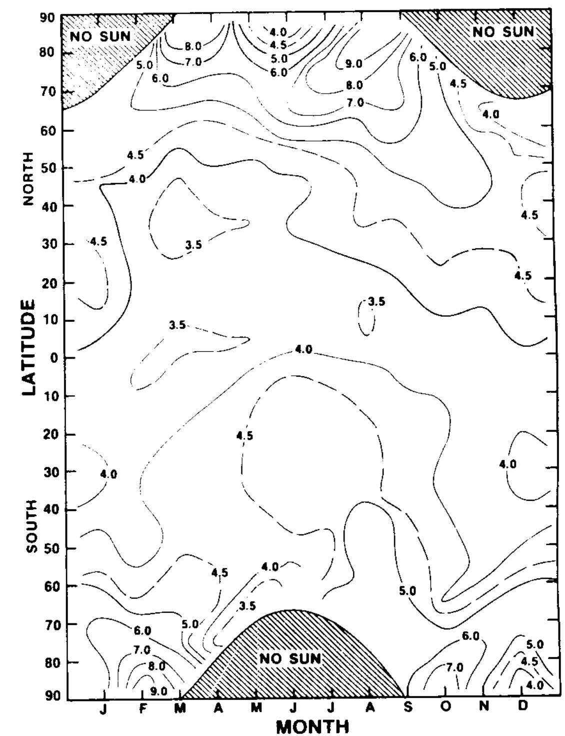

Fig. 2-Reduction in mean monthly solar energy absorbed in W/m^2 with addition of aerosol layer of $\tau_{vis} = 0.1$.

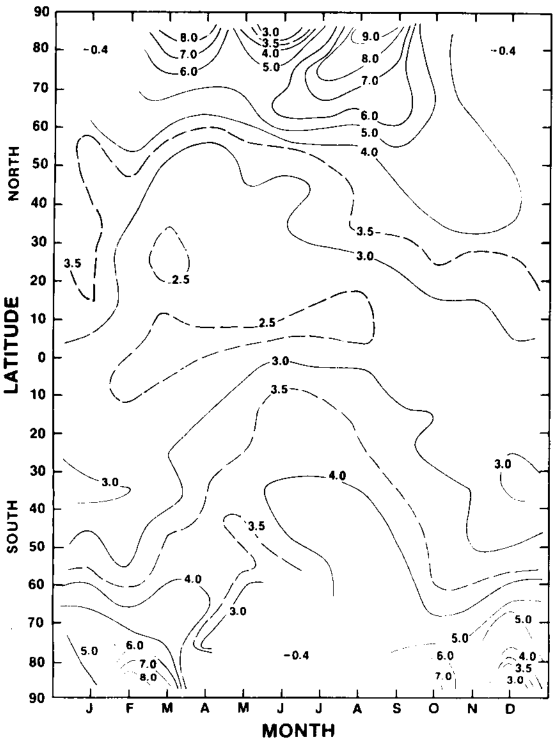

Fig. 3–Reduction in the zonal radiation balance in W/m^2 with addition of aerosol layer of $\tau_{vis} = 0.1$.

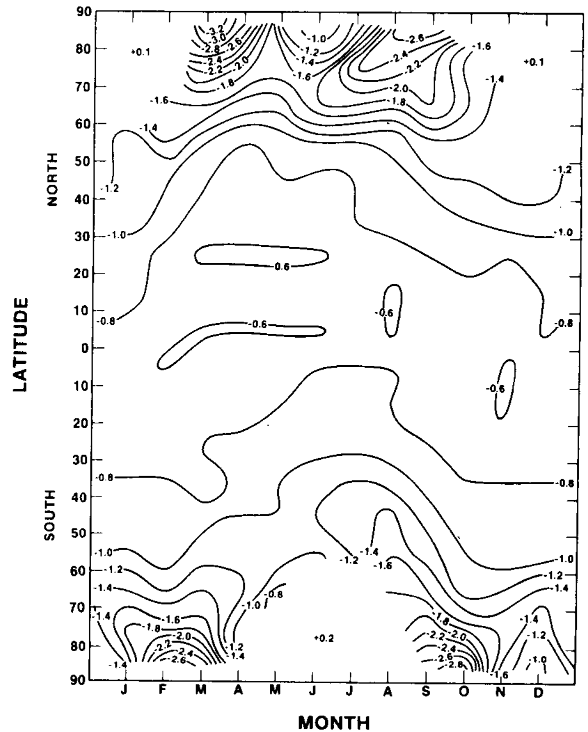

Fig. 4–Change in the effective black-body radiative temperature of the atmosphere in °K with addition of aerosol layer of $\tau_{vis} = 0.1$.